2000 Years of Drought Variability in the Central United States

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ABSTRACT

Droughts are one of the most devastating natural hazards faced by the United States today. Severe droughts of the twentieth century have had large impacts on economies, society, and the environment, especially in the Great Plains. However, the instrumental record of the last 100 years contains only a limited subset of drought realizations. One must turn to the paleoclimatic record to examine the full range of past drought variability, including the range of magnitude and duration, and thus gain the improved understanding needed for society to anticipate and plan for droughts of the future. Historical documents, tree rings, archaeological remains, lake sediment, and geomorphic data make it clear that the droughts of the twentieth century, including those of the 1930s and 1950s, were eclipsed several times by droughts earlier in the last 2000 years, and as recently as the late sixteenth century. In general, some droughts prior to 1600 appear to be characterized by longer duration (i.e., multidecadal) and greater spatial extent than those of the twentieth century. The authors' assessment of the full range of past natural drought variability, deduced from a comprehensive review of the paleoclimatic literature, suggests that droughts more severe than those of the 1930s and 1950s are likely to occur in the future, a likelihood that might be exacerbated by greenhouse warming in the next century. Persistence conditions that lead to decadal-scale drought may be related to low-frequency variations, or base-state shifts, in both the Pacific and Atlantic Oceans, although more research is needed to understand the mechanisms of severe drought.

1. Introduction

Drought is one of the most damaging climate-related hazards to impact societies. Although drought is a naturally occurring phenomenon throughout most parts of the world, its effects have tremendous consequences for the physical, economic, social, and political elements of our environment. Droughts impact both surface and groundwater resources and can lead to reductions in water supply, diminished water quality, crop failure, reduced range productivity, diminished power generation, disturbed riparian habitats, and suspended recreation activities, as well as a host of other associated economic and social activities (Riebsame et al. 1991).

The droughts of the 1930s, 1950s, and 1980s caused great economic and societal losses in the Great Plains of the United States, a region particularly prone to drought (Karl and Koscielny 1982; Diaz 1983; Karl 1983) (Fig. 1). This area shows signs of becoming increasingly vulnerable to drought because of factors such as the increase in cultivation of marginal lands and the escalated use of groundwater from the Ogallala Aquifer, where water withdrawal has exceeded recharge for many years (Glantz 1989; White and Kromm 1987). Estimates for the return intervals for a Great Plains drought of 1930s duration and intensity, based on the properties of the twentieth-century record, vary from 75 to 3000 years (Bowden et al. 1981; Yevjevich 1967). Estimates of this type do not provide a very clear understanding of how rare the severe droughts of the twentieth century were in the context of the last 2000 years, nor whether drought of even greater magnitude is possible.

Paleoclimatic data offer a way to evaluate the severity, duration, and extent of twentieth-century

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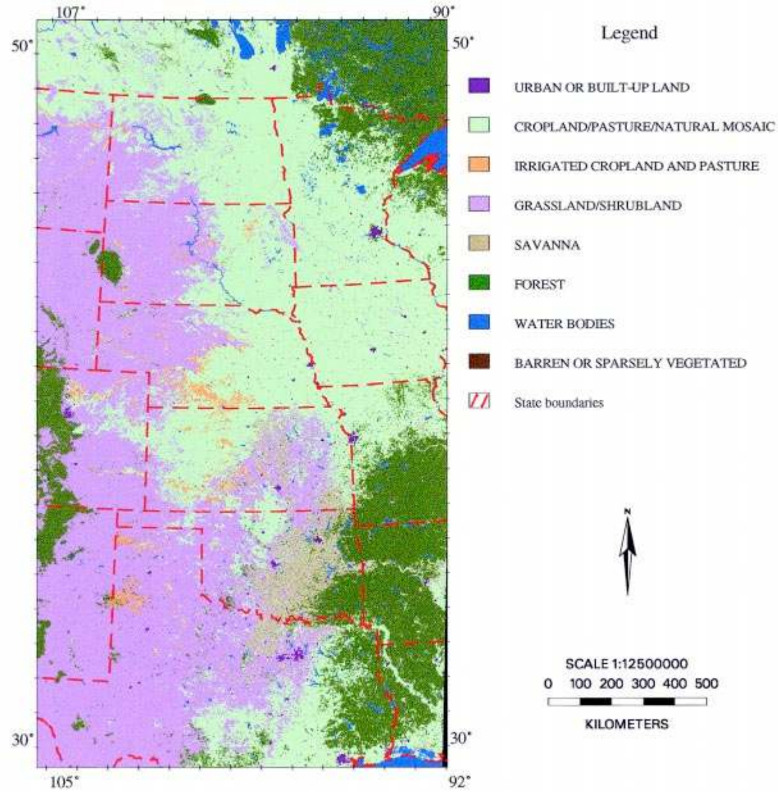

FIG. 1. Advanced Very High Resolution Radiometer Data (AVHRR)–derived 1-km land cover map of the Great Plains (Townshend et al. 1994). Large portions of this area, used for both agriculture and livestock grazing, are highly susceptible to drought.

droughts in the context of the past two millennia (e.g., Overpeck 1996). In this review paper, we bring together evidence of a greater range of drought variability than found in the instrumental record, from all available sources of paleoclimatic data, including historical documents, tree rings, archaeological remains, lake sediment, and geomorphic data, to evaluate the representativeness of twentieth-century droughts in terms of those that have occurred under naturally varying climate conditions of the past several thousand years. The persistence of drought-causing atmospheric conditions is examined through a review of the current literature on twentieth-century droughts, as well as through an examination of whether base-state shifts and low-frequency variation in oceanic/atmospheric systems can yield the persistence needed for the multidecadal- to century-scale droughts of the past. Finally, the prospects of future drought are considered, both in view of the full range of past natural drought variability, and in terms of land use practice and human greenhouse gas–induced climate change.

2. Paleoclimatic evidence for Great Plains drought, A.D. 1–1900

A variety of paleoclimatic data sources can each be tapped to provide key insights into Great Plains drought. Taken together, these proxy data offer a much more complete picture of natural drought variability than offered by instrumental data or any one proxy source alone. A summary of proxy paleodrought data sources and their characteristics is given in Table 1.

a. Seventeenth–nineteenth century drought in the Great Plains

Temperature and precipitation records, extending from 1851 to 1890, exist for early meteorological stations and forts in the Great Plains but are quite fragmented and patchy. Data (locations are shown in green on map in Fig. 2) have been analyzed by Mock (1991), who determined that no drought since 1868 has been as severe as that of the 1930s. However, due to the scarcity of records, he was unable to make a full assessment of a drought in 1860, which may have exceeded the severity of the 1930s drought. Historical accounts from newspapers and diaries provide additional documentation of nineteenth-century drought events. The 1860 drought was reported in Kansas newspapers, which continued to mention the severity of this drought for several decades after the event (Bark 1978). Less severe droughts were also reported in historical documents and early meteorological records for several years around 1860, in the late 1880s, and in the early 1890s (Ludlum 1971; Bradley 1976; Bark 1978). The map in Fig. 2 shows general locations of data sources and drought years documented in historical data, while Fig. 3 (top) shows a time line of these droughts. Accounts of early explorers document periods of blowing sand (an indication of drought conditions) for an area extending from northern Nebraska to southern Texas (Muhs and Holliday 1995). These areas are shown in brown in the map in Fig. 2, along with dates of documented

Table 1. Paleodrought data sources and characteristics.

Proxy data source	Continuous record?	Length of records	Resolution	Dating accuracy	Spatial coverage*	Limitations and potential biases
Early instrumental	Not always	Years–decades	Daily–monthly	Day–month	Local	Quality of instruments and collection of data inconsistent
Historical accounts	No	Decades–centuries	Daily–seasonal	Day–season	Local–regional	Observations subject to human perspective
Tree rings	Yes	Centuries–millennia	Seasonal–annual	Year	Local to ~300 km	(a) Dry extremes more reliably represented than wet, (b) quantitative reconstructions limited to "analog" conditions within range of instrumental variations
Lake sediments	Usually	Millennia	Seasonal (varves)–decadal	±2%–4% (^{14}C)** ±2% (varves)	Local–regional	(a) Suitable lakes not abundant, (b) analog problem as with tree rings
Alluvial sediments	No	Millennia	Century	±2%–5% (^{14}C)	Local	Fluvial response to climate change not well constrained
Eolian sediments	No	Millennia	Century	±2%–6% (^{14}C)	Local–regional	Eolian/soil response to climate not well constrained
Flooded trees	No	Millennia	Annual–century	2%–10% (^{14}C) year (dendrochronologically dated)	Local	(a) Little of this type of data available, (b) rising lake levels inferred from tree deaths, (c) timing of lake level drop difficult to estimate
Archaeological data	No	Centuries–millennia	Annual–century	Decade–century (^{14}C) year (dendrochonologically dated)	Local	Climate inferred from human activity

*Spatial coverages listed refer to general spatial representativeness of a single site. Some proxy sources (e.g., tree rings) have been used to reconstruct much broader-scale climatic variability when sufficient networks of sites are available.

**The age models of the original peer-reviewed papers were used, after checking to make sure they were internally and regionally consistent, based on common ^{14}C half-life and converted to calendar years (i.e., Stuiver and Polach 1977; Stuiver and Becker 1986)

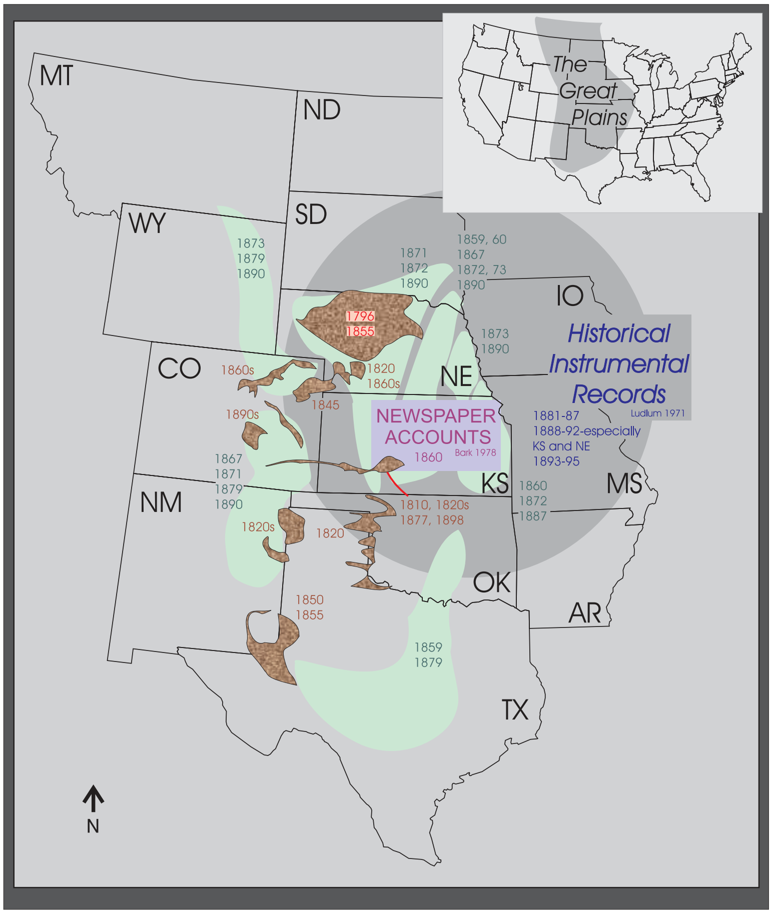

FIG. 2. Locations of sources of historical drought data for the Great Plains, 1795–1895. Green shaded areas represent climate regions based on cluster analysis from Mock's (1991) analysis of nineteenth-century climate records. The dates (dark green) represent years in which droughts were reported in more than one region for two or more consecutive seasons. Brown areas are regions of sand dunes and eolian activity, accompanied by the years (in red) in which active sand movement was reported (Muhs and Holliday 1995). The gray region represents the general region of early meteorological stations from which Ludlum (1971) derived drought years (in blue). Newspaper accounts are from a variety of newspapers in eastern and central Kansas (Bark 1978).

eolian activity, and on the time line in Fig. 3. Several periods of eolian activity were reported in many areas between 1840 and 1865, with other intervals in the late 1700s and early 1800s, as well as at the end of the nineteenth century (Muhs and Holliday 1995). Interestingly, although some eolian activity was reported in the 1930s and 1950s, these twentieth-century droughts were not severe or long enough to cause regional mobilization of dunes (Muhs and Maat 1993; Madole 1994; Muhs and Holliday 1995).

Numerous reconstructions of precipitation and summer drought have been generated for the Great Plains from tree-ring chronologies located in regions proximal to the Great Plains, shown on the map in Fig. 4a (Table 2 contains the key for the symbols in this figure). Regression-based tree-ring reconstructions of climate tend to underestimate extreme values, a consequence of the regression techniques used in producing the reconstructions. However, dry extremes are better replicated than wet extremes, and reconstructions of drought extent and duration are reasonably accurate. For example, in Fig. 5, a comparison of observed and reconstructed mapped Palmer Drought Severity Index (PDSI) (Palmer 1965) values for the severe drought years of 1934 and 1956 shows that drought severity is generally about one PDSI value lower (less severe) for the reconstructed values than for the observed values (Cook et al. 1998). However, the spatial extents of the droughts are well replicated by the reconstructed values, as are drought durations of the 1930s and 1950s events. Although the absolute severity is not duplicated in the tree-ring reconstructions, assessments of the relative severity of twentieth-century droughts compared to droughts in previous centuries can still be made. The amount of variance in the observed drought and precipitation series explained by tree-ring chronologies varies, with average values of about 55%, ranging up to 67% (Table 3). These values are good compared to those obtained in dendroclimatic studies in the semiarid to arid western United States, where trees are notably sensitive to climate. The tree-ring records, of course, are unable to explain all of the drought or precipitation variability because tree growth is usually not solely affected by precipitation or drought conditions (Douglass 1914, 1929).

Many of the tree-ring reconstructions suggest that the droughts of the 1930s and 1950s have been equaled or, in some regions, surpassed by droughts in the past several centuries. This is illustrated in the graphs of PDSI reconstructions from Cook et al. (1996) and Cook et al. (1998) for grid points in eastern Montana, central Kansas, and north-central Texas in Fig. 6. Other studies support this finding. Stockton and Meko (1983) reconstructed annual precipitation for four regions flanking the Great Plains (centered in

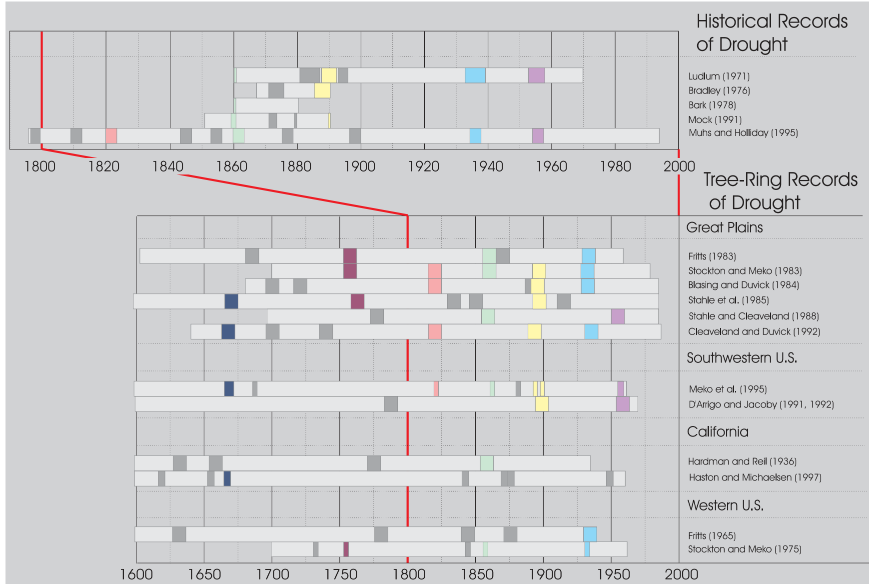

FIG. 3. Paleoclimatic records of Great Plains and western U.S. drought (1600–present) based on historical and tree-ring data. The pale gray horizontal bars reflect the length of the series, and the dark gray and colored bars indicate periods of drought in 3–10-yr increments. Colors mark more widespread droughts that occurred over the same time period in a number of records. The historical droughts are all those reported in the literature. The droughts recorded by tree-ring data are those listed in the literature as the most extreme (e.g., the five most severe 10-yr drought periods in a record). For the few reconstructions that were not accompanied by specific lists of droughts, periods of drought that equaled or exceeded twentieth-century droughts are shown.

Iowa, Oklahoma, eastern Wyoming, and eastern Montana). Although they found the individual years of 1934, 1936, and 1939 to be among the driest 10 of 278 years investigated (1700–1977), they found several periods of widespread prolonged drought (3–10 years) that equaled or surpassed the 1930s drought in intensity and duration: the late 1750s, early 1820s, early 1860s and 1890s. Periods of extreme drought revealed by other dendrochronological assessments for the west-central Great Plains coincide with these periods (Weakly 1965; Wedel 1986; Lawson 1970; Lawson and Stockton 1981). Stahle and Cleaveland's (1988) reconstructions of June PSDI in Texas showed the most severe and uninterrupted drought since 1698 was the 1950s drought, but the three driest decades (with some interspersed years of nondrought conditions), by decreasing severity, were 1855–64, 1950–59, and 1772–81. Another dendroclimatic study from the southern plains found prolonged (10 years or more) droughts in Arkansas around 1670, 1765, 1835, 1850, and 1875 that were comparable to twentieth-century events (Stahle et al. 1985), whereas a study in the Texas–Oklahoma–Arkansas region found the drought of the 1950s was exceeded only in 1860 in the last 231 years (Blasing et al. 1988), a particularly noteworthy year in the historical data, as mentioned above. In a reconstruction of precipitation for the corn belt of Iowa and Illinois, no droughts in the past 300 years were found to be appreciably worse than the 1930s drought, but two were of about the same magnitude (late 1880s–1890s and around 1820) (Blasing and Duvick 1984). Reconstructions of precipitation in Iowa (1640–1982) indicated that four 10-yr periods were drier than the period 1931–1940, and in order of decreasing dryness, these were 1816–25, 1696–1705, 1664–73, and 1735–44 (Cleaveland and Duvick 1992). Figure 3 summarizes the timing of droughts in these dendroclimatic studies and illustrates the regional impacts of some of these periods of drought.

The widespread and persistent nature of some of the severe Great Plains droughts of the past three centuries can be compared to twentieth-century droughts using the maps of tree-ring reconstructions of gridded PDSI for the United States (Cook et al. 1996; see maps of other droughts in the past three centuries at the NOAA/NESDIS Web site at http://www.ngdc.noaa.gov/paleo/drought.html). For example, Fig. 7 shows that the prolonged drought that centered around 1820 appears to be at least equivalent in extent and duration to the 1950s drought (Cook et al. 1998). The latter part of the 1750s was also a period of prolonged and widespread drought, comparable to those of the twentieth century.

Multiple sources of proxy data, including tree-ring reconstructions and historical records and accounts, work together to confirm the occurrence of several nineteenth-century droughts, as shown in Fig. 6. The 1820s drought is one of several that is documented in the historical accounts of eolian activity (Muhs and Holliday 1995), as well as in tree-ring reconstructions (Lawson and Stockton 1981; Stockton and Meko 1983; Blasing and Duvick 1984; Cleaveland and Duvick 1992; Cook et al. 1998). The drought that occurred about 1860 is notable in much of the historical data [eolian activity, newspapers, and early meteorological records (Ludlum 1971; Bark 1978; Mock 1991; Muhs and Holliday 1995)] and in drought reconstructions for the central and northern Great Plains (Fritts 1983; Stockton and Meko 1983), as well as in eastern California (Hardman and Reil 1936) and throughout the southwestern and western United States (Stockton and Meko 1975; Meko et al. 1995). While the historical evidence of eolian activity suggests these two nineteenth-century droughts were more severe than twentieth-century droughts, it is not clear from the dendrochronological records that nineteenth-century droughts were indisputably more extreme. Rapid increases in Native American and Euro-American populations as well as bison populations may have led to severe land cover degradation and increased eolian activity between 1820 and 1850 (West 1997). In any case, it is clear that major multiyear Great Plains drought has occurred naturally once or twice a century over the last 400 years.

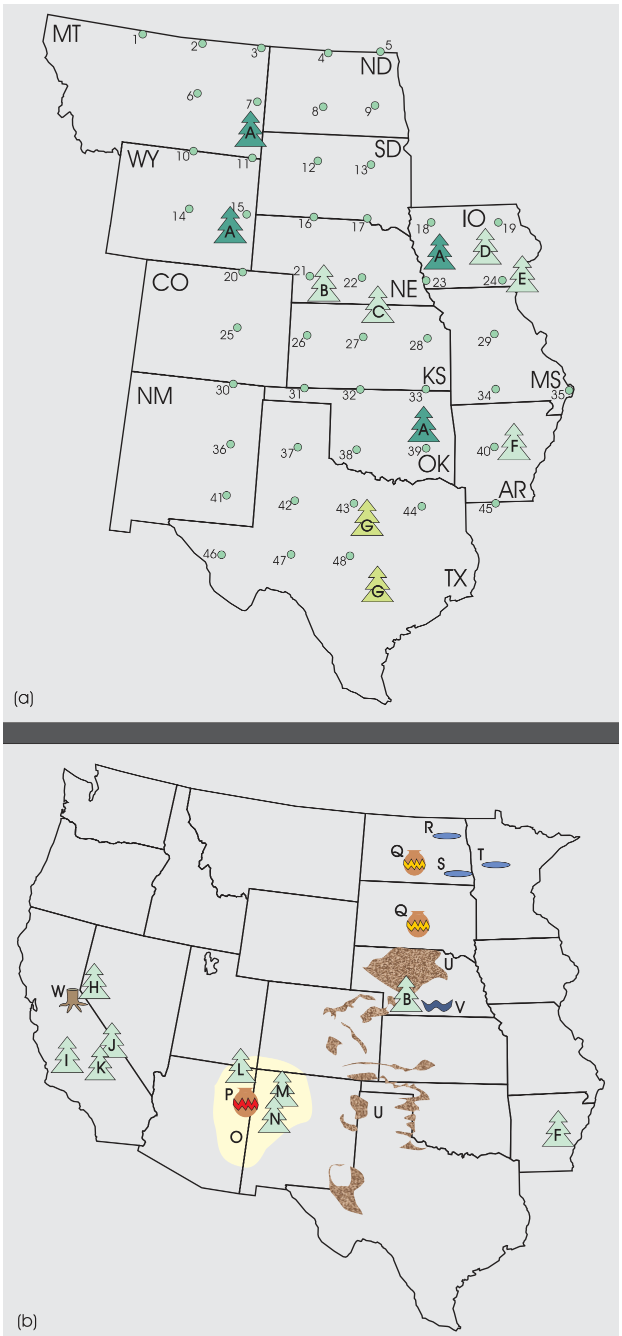

FIG. 4. (a) Locations of drought-sensitive tree-ring chronologies and reconstructions of precipitation or drought in the Great Plains. Numbered dots are locations of Cook et al.'s (1996) gridded PDSI reconstructions. The key for lettered symbols is in Table 2. Statistical relationships (explained variance) between observed and reconstructed series of tree-ring chronologies are listed by author and/or grid number in Table 3. Note that while reconstructions are for regions within the Great Plains, the reconstructions are generated from trees located in areas flanking the Great Plains reconstructions (the exception is Weakly's southwestern Nebraska chronology). Tree growth reflects large-scale climate variations, and thus trees proximal to the Great Plains have been used successfully to reconstruct climate variations in this region. (b) Locations of many of the paleoclimatic records documenting drought in the Great Plains and western United States for the period A.D. 1–1600. The key for lettered symbols is in Table 2.

b. Thirteenth-to-sixteenth-century megadroughts

Prior to the seventeenth century, the availability of high-resolution proxy data for Great Plains drought is reduced, but useful information can still be gleaned from a wide variety of proxy data, including data from other areas of the western United States (Table 1). We include these more distant records because they provide corroborative information for droughts documented in the few available Great Plains records and allow an assessment of the extent of some of these great droughts. Instrumental records indicate that the major droughts impacting the Great Plains in the twentieth century also affected areas of the western United States (see Fig. 5); thus we feel that our use of proxy data from the western United States to support evidence of drought in the Great Plains is justified. There are few tree-ring chronologies for the Great Plains that extend prior to the 1600s, but there are long chronologies for other areas in the western United States that reflect spatially extensive droughts. Other proxy data with a coarser temporal resolution or less accurate temporal control than tree-ring data include those from lake sediment, alluvial, eolian, and archaeological

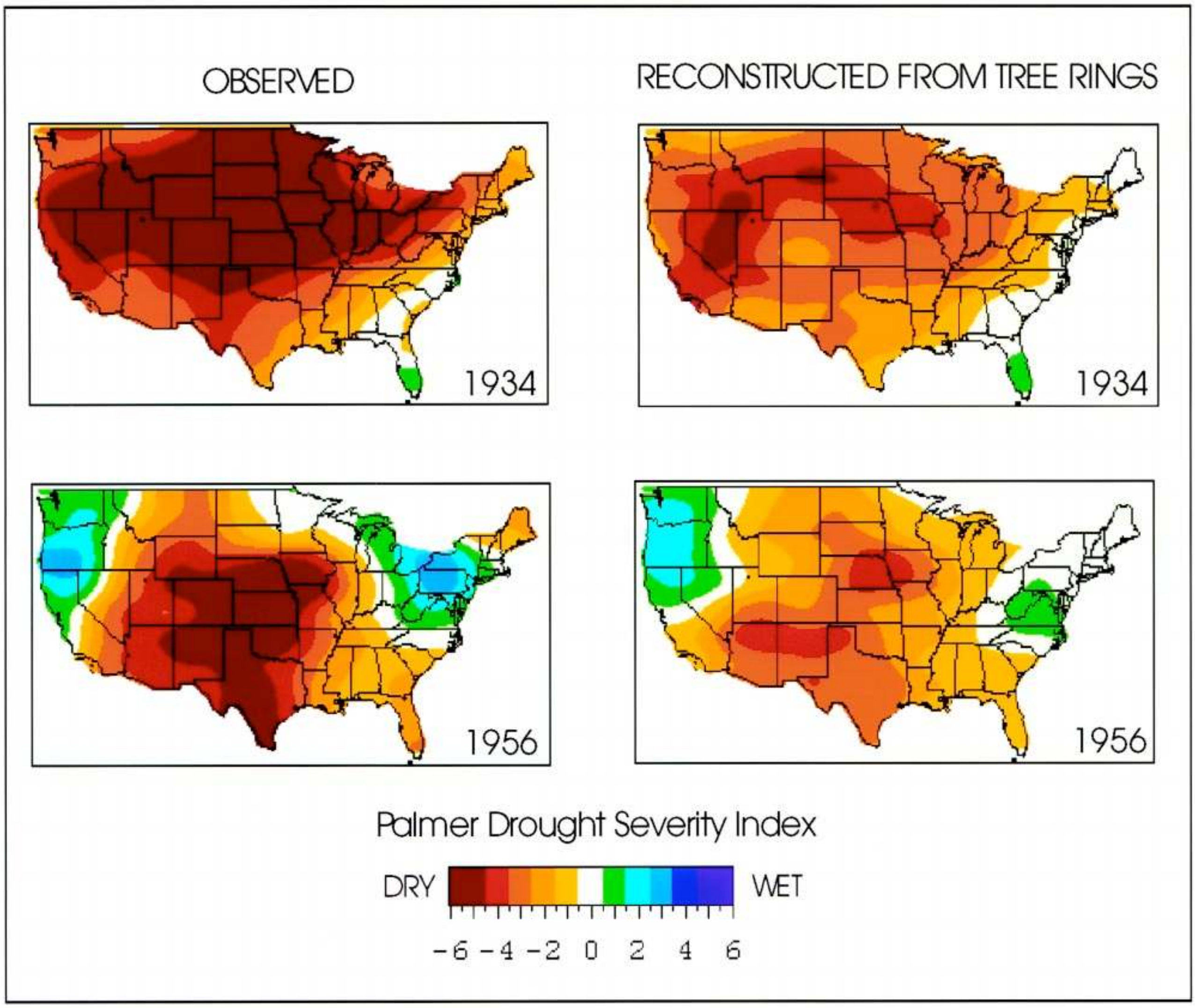

FIG. 5. Comparison of observed and tree-ring reconstructed PDSI values for two of the most extreme drought years in the twentieth century, 1934 and 1956. Although the severity of these droughts is not fully captured by the tree-ring reconstructions, the reconstructions duplicate the spatial extent and duration (see Fig. 6) of these droughts. Images are from the NOAA/NESDIS Web site (see text) (Karl et al. 1990; Guttman 1991; Cook et al. 1996).

sources. These data provide evidence to support the droughts documented in the few available extralong Great Plains tree-ring records, as well as for the period prior to that covered by tree-ring reconstructions. Locations of these proxy records are shown in Fig. 4b and described in Table 2. Thus, although the rapid decline in the number of annually resolved drought records prior to about 1600 makes it difficult to resolve interannual variations in drought frequency and magnitude, paleoclimatic records can provide key constraints on the full range of natural decadal to interdecadal drought variability.

During the thirteenth to sixteenth centuries, there is evidence for two major droughts that likely significantly exceeded the severity, length, and spatial extent of twentieth-century droughts. The most recent of these "megadroughts" occurred throughout the western United States in the second part of the sixteenth century. The dendrochronological records that reflect this drought and their locations are indicated in Figs. 8 (time line) and 9a (map, key in Table 2). This drought is indicated in a southwestern Nebraska chronology (Weakly 1965) as well as in a reconstruction of Arkansas drought (Stahle et al. 1985). Weakly (Wedel 1986) notes two periods of what he terms "very severe" drought, from 1539 to 1564 and from 1587 to 1605. Stahle et al. (1985) suggest that the period 1549–77 was likely the worst drought in Arkansas in the past 450 years. Other tree-ring reconstructions for the broader western United States reflect this drought,

TABLE 2. Key for map symbols in Figs. 4a, 4b, 9a, and 9b.

Map letter	Reference	Proxy data source	Proxy variable	Location
A	Stockton and Meko (1983)	Tree rings	Precipitation	Four regions flanking Great Plains
B	Weakly (1965)	Tree rings	Precipitation	Southwest NE
C	Fritts (1983)	Tree rings	Precipitation	Central plains
D	Cleaveland and Duvick (1992)	Tree rings	Drought	IA
E	Blasing and Duvick (1984)	Tree rings	Precipitation	Western corn belt
F	Stahle et al. (1985)	Tree rings	Drought	AR
G	Stahle and Cleaveland (1988)	Tree rings	Drought	North and south TX
H	Hardman and Reil (1936)	Tree rings	Flow	Truckee River, CA
I	Haston and Michaelsen (1997)	Tree rings	Precipitation	South CA
J	Hughes and Graumlich (1996)	Tree rings	Precipitation	White Mts./south Great Basin
K	Hughes and Brown (1992)	Tree rings	Drought	Central CA
L	Stockton and Jacoby (1976), Meko et al. 1995	Tree rings	Flow, precipitation	Colorado River, AZ, NM, CO, UT
M	D'Arrigo and Jacoby (1991, 1992)	Tree rings	Precipitation	Northwest NM
N	Grissino-Mayer (1996)	Tree rings	Precipitation	Northwest NM
O	see Dean (1994)	Tree rings	Precipitation	South CO Plateau
P	Euler et al. (1979), Dean et al. (1985), Peterson (1994)	Archaeological remains	Drought	Four Corners area
Q	Lehmer (1970), Wendland (1978)	Archaeological remains	Drought	Missouri Valley
R	Fritz et al. (1991)	Lake sediments	Salinity	Devil's Lake, ND
S	Laird et al. (1996), Laird et al. (1998)	Lake sediments	Salinity	Moon Lake, ND
T	Dean et al. (1994), Dean (1997)	Lake sediments	Aridity	Elk Lake, western MN
U	Muhs and Holliday (1995) and others	Eolian sediments	Drought	Western Great Plains
V	Brice (1966), May (1989), Martin (1992)	Fluvial sediments	Xeric conditions	Southwest NE, north KS
W	Stine (1994)	Flooded stumps	Lake levels	Sierra Nevada, CA

including a number of reconstructions from the Four Corners region (the junction of Arizona, New Mexico, Utah, and Colorado) (Rose et al. 1982; D'Arrigo and Jacoby 1991, 1992; Grissino-Mayer 1996). In their 1000-yr-long reconstruction of winter precipitation D'Arrigo and Jacoby (1991, 1992) found the 1950s drought was surpassed only by a 22-yr drought in the late 1500s. The reconstruction of Colorado River flow for 1520–1961 shows the period 1579–98 to reflect the longest and most severe drought in this record (Stockton and Jacoby 1976; Meko et al. 1995). In the White Mountains of eastern California, precipitation reconstructed from bristlecone pine shows a moderate drought in the late sixteenth century (Hughes and

TABLE 3. Variance in observed precipitation and drought series explained (r^2) or shared (r) by tree-ring chronologies.

Study	Region	Variable*	Years	Variance explained (r^2) or shared (r)**
Weakly (1965)	Western NE	Annual precipitation at North Platte	1210–1965	$r = 0.63$ (ring widths) $r = 0.75$ (ring area)
Fritts (1983)	Central Great Plains	Annual regional precipitation	1600–1963	$0.20 \le r^2 < 0.40$
Stockton and Meko (1983)	Eastern MT Eastern WY IA OK	Annual regional precipitation	1700–1977	$r^2_{adj} = 0.52$ $r^2_{adj} = 0.54$ $r^2_{adj} = 0.44$ $r^2_{adj} = 0.40$
Blasing and Duvick (1984)	IA, IL	Annual regional precipitation	1680–1980	$r^2 = 0.62$
Stahle et al. (1985)	AK	June PDSI	1531–1980	$r^2_{adj} = 0.40$
Stahle and Cleaveland (1988)	North TX South TX	June PDSI	1698–1980	$r^2_{adj} = 0.59$ $r^2_{adj} = 0.60$
Cleaveland and Duvick (1992)	IA	July PDHI	1640–1982	$r^2_{adj} = 0.67$
Cook et al. (1996)	United States (Great Plains gridpoint results reported here; see Fig. 5a for locations)	Summer PDSI	1700–1979	
	Grid points 34, 35, 47		$r^2 \ge 0.60$	
	6, 7, 10, 11, 19, 24, 25, 29, 33, 39, 40, 42, 43, 44, 46, 48			$0.50 \le r < 20.60$
	1, 2, 3, 14, 15, 18, 20, 22, 23, 26, 27, 28, 30, 32, 36, 38, 41			$0.40 \le r^2 < 0.60$
	8, 9, 12, 13, 16, 17, 21, 31, 37, 45			$0.30 \le r^2 < 0.40$
	4,5		$r^2 < 0.30$	

*PDSI: Palmer Drought Severity Index; PHDI: Palmer Hydrological Drought Index (Palmer 1965),
**The use of r^2 or r^2_{adj} depends on how results were reported in specified studies.

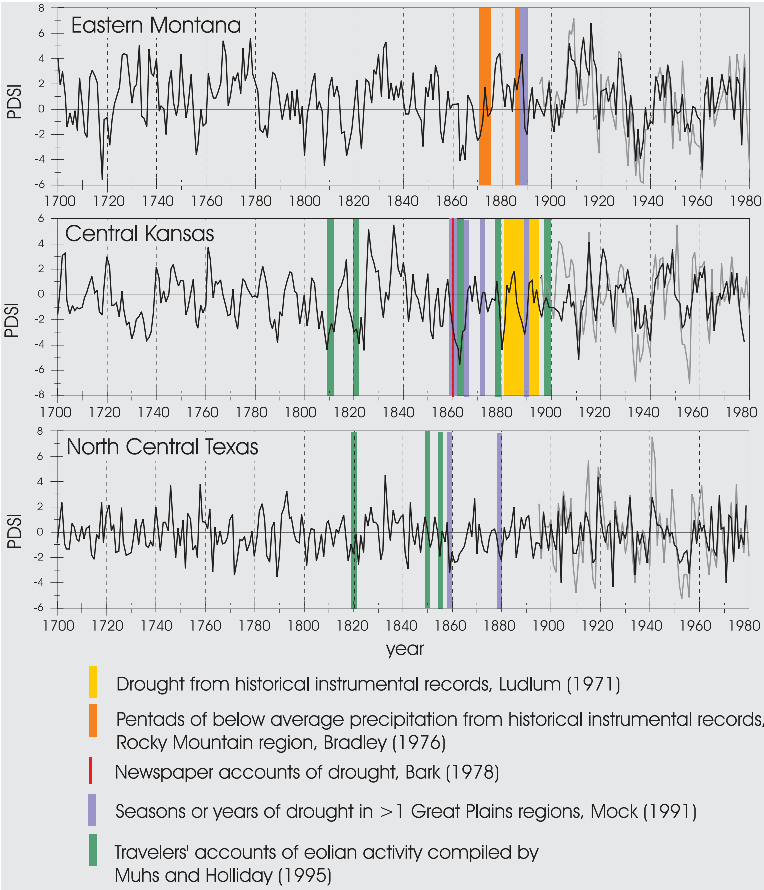

FIG. 6. Palmer Drought Severity Index records (1700–1979) for eastern Montana (grid point 7 in Fig. 4a) (top), central Kansas (grid point 27) (middle), and north-central Texas (grid point 42) (bottom), as reconstructed using tree-ring data [Cook et al. 1996; see also the NOAA/NESDIS Web site (see text)]. Also shown are the observed PDSI values for grid points (light gray lines). The colored vertical bars indicate years of drought from historical accounts.

Graumlich 1996). A flow reconstruction for the Truckee River in eastern California reflects this sixteenth-century drought (Hardman and Reil 1936), as do reconstructions of precipitation for northern and southern regions of California west of the Sierra Nevada (Haston and Michaelsen 1997). Additionally, reconstructions of western U.S. regional precipitation indicate a drought beginning in the southwest around 1565 and spreading to the entire western United States by 1585 (Fritts 1965), corresponding to drought evidence in both lake sediment data from western Minnesota (Dean et al. 1994) and scarcity of old, living conifers established before about 1600 in the southwest (Swetnam and Betancourt 1998). Recent analysis of eolian sedimentation dates in the Wray dune field of eastern Colorado by Muhs et al. (1997) estimate the most recent period of eolian activity to have occurred in the past ~400 years (^{14}C yr before present), while Stokes and Swinehart (1997) document an optically dated period of eolian activity in the Nebraska Sand Hills that also coincides with the late sixteenth-century drought. Eolian activity is primarily due to drought severe enough to remove vegetation (Muhs and Holliday 1995), and the late 1500s drought was likely severe and long enough to have cleared sand deposits of live vegetation.

The other megadrought of the thirteenth to sixteenth centuries occurred in the last quarter of the thirteenth century. The time line in Fig. 8 shows the tree-ring records that reflect this drought, while Fig. 10 shows the proxy records that reflect this drought in a coarser temporal context (locations of the proxy records are shown in Fig. 9b, key in Table 2). Most of the proxy records mentioned for the sixteenth-century drought that extend back to the thirteenth century also record this severe multidecadal drought, including tree-ring chronologies and/or reconstructions for southwestern Nebraska (Weakly 1965), northern New Mexico (Grissino-Mayer 1996), the Four Corners area (Rose et al. 1982), and the White Mountains (Hughes and Graumlich 1996). Weakly (1965) reported a 38-yr drought from 1276 to 1313 in his southwestern Nebraska tree-ring chronology, the longest drought in the past 750 years. Other less finely resolved proxy data also testify to the occurrence of this drought, which in some areas appears to have rivaled or exceeded even the sixteenth-century drought and was almost certainly of much greater intensity and duration than any drought of the twentieth century. Recent analysis of eolian sediments in the Nebraska Sand Hills suggests an onset of eolian activity beginning within the past 800 ^{14}C years (Muhs et al. 1997). A period of drought at this time is documented in the varve record of western Minnesota (Dean et al. 1994). Archaeological data from the Great Plains and Four Corners areas also provide documentation of this drought (Bryson et al. 1970; Lehmer 1970; Wendland 1978; Euler et al. 1979; Dean et al. 1985; Dean 1994; Peterson 1994). In the Southwest this drought, sometimes referred to as the “Great Drought,” coincided with the abandonment of Anasazi settlements, redis-

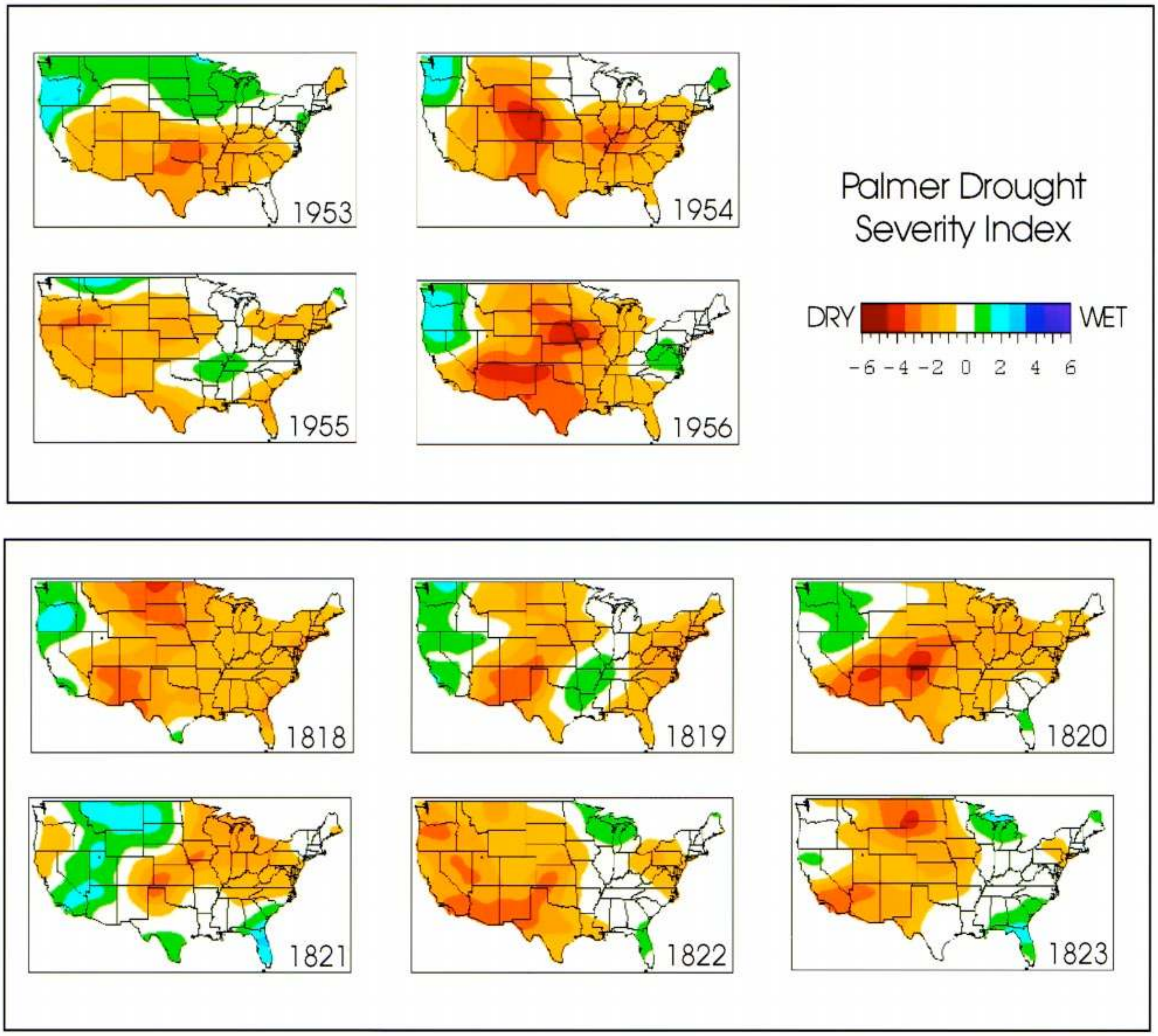

FIG. 7. Comparison of duration and extent of two severe droughts, the 1950s drought (top) and the drought centered on 1820 (bottom), both reconstructed from tree-ring chronologies (Cook et al. 1996; Cook et al. 1998).

tribution of populations, and widespread societal reorganization (Douglass 1935; Dean 1994). This period of drought is also reflected in unprecedentedly low lake levels as reconstructed through the dating of tree stumps rooted in what are today bottoms of several streams and lakes in the Sierra Nevada of eastern California (Stine 1994).

Several tree-ring reconstructions allow a temporal evaluation of the thirteenth and sixteenth-century megadroughts relative to more recent droughts (Dean 1994; Grissino-Mayer 1996; Hughes and Graumlich 1996). Of the reconstructions that reflect both sixteenth- and thirteenth-century droughts, the sixteenth-century drought appears to have been the most severe and persistent drought in the Southwest in the past 1000–2000 years, whereas the thirteenth-century drought was the most persistent and severe drought in the California mountain ranges and, likely, the Great Plains (Weakly 1965). It is more difficult to evaluate the spatial extent of the two major paleodroughts. At a minimum, both droughts appear to have impacted the Great Plains, Southwest, southern and western Great Basin, and Sierra Nevada (Figs. 9a,b). A survey of other tree-ring chronologies for the northwestern Great Basin and northeastern Utah (from the International Tree-Ring Data Base, World Data Center-A, Paleoclimatology, Boulder, Colorado) shows marked periods of low growth in the latter part of the thirteenth century in

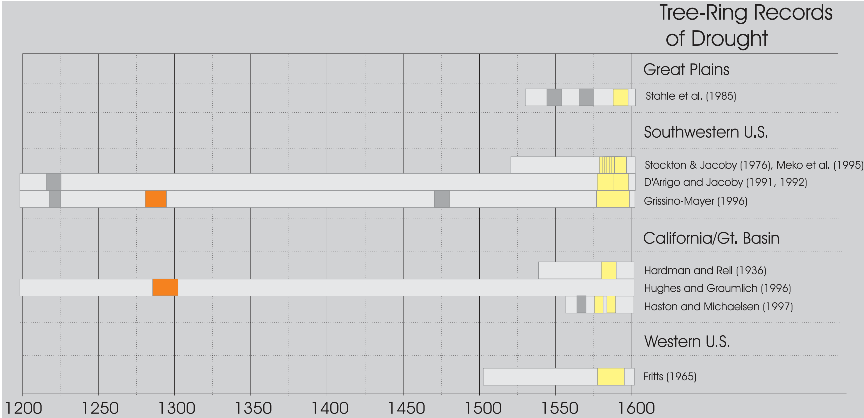

FIG. 8. Paleoclimatic records of Great Plains and western U.S. drought (thirteenth century through the sixteenth century) based on tree-ring data. As in Fig. 3, the pale gray horizontal bars reflect the length of the series, and the dark gray and colored bars indicate periods of drought. Yellow bars mark records that reflect the late fifteenth-century drought, while orange bars mark records that reflect the late thirteenth-century drought.

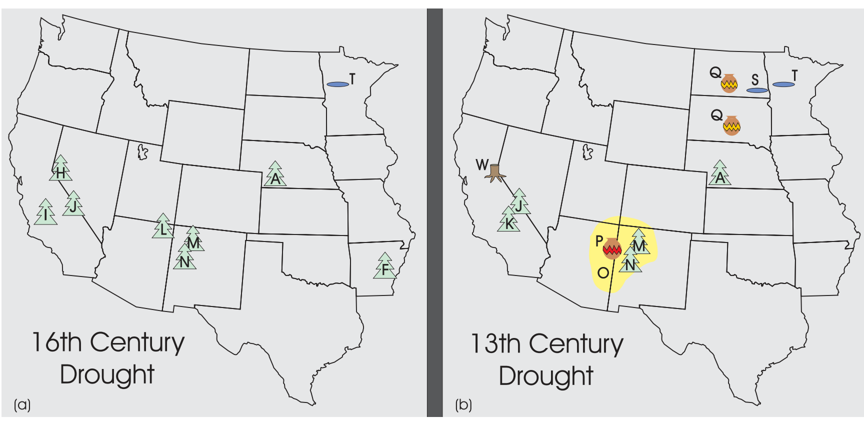

FIG. 9. (a) Location of paleoclimatic records that document the late sixteenth-century drought. Proxy data reflect the widespread nature of this drought, which was especially notable in the Southwest but also detected in records from areas ranging from the northeastern and southeastern Great Plains to the California coast. Most proxy records indicate the duration of this drought was close to 20 yr. (b) Location of paleoclimatic records that document the late thirteenth-century drought. Fewer proxy records are available for this drought, but most that do exist for this period reflect drought that was at least 25 yr in duration and that appears to have ranged from the northern Great Plains, through the Southwest, and to the southern end of the Sierra Nevada. The key for the lettered symbols is in Table 2.

these areas as well. For comparison, severe drought conditions in 1934 (see Fig. 5) also covered most of these areas, but the 1934 conditions were part of a drought that lasted only several years, as opposed to decades (Karl et al. 1990; Guttman 1991; Cook et al. 1996).

c. *Evidence for drought, A.D. 1–1200*

The temporal resolution and interpretation of most proxy data for the period A.D. 1–1200 make it difficult to assess droughts of this period in the same way as more recent droughts. Many of the proxy records that exist for this period extend back several millennia or more. Because of their great length, even proxy records with annual resolution are typically ana-lyzed in terms of multidecadal- to century-scale variations. Consequently, our assessment of drought within this time frame focuses on low-frequency (decade to century scale) drought variability relative to the twentieth century. However, even given this low-frequency perspective, proxy records suggest that droughts of the period A.D. 1–1200 occurred on a scale that has not been duplicated since Europeans came to the Great Plains.

At least four periods of widespread drought between A.D. 1 and 1200 are found in a variety of proxy data from the Great Plains and the western United States as illustrated in Fig. 10. Of these four, the most recent is the least well documented. A limited number of proxy records suggest that a drought began around mid A.D. 1100, although it is difficult to separate this drought from the late thirteenth-century drought in some of the less finely resolved records. This drought is suggested in the Southwestern archaeological data as a forerunner to the more severe late 1200s drought (Euler et al. 1979; Dean et al. 1985) and is also documented in White Mountains and Four Corners tree-ring records (LaMarche 1974; Rose et al. 1982), in a preliminary Colorado Front Range tree-ring chronology (P. Brown 1997, personal communication), and in western Minnesota lake sediments (Dean et al. 1994; Dean 1997). Archeological and pollen data have also been cited as evidence for an onset of markedly drier conditions in the northern Great Plains about this time

(Lehmer 1970; Wendland 1978), and in northwestern Iowa after about A.D. 1100 and intensifying by A.D. 1200 (Bryson and Murray 1977). The next major drought is characterized primarily by an onset of eolian activity in the western Great Plains. It is difficult to determine the exact date of onset, but activity began sometime after ~A.D. 950 (Forman et al. 1992; Forman et al. 1995; Madole 1994, 1995; Muhs et al. 1996). Other proxy data that help confirm this period of drought are those from North Dakota lake sediments (Laird et al. 1996; Laird et al. 1998) and alluvial sediment records from western Nebraska and Kansas (Brice 1966; May 1989; Martin 1992). Although there is dendroclimatic and lake-level evidence of drought in the Sierra Nevada and White Mountains between ~A.D. 900 and 1100, (LaMarche 1974; Stine 1994; Hughes and Graumlich 1996), there is no evidence of an onset of drought conditions occurring in the Southwest at this time.

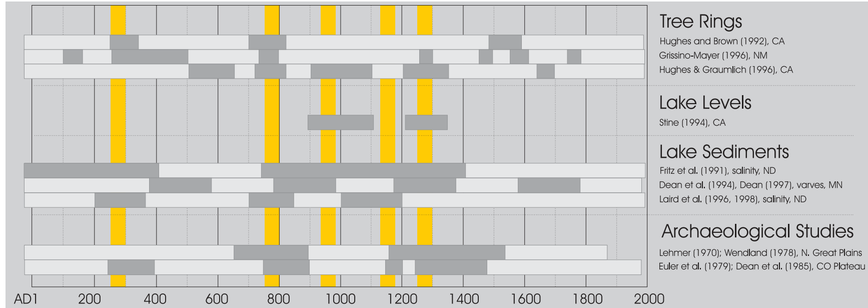

FIG. 10. Paleoclimatic records of Great Plains and western U.S. century-scale drought, A.D. 1–present, as recorded in a variety of paleoclimatic data. The pale gray horizontal bars reflect the length of the series, and the dark gray indicate periods of drought. Orange vertical bars represent multidecadal droughts that appear to have been widespread.

The third major drought episode of the A.D. 1–1200 period occurred roughly between A.D. 700 and 900. In archaeological evidence in the Four Corners area, A.D. 750 was the starting date for a drought that lasted several centuries (Euler et al. 1979; Dean et al. 1985; Peterson 1994), and a tree-ring reconstruction of drought for New Mexico also reflects this drought (Grissino-Mayer 1996). Drought is recorded in western Minnesota lake varves at this time (Dean et al. 1994; Dean 1997) while North Dakota lake sediments indicate drought conditions typified the period A.D. 700–850 (Fig. 11; Laird et al. 1996; Laird et al. 1998). In another more coarsely resolved record of lake sediments in North Dakota, high salinity conditions are indicated to have begun about this time and continued through the fifteenth century, a period containing the droughts of the tenth, twelfth, and late thirteenth centuries (Fritz et al. 1991). In the central California drought record from giant sequoia, the period A.D. 699–823 had the highest drought frequency in the past 2000 years (Hughes and Brown 1992). Drought appears to have occurred in the White Mountains about this time as well (Hughes and Graumlich 1996). The onset of the earliest of these four droughts occurred about the middle of the third century and appears to have lasted up to three centuries. A dendroclimatic reconstruction of precipitation for northern New Mexico (Grissino-Mayer 1996) shows this to be a period of consistently average to below-average precipitation until about A.D. 500. Drought-sensitive giant sequoia in central California suggest that the period A.D. 236–377 was one of the three periods with the highest frequency of drought within the past two millennia (Hughes and Brown 1992). During the closely corresponding period, A.D. 200–370, more frequent drought conditions were indicated by high lake salinity in North Dakota lake sediments (Laird et al. 1996; Laird et al. 1998). Archaeological remains in

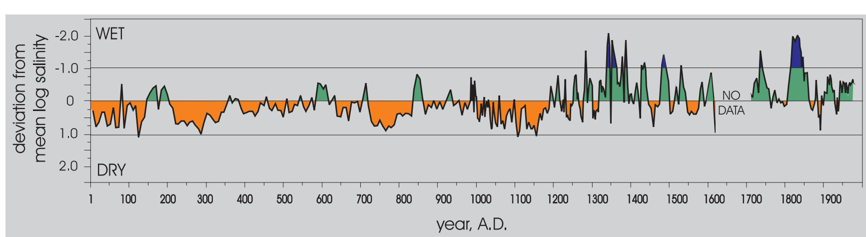

FIG. 11. North Dakota Moon Lake salinity record, here spanning A.D. 1–1980 (Laird et al. 1996). Deviations from the mean (based on past 2300 yr) log salinity values are shown with negative values indicating low salinity and wet conditions and positive values indicating high salinity and dry conditions. Note the shift in salinity values around A.D. 1200, likely reflecting a change in drought regime from more frequent, intense droughts prior to A.D. 1200 to relatively wetter conditions in the last 750 yr. The average temporal resolution of the chronology is about one sample per five years, with an estimated error in the absolute chronology of ±50–60 yr. The 92-yr gap in the data from 1618 to 1710 is due to desiccation in this section of the core.

the Four Corners (Euler et al. 1979; Dean et al. 1985) indicate drought conditions from A.D. 250 to 400.

3. The paleoclimatic perspective: A summary

Paleoclimatic data provide evidence that twentieth-century droughts are not representative of the full range of drought variability that has occurred over the last 2000 years. The collection of dendroclimatic reconstructions for the Great Plains region suggests that the severe droughts of the twentieth century, although certainly major in terms of their societal and economic impacts, are by no means unprecedented in the past four centuries. Moreover, when all proxy data, including historical accounts of eolian activity, are considered, it is likely that droughts of a magnitude at least equal to those of the 1930s and 1950s have occurred with some regularity over the past 400 years. A look farther back in time reveals evidence that the multidecadal drought events of the late thirteenth and/or sixteenth centuries were of a much greater duration and severity than twentieth-century droughts. Interestingly, in the interval between these two big droughts, there is little evidence of severe and/or widespread drought.

Laird et al. (1996) and Laird et al. (1998) suggest that their North Dakota lake sediment data reflect a drought regime shift occurring around A.D. 1200, with droughts prior to this time characterized by much greater intensity and frequency (Fig. 11). Although the type of proxy data that extend back several thousand years tend to have a decadal- to century-scale temporal resolution and dating accuracy that confounds close comparisons, the few annually resolved paleoclimatic records that do exist for this period provide some evidence for longer periods of drought or periods of more frequent drought prior to the thirteenth century (LaMarche 1974; Dean 1994; Grissino-Mayer 1996; Hughes and Graumlich 1996). Several tree-ring records and reconstructions for the Southwest and the White Mountains/Great Basin region support the idea of a major drought regime shift after the late thirteenth-century drought. The timing is somewhat later than suggested by Laird et al. (1996) and Laird et al. (1998), but the difference may be due to the greater precision in dendrochronological dating compared to radiocarbon dating. For the most part, these longer records have been analyzed in terms of low-frequency variations, but twentieth-century variations can still be evaluated in the context of the previous 2000 years. Dendroclimatic evidence suggests that many droughts prior to the late thirteenth-century drought were at least decades in duration. In contrast, the droughts since the thirteenth-century event apparently have tended to be a decade or less in duration, with the exception of the late sixteenth-century multidecadal drought in the Southwest. The North Dakota lake sediment record along with these tree-ring records from the Southwest, the Great Basin, and the White Mountains suggest that a drought regime shift may have occurred not only in the Great Plains, but over much of the western United States as well. The evidence for a drought regime shift around 700 years ago is intriguing, but more investigations incorporating millennium-length records of highly resolved, precisely dated paleoclimatic data are needed to confirm and understand the full nature and extent of this event.

An assessment of the available proxy data suggests that droughts of the twentieth century have been characterized by moderate severity and comparatively short duration, relative to the full range of past drought variability. This indicates the possibility that future droughts may be of a much greater severity and duration than what we have yet experienced. It is imperative to understand what caused the great droughts of the past 2000 years and if similar droughts are likely to occur in the future.

4. The causes of Great Plains drought

An inquiry into the mechanisms behind Great Plains drought begins with an examination of precipitation climatology and the atmospheric conditions associated with twentieth-century drought. The semiarid to subhumid climate of the Great Plains is influenced by several different air masses, each with spatially and seasonally varying impacts on the region: dry westerly flow of air from the Pacific; the cold, dry arctic air masses from the north; and the warm, moist tropical air masses from the south (Bryson 1966; Bryson and Hare 1974). The polar jet stream brings Pacific moisture to the area in the cool season, but the region is generally quite dry in winter (Doesken and Stanton 1992). In summer, although the central Great Plains is under the drying influence of a strong subtropical ridge, moisture is drawn into the area from the Gulf of Mexico by the Great Plains nocturnal low-level jet (LLJ). The LLJ is a synoptic-scale feature associated with convective storm activity and represents the intrusion of the Atlantic anticyclonic subtropical gyre (associated with the Bermuda high) into the interior United States (Tang and Reiter 1984; Helfand and Schubert 1995; Higgins et al. 1997). Another source of summer precipitation is mesoscale convective complexes (MCCs), which can contribute between 30% and 70% of the total warm season precipitation over much of the Great Plains (Fritsch et al. 1986). Less consistently, synoptic-scale upper-level disturbances also contribute summertime moisture (Helfand and Schubert 1995; Mock 1996). In spring, the mixing of cold air masses from the Arctic with warm, moist air tropical masses from the Gulf of Mexico causes an increase in precipitation (Bryson 1966). During this season, meridional troughs and cutoff lows in midlatitude frontal systems draw moisture from the Gulf of Mexico into the western Great Plains (Hirschboeck 1991). Fall is a relatively dry season as Pacific air dominates most of the region (Bryson 1966; Mock 1996).

Drought in the Great Plains can occur during any season, but since late spring and summer are the seasons when most of the precipitation falls, these are the most important drought seasons. In general, Great Plains drought is characterized by a semipermanent mid- to upper-tropospheric anticyclone over the plains, sustained by anticyclones in both the eastern central Pacific and eastern central Atlantic and accompanied by intervening troughs (Namias 1955, 1983) that can persist throughout the summer. Under this configuration, the jet stream is diverted to the north and the plains anticyclone blocks moisture from the Gulf (Borchert 1950). The Great Plains region is commonly not homogeneous with respect to drought because of the spatially variable influence of the circulation features related to seasonal precipitation (Karl and Koscielny 1982). Figure 12 illustrates this by showing the spatial distribution of PDSI values for three different twentieth-century drought years and accompanying PDSI time series for three different regions. The position of the semipermanent ridge of high pressure appears to be particularly important. At times when the ridge is displaced east of its usual position over the west-central United States, Gulf of Mexico moisture is unable to penetrate into the central United States (Oglesby 1991), but there appear to be varying degrees of displacement. The 1950s drought was most severe in the southern Great Plains, suggesting a complete failure of Gulf of Mexico moisture to enter the central United States (Borchert 1971). In contrast, the 1988 drought was characterized by an inverted U shape, in which drought was largely restricted to the northern Plains as well as the west coast and southeastern United States, while Gulf moisture was able to find a way into the south-central United States (Oglesby 1991) (Fig. 12). Once a drought-inducing circulation pattern is set up, dry conditions can be perpetuated or amplified by persistent recurrent subsidence leading to heat waves, clear skies, and soil moisture deficits (Charney 1975; Namias 1983; Oglesby and Erickson 1989).

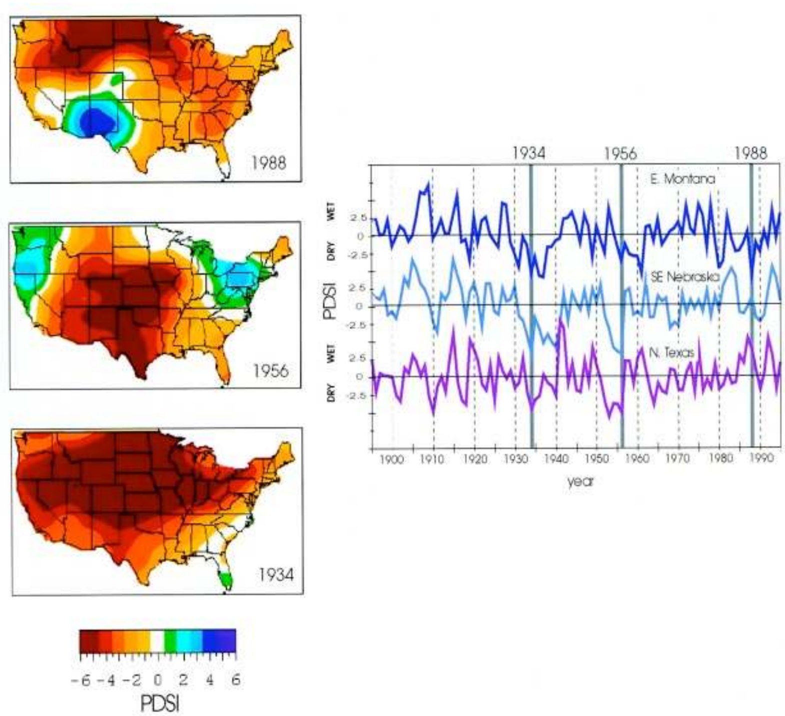

FIG. 12. Spatial distribution of observed PDSI values for three severe twentieth-century drought years (1988, 1956, 1934) (left) and time series of observed PDSI for three grid points in the Great Plains, 1900–94 (right). Gray vertical bars in the time series mark the drought years mapped. This set of maps shows that although PDSI values are low for all three grid points in 1934, in 1956 drought was more severe in the central and southern Great Plains, whereas in 1988, drought is only reflected in the Montana time series, and on the map, across the northern Great Plains [Karl et al. 1990; Guttman 1991; Cook et al. 1996; see also the NOAA/NESDIS Web site (URL given in text)].

What sets up these drought-producing atmospheric circulation patterns, and what drives the variability that leads to the spatial distribution of drought in the Great Plains? There is strong evidence that the state of the oceans, both Pacific and Atlantic, can lead to drought conditions in the Great Plains, directly or indirectly, by inducing perturbations in patterns of atmospheric circulation and the transport of moisture (Trenberth et al. 1988; Palmer and Branković 1989; Trenberth and Guillemot 1996; Ting and Wang 1997). The position of the ridge of high pressure over the plains has been found to be associated with the strength and position of the surface Bermuda high (also called the Atlantic subtropical high) over the Atlantic, which is also linked to the strength of the LLJ (Helfand and Schubert 1995). When this Bermuda high is in a position farther east and north than normal, moist air flows around the high and into the eastern coast, while the Great Plains remains dry. In its usual position, farther south and west, the moist Gulf of Mexico air moves around the high and into the central United States (Forman et al. 1995). The position of this high is likely related to sea surface temperatures (SSTs) in the Gulf of Mexico (Oglesby 1991).

Conditions in the Pacific also influence Great Plains drought-associated circulation patterns. A number of studies have linked conditions in both the equatorial Pacific [El Niño–Southern Oscillation (ENSO)-related conditions] and the northern Pacific to spring and summer precipitation variability in the Great Plains (Trenberth et al. 1988; Kiladis and Diaz 1989; Palmer and Branković 1989; Bunkers et al. 1996; Phillips et al. 1996; Ting and Wang 1997). SSTs in the equatorial Pacific appear to have more influence on summer drought conditions in the northern Great Plains, whereas northern Pacific SSTs are more closely linked to conditions in the central and southern Great Plains (Ting and Wang 1997). The two modes of SST (i.e., the patterns of covariance between SST and precipitation in the two regions) operate independently for the most part and can compound or cancel out one another's impacts on precipitation in the Great Plains. For instance, in 1988, SSTs from both areas contributed to drought conditions, whereas in 1987, modes were in opposition, and precipitation was near normal throughout the Great Plains. The interplay between conditions in the tropical and northern Pacific have been linked to decadal-scale PDSI variability in areas of the United States that include the Great Plains (Cole and Cook 1998). Although Pacific SSTs are not directly linked to the transport of Pacific moisture into the Great Plains, they apparently cause changes in circulation patterns, which, in turn, influence the transport of Gulf of Mexico moisture into the region and the position of the jet stream (Ting and Wang 1997). The position of the jet is associated with the locations of surface fronts and cyclogenesis (Barry and Chorley 1987).

Persistence of drought-producing conditions is a key feature of drought (Namias 1983). The causes of persistent drought-producing conditions on the timescales of months to a season are fairly well understood, but the causes of droughts with durations of years (e.g., 1930s) to decades or centuries (i.e., paleodroughts) are not well understood. Twentieth-century droughts have occurred on subdecadal timescales and persistence related to these droughts has been attributed to anomalous circulation patterns supported by low soil moisture, strong surface heating, and large-scale subsidence (Namias 1983). These synoptic-scale to mesoscale patterns are also maintained by variability in modes of seasonally related large-scale atmospheric circulation (Diaz 1986; Barnston and Livezey 1987; Diaz 1991).

A principal difference between major droughts of the twentieth century and major droughts of the more distant past is the duration, which is on the order of seasons to years compared to decades to centuries. What caused persistence of drought conditions on these timescales? A number of mechanisms may be influencing persistence on decadal timescales. One possible cause of long-term persistence may be related to persistent anomalous boundary conditions influenced by low-frequency variations in the thermal characteristics of oceans (Namias 1983). There is evidence that variations in large-scale patterns of atmospheric circulation and atmosphere–ocean interactions that impact regional precipitation occur on the order of decades to centuries. A recent example of decadal-scale variation is the change in conditions in the North Pacific atmosphere and ocean beginning in the mid-1970s, which impacted climate conditions across the United States (Miller et al. 1994; Trenberth and Hurrell 1994). In the Atlantic Ocean, decadal-scale variations in the Northern Atlantic oscillation (NAO) have been detected and linked to climatic conditions in Europe and the Mediterranean (Hurrell 1995). Other less well investigated twentieth-century decadal shifts in atmospheric circulation patterns have been characterized by changes in zonal versus meridional circulation over North America (Dzerdzeevskii 1969; Granger 1984). Another possibly important source of decadal-scale precipitation variability is the ~20 yr solar–lunar cycle. This cycle has been investigated for years by many researchers (e.g., Mitchell et al. 1979; Stockton et al. 1983; Currie 1981, 1984a,b; Cook et al. 1997) and is a feature that is increasingly being discussed as a control on drought occurrence in the western United States, although the physical link between these cycles, atmospheric circulation, and solar–lunar variability has not yet been determined. A number of proxy records reflect oscillations at the wavelength of these cycles, including tree-ring width chronologies for the western United States (Cook et al. 1997), salinity in North Dakota lake sediments (Laird et al. 1998), and varve thickness records in western Minnesota (Anderson 1992).

At longer timescales, low-frequency variability in ocean SSTs and ocean–atmosphere interactions is a likely source of persistent Great Plains drought conditions in the past. Research has shown changes in the conditions in oceans, such as occurred recently in the North Pacific Ocean, are manifested in long-term climate and proxy records that suggest low-frequency variations have occurred in both Pacific and Atlantic Oceans (e.g., Michaelsen 1989; Duplessey et al. 1992; Rasmussen et al. 1995; Jennings and Weiner 1996; Keigwin 1996). Tree-ring chronologies from the southwestern United States that are sensitive to variations in ENSO reveal a tendency toward low-frequency variations in ENSO events on century scales (Michaelsen 1989). It is known that ENSO events are linked to precipitation in the Great Plains on an event basis (Trenberth et al. 1988; Kiladis and Diaz 1989; Palmer and Branković 1989; Bunkers et al. 1996; Phillips et al. 1996; Ting and Wang 1997), and Rasmussen et al. (1995) suggest that variations in ENSO intensity at the century timescale may broadly correspond to a modulation of interdecadal variations in drought in the Great Plains, with more severe drought epochs (i.e., 1930s–1950s) coinciding with intervals of low ENSO variability. At present, it is not known whether decadal- to century-scale ENSO variability is due to internal variability or external mechanisms, or a combination of both. Currently, there are no good long centuries-long records of North Pacific variability.

Variations in the base state of the Atlantic Ocean may be an important influence on Great Plains precipitation if these variations change the position of the Bermuda high/Atlantic gyre or affect in other ways the ability of Gulf of Mexico moisture to penetrate into the interior United States. For example, Forman et al. (1995) suggested that dune reactivation about 1000 years before present was due to a small easterly shift of the Bermuda high from its usual position in combination with a slight eastward shift of a western-central U.S. ridge aloft, a set of conditions that leads to drought today. There are several sources of proxy data in the North Atlantic Ocean that suggest low-frequency changes in the conditions of this ocean have occurred. A 1300-yr-long record of changes in the East Greenland Current from sediment cores on the coast of eastern Greenland shows a cold interval beginning around A.D. 1270 and peaking around 1370 (Jennings and Weiner 1996), which roughly coincides with the western U.S. drought of the late thirteenth century. However, another cold period in this North Atlantic record spanned the mid–sixteenth century to the early twentieth century, a period not notable for drought in the Great Plains (in fact, the early part of this period was characterized by a lack of drought). In the Sargasso Sea, century-scale variations in SSTs are reflected in $\delta^{18}O$ changes in planktonic foraminifera from marine sediments (Keigwin 1996). Temperatures yielded from this record indicate oscillations from a minimum in A.D. 250–450, to a maximum in A.D. 950–1050, to another minimum in A.D. 1500–1650. All three of these periods correspond to periods of Great Plains drought. Although periods of major Great Plains drought appear to correspond to extremes in the SST record of either sign, perhaps an Atlantic–drought link is related to periods of anomalous conditions or periods of significant change in SST. It is also likely that the effects of anomalous conditions in the Atlantic on Great Plains drought may interact with the impacts of conditions in the Pacific in ways that may enhance or diminish drought conditions.

5. Droughts of the future

A review of the available paleoclimatic data indicates that twentieth-century droughts do not represent the full range of potential drought variability given a climate like that of today. In assessing the possible magnitude of future drought, it is necessary to consider this full range of drought. It is possible that the conditions that lead to severe droughts, such as those of the late sixteenth century, could recur in the future, leading to a natural disaster of a dimension unprecedented in the twentieth century. Two factors may further compound the susceptibility of the Great Plains to drought in the future: 1) increased vulnerability due to human land use practices, and 2) enhanced likelihood of drought due to global warming.

As the limits of productive agricultural lands have been reached, more marginally arable lands have been put into agricultural production in times of favorable climatic conditions and through the use of irrigation. This practice has resulted in an increasing vulnerability to drought in many areas of the Great Plains (Lockeretz 1978; Barr 1981; Hecht 1983). Although the total acreage of irrigated land is not great, irrigation has been an important factor in the increase in cultivated acreage. The High Plains (Ogallala) Aquifer supplies 30% of the ground water used for irrigation in the United States (United States Geological Survey 1997) and is the primary source of water for irrigation in the Great Plains. Since the time of development, pumping of this ground water resource has resulted in water-level drops of more than 15 m in parts of the central and southern plains, with drops that exceed 30 m in several locations, and is already depleted in some areas (Glantz 1989; White and Kromm 1987; United States Geological Survey 1997).

The impacts of drought in these marginal areas have been tempered through social support, but these mitigation measures have been costly. Federal aid costs (disaster assistance, crop insurance, and emergency feed assistance) for the 1988 drought amounted to $7 billion with additional aid supplied by individual states (Riebsame et al. 1991). Total costs associated with this most recent severe drought amounted to over $39 billion (Riebsame et al. 1991). The duration of this drought was about 3 years and the percent of the contiguous United States in severe or extreme drought (Palmer Drought Hydrologic Index ≤ -3.0) peaked at 37% in 1988 (Riebsame et al. 1991). In contrast, the 1930s drought lasted about 7 years, and at its peak almost 70% of the contiguous United States experienced severe or extreme drought (Riebsame et al. 1991). It is difficult to calculate and compare the costs and losses associated with drought, but the costs of mitigating impacts of a 1930s-magnitude drought today would surely be considerable.

General circulation models (GCMs) have been used to estimate the climate change that will accompany increases in tropospheric greenhouse gases leading to a doubling of atmospheric CO_2, calculated to occur in the mid– to late twenty-first century. Most state-of-the-art simulations suggest drier summers will prevail in the central United States under a $2 \times CO_2$ climate scenario (Manabe and Wetherald 1987; Rind et al. 1990; Wetherald and Manabe 1995; Gregory et al. 1997). Model simulations show an earlier drying of soils in spring due to the coincidence of less winter precipitation in the form of snow and warmer temperatures, conditions leading to greater evapotranspiration relative to precipitation in late spring and summer (United States Global Chance Research Program 1995; Gregory et al. 1997). Dry conditions may be further enhanced by a decrease in summer precipitation and relative humidity (Wetherald and Manabe 1995; Gregory et al. 1997). In addition, some GCM studies have suggested an increase in the occurrence of extreme events with global warming (Overpeck et al. 1990; Rind et al. 1990), and although recent modeling results report modest decreases in mean values of summer precipitation and soil moisture in the central United States, a marked increase in the frequency and duration of extreme droughts under $2 \times CO_2$ conditions is also reported (Gregory et al. 1997).

Paleoclimatic data strongly support evidence for Great Plains droughts of a magnitude greater than those of the twentieth century, while current land use practices and GCM predictions point to an increased vulnerability to Great Plains droughts in the next century. Given the likelihood that we are not able to predict the exact extent and duration of the next major drought, it would be wise to adopt a probabilistic approach to drought forecasting and planning that incorporates the range of variability suggested by the proxy data. The paleoclimatic data suggest a 1930s-magnitude Dust Bowl drought occurred once or twice a century over the past 300–400 years, and a decadal-length drought once every 500 years. In addition, paleoclimatic data suggest a drought regime change about 800 years ago, which was likely due to some change in the base state of the climate. An increase in global temperatures is one mechanism that could possibly induce such a base-state change in climate and thus confront society with some costly surprises in the form of multidecadal drought. The prospect of great drought in the future highlights the need to place higher priority on narrowing the uncertainty about future drought by improving our understanding of the causes of drought and our ability to predict great droughts in the future.

Assessments of future drought variability must tap paleoclimatic data, in combination with climate models, to understand the full range of natural interannual to interdecadal drought variability, and to estimate the human-induced climate changes that might occur, superimposed on natural variability. Our current understanding of drought and drought prediction is based on twentieth-century climate variability. This review of the paleoclimatic data for Great Plains drought over the past 2000 years provides a number of lines of evidence that support our conclusion that twentieth-century variability is just a subset of the total climatic variability that can be expected to occur under naturally occurring climatic conditions. We need to gain an understanding of the processes behind the bigger, longer droughts of the last 2000 years. Equally important, we have to make sure our predictive models are capable of simulating these processes and the full range of drought variability. This will require additional paleoclimatic data to map out the exact time–space character of past droughts and associated Pacific and Atlantic influences, and to test the ability of models to simulate the full range of potential drought.

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