

Holocene paleoclimate records from a large California estuarine system and its watershed region: linking watershed climate and bay conditions

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Abstract

The San Francisco Bay-Delta system includes a watershed that covers a large area of California and provides water to two-thirds of the State's population. Climate over the estuary and its watershed in the dry summer months is controlled by the subtropical high which dominates and deflects storms from California. The subtropical high weakens and migrates south as the Aleutian Low strengthens, bringing wet winter storms to the region. Paleoclimatic records from the Bay and its greater watershed, spanning the Holocene, are reviewed here in order to better understand natural variations of precipitation and runoff and the linkages between those variations and the salinity and ecosystems of the estuary. To better understand regional-scale climate patterns, paleoclimate records from coastal California and the Great Basin are also considered. Large fluctuations in climate have occurred during the period of interest, and there is generally good agreement between the paleoclimate records from different regions. Early Holocene climate throughout California was marked by rising temperatures and reduced moisture as seen in fire records from the watershed. This warmth and aridity peaked about 5000–7000 years ago and was followed by a cooling trend, with variable moisture conditions. The Estuary formed relatively rapidly in response to a high rate of sea level rise that dominated the Holocene until about 6000 years ago, and the subsequent reduced rate of inundation allowed vast tidal marshes to form along the edges of the estuary, which have since been recording changes in environmental conditions.

The impacts of changing regional climate patterns are experienced in the San Francisco Bay-Delta system, as altered fresh water flows result in altered estuary salinity. For example, approximately 3800 cal yr B.P., records from throughout the state indicate a cool, moist period, and Bay salinity was reduced; this period was followed by a general drying trend throughout California over the last two millennia, punctuated by decades to centuries-long droughts and brief, extremely wet events. In particular, during the period ca. 1000–800 cal yr B.P. (A.D. 950–1150) conditions seem to have been unusually dry in many parts of the watershed, reducing the fresh water flows to the estuary, and shifting tidal marsh plant assemblages toward less diverse, but more salt-tolerant plants. In contrast, the Little Ice Age (ca. 550–200 cal yr B.P.) brought unusually cool and wet conditions to much of the watershed, and lowered salinity in the Bay. Many reconstructions suggest that notably stable conditions have prevailed over the instrumental period, i.e., after ca. A.D. 1850, even including the severe, short-term anomalies experienced during this period. Interdecadal variability is common in many of the records, with timescales of ca. 55, 70, 90, 100, 150, and 200 years.

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1. Introduction

For most of the Holocene (the last 10,000 years), the San Francisco Bay-Delta region has provided a relatively comfortable home for human populations, with reliable

resources and a mild climate (Fagan, 2003). Approximately 8.9 million people (US Census, 2000) live and work in the 12 counties that surround the San Francisco Bay-Delta today, making this region an important population and economic center in the state of California. The greater watershed of the estuary supplies freshwater to 20 million people. During the past two centuries, the San Francisco Bay estuary and its watershed have undergone major

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human-caused changes, including damming and diversion of major rivers, filling and diking of surrounding wetlands, and extensive land-use changes (Nichols et al., 1986; SFEP, 1992). As a result, it has been difficult to describe the natural state of the estuary, including the character of natural variations of river inflows to the estuary, natural salinity, and natural ecosystems. Improvements in our understanding of the long-term (decades to millennia) natural variations of the estuary can provide an important context for interpreting human-caused changes in the estuary, for setting restoration targets for habitats, for estimating the limits of the various resource systems that draw from the estuary and watershed, and for evaluating the risks from global climate changes.

1.1. Setting

The San Francisco Bay estuary consists of a series of bedrock basins with constricting narrows, or straits connecting them. Pacific Ocean water enters the Bay at the first of these constrictions, the Golden Gate, and inland of this lies the Central Bay, followed upriver to the north by San Pablo Bay, Carquinez Strait (a second bedrock constriction), Suisun Bay and the Delta where the Sacramento and San Joaquin rivers meet (Goals Project, 1999; Malamud-Roam et al., in press). The estuary drains much of the northern half of California (over 40% of the state, Peterson et al., 1989). Fresh water inflow largely determines the health of the estuary ecosystems and, in combination with sea level variations, the salinity of Bay waters (Jassby et al., 1995; Peterson et al., 1995, 1996; Dettinger and Cayan, 2003). The interaction of fresh river water and saline ocean water creates an environment with seasonally and inter-annually varying salinity conditions, to which the organisms within the Bay ecosystems have adapted (Josselyn, 1983).

California's Mediterranean climate, with wet winters and dry summers, affects the timing of fresh water flows into the Delta and estuary. The volume of delta flow is greatest in the winter months, as 55% of California's annual precipitation typically falls over the state during the months of January, February and March (Peterson et al., 1995) in the form of rain over lower elevations and snow at higher elevations. The higher elevation snow pack provides natural storage that releases flows into rivers contributing to the Bay during the dry warm-season months. In response to seasonal changes in flows, Bay salinity fluctuates between fresher conditions in the winter and spring, and saltier conditions in the summer and fall. In addition to these seasonal changes, precipitation and river inflow vary inter-annually with changes in regional climate. Variations in precipitation on interannual time-scales have had dramatic consequences for the state: in the winter of 1861 massive flooding inundated the Central Valley, forming "an immense, quiet inland sea a hundred miles long" that did not drain until the late Spring (Kelley, 1989). This flooding initiated the Califor-

nia's history of flood control management, a complex system-wide series of dams, levees, weirs, and bypass channels. Though their impacts appear less immediately dramatic, droughts similarly threaten the state. The 1976–77 resulted in losses of an estimated \$2.6 billion to the state's economy (CDWR, 1978), and Nichols et al. (1990) argue that the combination of extreme precipitation in 1986, followed by drought lasting from 1987 to 1992, may have weakened Bay ecosystems causing them to be more susceptible to invasion by the Asian clam, *Potamo-corbula amurensis*.

The watershed of San Francisco Bay-Delta (Fig. 1) system encompasses the western slopes of the Sierra Nevada Range, the southern slopes of the southern Cascade Mountains, parts of the Klamath and Coast Ranges, and the Central Valley. At 153,000 square kilometers, the San Francisco Bay watershed is almost

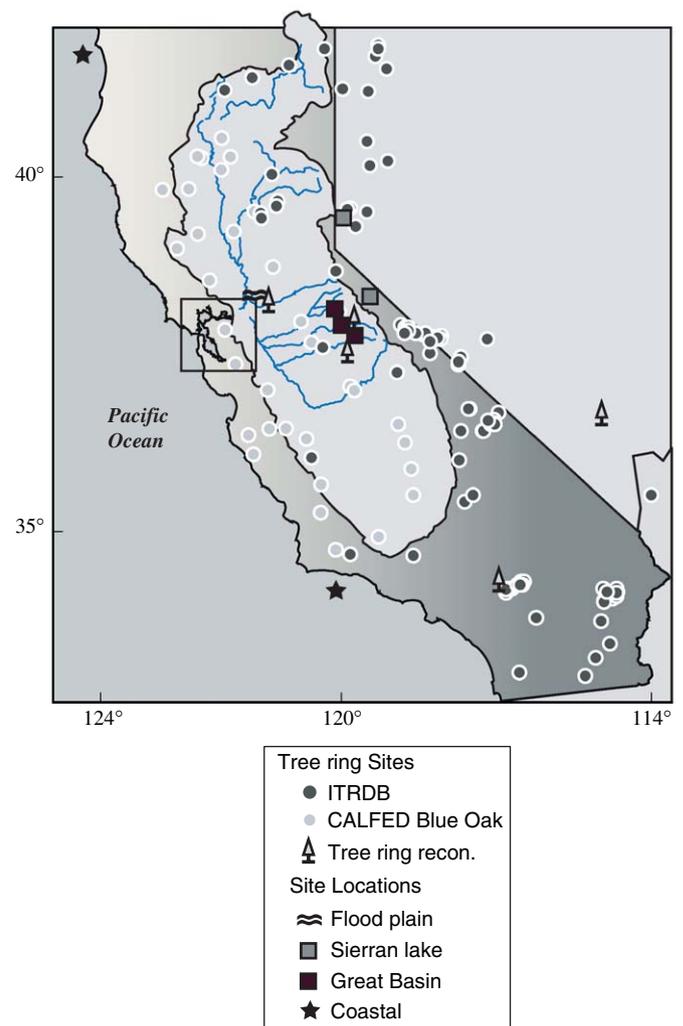


Fig. 1. Map of the study area: San Francisco Bay, California estuary and its watershed. The watershed is comprised of the Sacramento river basin in the north and the San Joaquin river basin in the south. Paleoclimate study sites throughout the San Francisco Bay/Delta system and its watershed. Symbols may reflect multiple sites. Also shown are sites along Coastal California and in the Great Basin.

half the size of California (Atwater et al., 1977) and is vitally linked to estuarine ecosystems, delivering fresh water, nutrients and sediments via rivers and streams and provides critical spawning grounds and migration routes for salmon, trout, and other anadromous fish. Climate, hydrology, geology, and topography vary dramatically over the length and width of the watershed and interact to determine the amount and timing of fresh water delivered to the Bay. The Sacramento River basin supplies about 85% of San Francisco Bay's fresh water, fed mostly by rainfall runoff from storms on the western slopes of the medium altitude, relatively warm northern Sierra Nevada, while the San Joaquin River drains the southern portion of the watershed, mostly in the form of spring snow melt off the southern Sierra Nevada, contributing about 15% of the inflow (Peterson et al., 1989).

San Francisco Bay is closely linked through its watershed with global and regional climate processes (Peterson et al., 1989, 1995). Variations of annual precipitation in California are related to changing patterns of winter-spring atmospheric circulation over the North Pacific Ocean and the tropics, including variations of El

Niño/Southern Oscillation (Cayan and Webb, 1992). Central California is similarly the transition zone for the multi-decadal ENSO-like expressions of the Pacific Decadal Oscillation (PDO; Mantua et al., 1997; Benson et al., 2003) and for other shorter and (especially) longer term climatic influences that are less understood. Superimposed on these natural climate fluctuations are recent warming trends in winter-spring temperatures (Cayan et al., 2001) and associated snowmelt-and streamflow-timing trends (Roos, 1991; Dettinger and Cayan, 1995; Mote, 2003; Stewart et al., 2005). These variations in climate over the watershed are the primary controls on natural variations of salinity in San Francisco Bay (Cayan et al., 1999; Dettinger and Cayan, 2003).

The goals of this paper are: (a) to review the variety of paleoclimatic resources for the San Francisco Bay and watershed in order to identify major climate variations in the pre-industrial past, and (b) to compare the records from the larger watershed region with the Bay records in order to determine the linkages between climate experienced over the larger watershed region and conditions in the San Francisco Bay. The resulting depiction of the past

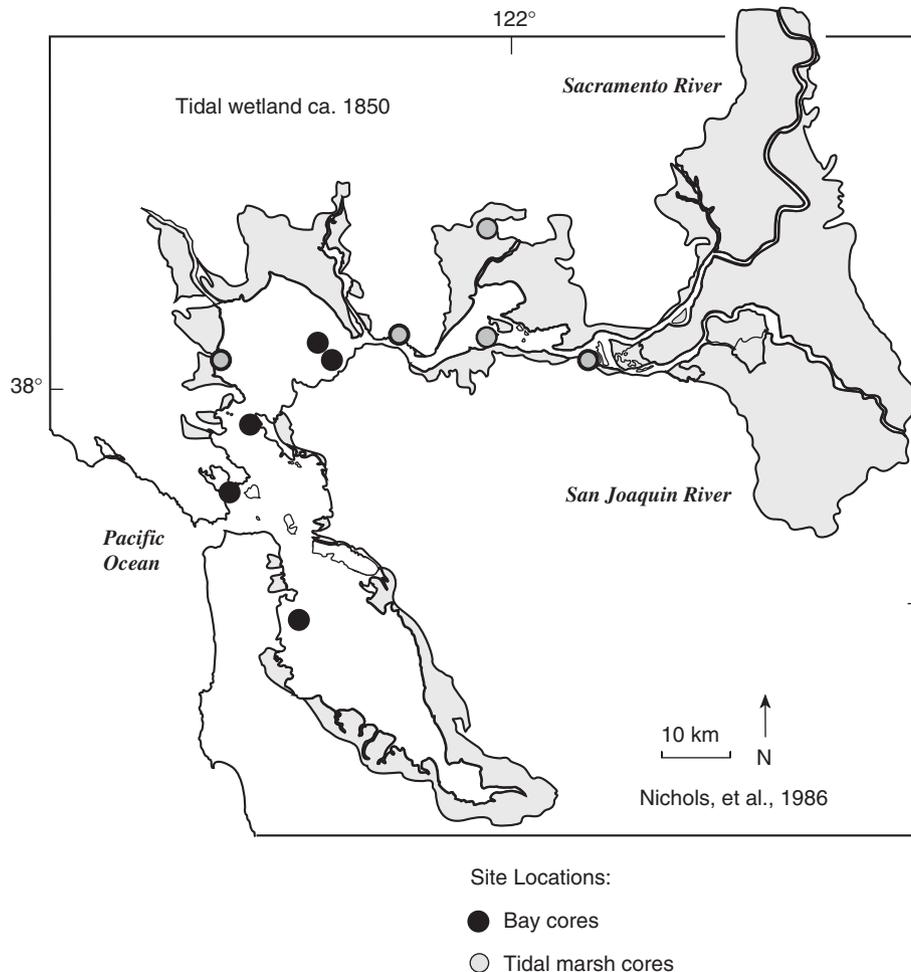


Fig. 2. Map of San Francisco Bay showing the Golden Gate where seawater enters, the Delta where San Joaquin and Sacramento rivers enter and coring locations.

climates of the estuary and watershed provide important contexts for development of scenarios of future climates and climate changes in the estuary, by illustrating the range of natural climatic variations upon which human-induced climate changes are likely to be superimposed.

1.2. Natural history of the San Francisco Bay

The modern Bay estuary began forming sometime between 10,000 and 8000 years ago as rising sea level entered the Golden Gate (Atwater et al., 1979). After an initial period of rapid rise following the last glacial period, global rates of sea level rise slowed ca. 6000 years B.P. (Fairbanks, 1989). Tidal marsh formation and development around the San Francisco Bay has been controlled by relative sea level rise (less than 2 mm/yr, Atwater, 1979), with no evidence of significant tectonic controls (Atwater et al., 1979; Goman and Wells, 2000).

Humans have occupied the surrounding tidal marshlands of the San Francisco Bay estuary for more than 5000 years (Ingram, 1998; Fagan, 2003). The growth in population accompanying the Gold Rush era led to the rapid loss of tidal wetlands around the Bay and Delta, which were diked for agriculture, salt ponds, and housing with a total loss of over 85% of the original (i.e., pre- A.D. 1850) acreage (Nichols and Wright, 1971; Atwater, 1979), some of that offset by new marshes around San Pablo Bay created from hydraulic mining sediments washed in from the Sierra Nevada (Atwater et al., 1979; Jaffe et al., 1998). Efforts in the latter half of the 20th century have been to protect wetland habitat, and mitigate any losses of tidal wetlands as part of the Clean Water Act of 1977 (amended) (Fig. 2).

2. Paleoclimate approaches and archives

In this review, the watershed of the San Francisco Bay is divided into two sub-regions: the Sacramento drainage in the north, and the San Joaquin drainage in the south and we review sites beyond the watershed, such as off-coastal sites and the Great Basin that reflect broad-scale climate events (Fig. 1). Paleoclimate reconstructions of precipitation, temperature and vegetation variations span the past several thousand years (see Table 1 (Friddell et al., 2003; Mensing et al., 1999; Soutar and Crill, 1977)).

2.1. San Francisco Bay watershed

2.1.1. Central Valley lowland floodplain archives

Lowland Central Valley rivers are influenced by climate change through both upstream watershed hydrology and downstream sea level fluctuations. Changes in climate conditions have affected river processes and forced changes in floodplain sedimentation styles and paleo-landforms. These paleo-landforms include main and secondary channels, over-bank deposits, alluvial levees, crevasse splays, abandoned channels and oxbows, and seasonal and

perennial lakes, marshes, and inter-channel wetlands (Gilbert, 1917; Bryan, 1923; Olmsted and Davis, 1961; Brice, 1977; Atwater and Marchand, 1980; Preston, 1981; Sullivan, 1982; Atwater et al., 1986; Florsheim and Mount, 2002, 2003). Sediment assemblages, particle size distributions, and stratigraphy within these landforms provide archives of paleoclimate variability and change in lowland parts of the San Francisco Bay watershed (Fig. 3). For example, analysis and dating of floodplain stratigraphy (of organic-rich and clayey sediment) is a powerful tool for assessing paleoflood magnitude and frequency (Kochel and Baker, 1988; Baker et al., 2002), which can be sensitive to climate change (Knox, 1993).

Holocene sea-level rise created freshwater tidal marsh over floodplain deposits in the Delta (Brown and Pasternack, 2004) and upstream floodplain aggradation created blocked tributary valley lakes (Florsheim and Mount, 2002) near the margins of the Delta. These lakes, formed at the boundary between Pleistocene glacial outwash fans and Holocene floodplain deposits (Atwater and Marchand, 1980), are dendritic, layered with clay, silt, and sand in fluctuating proportions (Strudley, 2001). Basin subsidence and Quaternary glacial outwash deposits created large lakes, like Tulare and Buena Vista lakes, in the southern San Joaquin Basin (Atwater et al., 1986). Prior to anthropogenic disturbances, lake levels and lacustrine sedimentation fluctuated with climate (Preston, 1981).

Geomorphic surfaces formed during past river regimes and floods, such as discontinuous low terraces or higher floodplain surfaces, indicate maximum flood stages (Levish, 2002) and are proxies for climate change and variability in the Central Valley. Archaeological artifacts, pollen, and charcoal found immediately above such surfaces are used to date these flood events to aid paleoclimate reconstructions.

2.1.2. Tree-ring evidence

One of the more established proxies for climatic conditions is the year-to year growth of trees as recorded in annual tree-ring widths (Fig. 4). Tree-ring widths, particularly in trees growing near their environmental limits, reflect regional climate variations. However, annual growth rings respond to complex combinations of climate factors (including interactions between variables such as precipitation, temperature, wind, evaporation), biological factors, and edaphic factors (e.g., soil porosity, pH, nutrient supply) (Hughes, 2002). Before any climatic interpretations can be drawn from the tree-ring widths, the non-climatic influences (e.g., tree specific effects of tree age and size) must be removed to the extent possible, commonly by using statistical techniques for standardizing the ring widths, such as long-term trends fitted with stiff splines, modified negative exponential functions, or straight lines (Cook and Kairiukstis, 1990). This standardization removes tree-specific effects of tree age and size, and averaging the resulting width indices from multiple

Table 1
Paleoclimate studies considered by this review, organized regionally and including the temporal coverage, proxy approach used, climate variable measured, and references

San Francisco Bay estuary	Time period covered (cal. yrs B.P.)		Proxy	Variable	Archives	References
	100	250 350 500 750 1000* 3000 5000 7000 9000 11000 13000				
Richardson Bay	Medium, continuous with gap		$\delta^{18}O_{Sr}$	Water salinity	2 cores (11 m & 2 m long)	Ingram & DePaolo 1993
San Pablo Bay	Fine to Medium, continuous with gap		$\delta^{18}O_{Sr}$, $\delta^{18}O$ & $\delta^{13}C$	Water salinity	3 cores (11 m, 2.5 m, & 6.7 m)	Ingram & DePaolo 1993, Ingram et al 1996b
Oyster Point	Fine to Medium, continuous with gap		$\delta^{18}O$ & $\delta^{13}C$	Water salinity	1 core (5.78 m)	Ingram et al 1996c
Tidal marshes	Medium, continuous		$\delta^{13}C$, Pollen, macrofossils, & Iron*	Vegetation change, hydroperiod	3 cores (6.2m, 4.9m & 4.0 m)	Goman 1996, Malamud-Roam 2002
China Camp	Medium, continuous with possible gap		$\delta^{13}C$ & Pollen	Vegetation change	2 cores (1.95 m, 3.68 m)	Malamud-Roam 2002
Bericia S.P.	Medium, continuous		Macrofossils, Iron	Vegetation change, hydroperiod	2 cores (7.15 m & 3.1m)	Goman & Wells 2000
Peyton Hill	Medium, continuous		$\delta^{13}C$, & Pollen	Vegetation change	1 core (4.25 m)	May 1999, Malamud-Roam 2002
Roe Island	Medium, continuous, possible gap		$\delta^{13}C$, Pollen, diatoms	Vegetation change, salinity	1 core (3.5 m)	Byrne et al. 2001
Rush Ranch	Medium, continuous		$\delta^{13}C$, macrofossils, Pollen	Vegetation change	3 cores (7.79 m, 10.61m & 3.5m)	Goman & Wells 2000, May 1999, Malamud-Roam 2002
Browns Island	Medium, continuous					
Sacramento Watershed	Episodic		Sediment deposits	Floods		Sullivan, 1982.
Floodplains	Fine, continuous		Tree-rings	Streamflow	17 chronologies (pines & Juniperus)	Earle, 1993, Meko et al. 2001
Sacramento R.	Fine, continuous		Tree rings	Salinity	71 chronologies	Stahle et al. 2001
Ft. Point	Fine, continuous				5 chronologies	
San Joaquin Watershed	Medium, continuous		Pollen, macrofossils, charcoal	Vegetation change	2 cores (7.8 m & 1 m); 9 cores - 9 meadows (var lengths, max 4.3 m)	Smith & Anderson 1992 Anderson & Smith 1994 Brunelle & Anderson 2003 Anderson 1990
Central Sierra Nevada	Medium, continuous			Fire frequency	3 cores(0.46 m; 2.72 m; 2.67 m); 9 cores Barrett L. (2.6 m); 2 cores Toga P (3.2 m); 5 cores Starkweather P (3.1 m)	
Southern Sierra Nevada	Fine, continuous		Tree-rings	Precipitation, temperature	3 chronologies (<i>P. balfouriana</i> , <i>J. occidentalis</i>)	Graumlich 1993
Sierra Nevada	Fine, continuous		Tree-rings	Temperature	1 chronology (<i>P. balfouriana</i>)	Scuderi 1993
Ctrl & So Sierra Nevada (Giant Sequoia Groves)	Fine, continuous		Tree-rings	Precipitation	3 chronologies (<i>S. giganteum</i>)	Hughes & Brown 1992
	Fine, continuous		Tree-rings / fire scars	Fire frequency	5 chronologies (<i>S. giganteum</i>)	Swetnam 1993
Great Basin	Coarse, continuous		Tree line	Precipitation, temperature	Tree stumps - 2 sites; 2 chronologies	LaMarche, 1973, 1974 a,b
White Mountains	Fine, continuous		δD	Temperature	Three trees (<i>P. longaeva</i>)	Feng & Epstein 1994
	Fine, continuous		$\delta^{13}C$ (soilmoisture)	Precipitation	Trees (<i>P. longaeva</i>)	Leavitt 1994
Pyramid Lake, Nevada			$\delta^{18}O$	Precipitation	2 cores: 1.535 m long, 1.059 m "box" core	Benson et al., 2002
Mono Lake	Coarse, episodic		Geomorphic features & tree stumps	Precipitation	Shorelines; deltas, stumps	Stine, 1990 a,b
	Medium, continuous		Pollen	Vegetation change	1 core (7.52 m)	Davis 1992
Mojave R.B.	Coarse, episodic		Flood deposits	Floods	2 core (6.4 m); shoreline geomorphology	Enzel et al., 1989, Enzel & Wells, 1997
	Coarse, episodic			Floods	251 sites on 19 rivers	Ely et al, 1993
Southwestern states	Fine, continuous		Tree-rings	Precipitation	1 chronology 6 chronologies	Hughes & Graumlich 1996 Hughes & Funkhouser 1998
Coastal California	Medium, continuous		Diatoms, alkenones, Pollen, CaCO ₃ %, TOC	Sea surface temperature, Vegetation change precipitation	1 core (>8 m)	Barron et al. 2003
No. CA border	Medium, continuous		Pollen	Vegetation change	1 core (6.87 m)	Davis 1992
San Joaquin marsh	Fine, continuous		Tree-rings	Precipitation	3 chronologies & 17 chronologies	Haston & Michaelsen 1994 & 1997
Santa Barbara	Fine, variable		Laminations	Precipitation, floods	3 box cores; 1 (2.87 m) core	Soutar & Crill 1977, Schimmelmann et al 1998, Mensing et al 1999, Field & Baumgartner 2000, Fridell et al., 2003
	Fine, continuous		charcoal, $\delta^{13}C$, $\delta^{18}O$ forams	Fires Sea surface temperature Sea surface temperature	1 core (11 m)	

The temporal spans of the various proxies are indicated in the table along the horizontal axis, with indications of whether the archives produce continuous or episodic records and the temporal resolution of the archives, e.g., fine (annual to decadal scale), medium (centennial to millennial) and coarse (millennial or greater).

trees within a site. The result, a dimensionless time-series of averaged, detrended value known as a “site chronology”, is designed to provide the best possible representation of the shared climatic effects on trees of a given species at a location. Such chronologies typically are based on 30–50 tree cores per study site to ensure statistical reliability (Fritts, 1976, 1990; Hughes, 2002; Hughes and Funkhouser, 1998, 2003).

Once a network of such site chronologies has been established for a region, the climate signals driving the common patterns of variation are identified using climate data from the instrumental period, e.g., the atmospheric circulation patterns that are associated with small rings in Giant Sequoia (*Sequoiadendron giganteum*) throughout its range on the western flanks of the Sierra Nevada. These

associations can then be used to infer the frequency of certain circulation patterns in the pre-industrial times from the older tree-ring widths (Hughes and Brown, 1992). Garfin (1998) extended this approach to a combined analysis of atmospheric conditions limiting growth in both mid- and high-elevation trees in the Sierra Nevada, and demonstrated that these tree rings contain information on circulation patterns in both winter and summer seasons, a finding that remains, as yet, underexploited.

LaMarche (1973, 1974a,b) inferred past climate variations from long-lived bristlecone pines (*Pinus longaeva*) in the White Mountains. He used relict wood dated either dendrochronologically or by radiocarbon to reconstruct temperatures from variations in the upper altitudinal limits of this species and moisture conditions from variations in

recent years, and a new analysis for this extended period is under way.

Linear regression techniques can be used on tree-ring chronologies to reconstruct changes in other physical parameters that are themselves related to climate. For example, some studies have used regional collections of tree-ring chronologies to reconstruct stream flow in major rivers (e.g., Earle, 1993; Meko et al., 2001). By calibrating tree-ring chronologies with stream flow measured at gage stations, reconstructions of paleo-stream flow have been reproduced. Both stream flow and tree-ring width in moisture-stressed trees are correlated to precipitation. Similarly, Stahle et al. (2001) explored calibrating tree-ring data from moisture-sensitive Blue Oaks (*Quercus douglasii*) in the Central Valley of California to produce a record of paleosalinity at Fort Point, San Francisco near the Golden Gate (Fig. 5). The study model assumed that growth for these trees integrates precipitation and temperature conditions over the winter/spring season in the Central Valley in

ways that also would be reflected in the estuarine salinity. Tree-ring width indices were shown to be positively correlated with precipitation over the Central Valley and the Sierra Nevada, and hence with river flow from the Sacramento and San Joaquin basins. Conversely, the tree-ring widths are negatively correlated with salinity.

2.1.3. Paleontological evidence

The species assemblages of pollen and macrofossils in sediments from lakes, ponds, marshes and meadows provide records of paleoclimatic and depositional environments in chrono-stratigraphic order. The resolution of such records is coarser than tree-ring records, in part because sedimentation rates for lake and meadow deposits tend to be low. As a result, a sediment sample collected for paleontological analysis may span several decades of deposition, depending on the sedimentation rate and depth of bioturbation. The sedimentary records are typically dated using a combination of radiocarbon dating (^{210}Pb is useful for more recent sediments), and by identifying known sedimentary features (e.g., previously dated volcanic ash layers or first appearances of pollen from known alien species). Because pollen and macrofossil analyses are primarily used to develop reconstructions of relatively gradual vegetation change in response to changing environmental conditions, higher resolution sampling might not add much information to the resulting reconstructions.

Palynology (primarily pollen analysis, but often including macrofossil analysis) is a well-established methodology for paleo-environmental reconstructions originating in the early, 1900s with L. von Post's publication "On forest tree pollen in south Swedish peat bog deposits" published in, 1916 (Traverse and Sullivan, 1983). This technique has the advantage of using direct evidence of past vegetation, i.e., remains that are preserved in a depositional environment, such as a marsh or lake. Palynologists frequently conduct studies of modern pollen rain for their research sites to calibrate the results of their core data, using weighted averaging and weighted averaging partial least squares regression (Bradley, 1999; Birks, 2005). A comprehensive

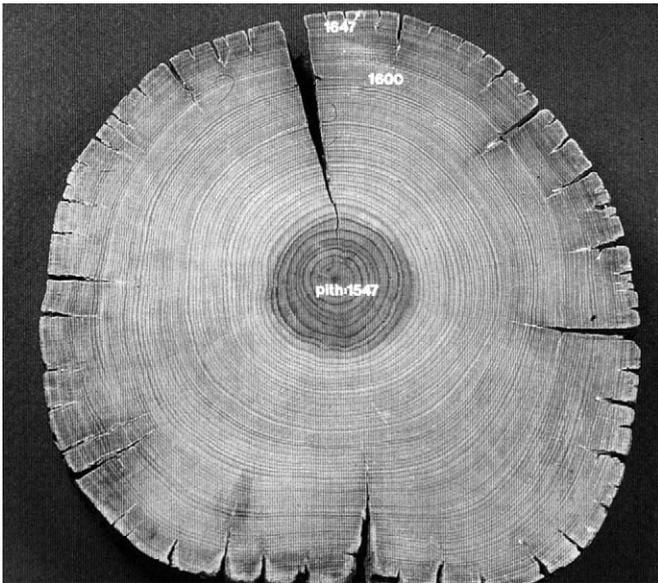


Fig. 4. Image of tree-ring slab showing the annual layering of tree growth rings. Photo: R.R. Parmenter.

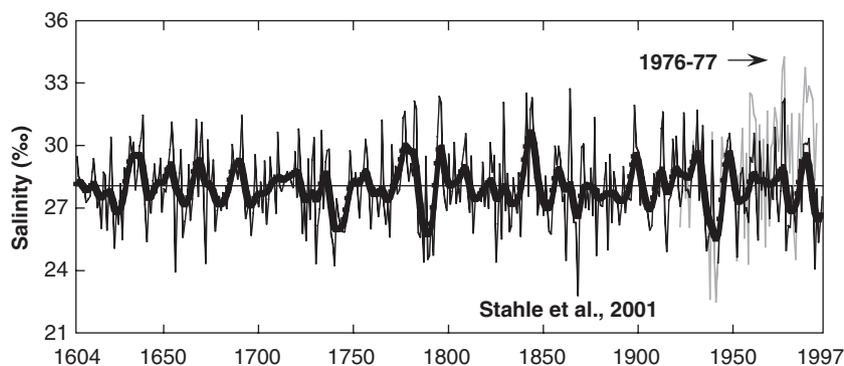


Fig. 5. A 400 year record of January–July salinity (Y-axis, in per mil) at Fort Point, San Francisco derived from Ancient Blue Oak tree ring records. The study included a period of observed salinity records for model calibration (in red). (Stahle et al., 2001).

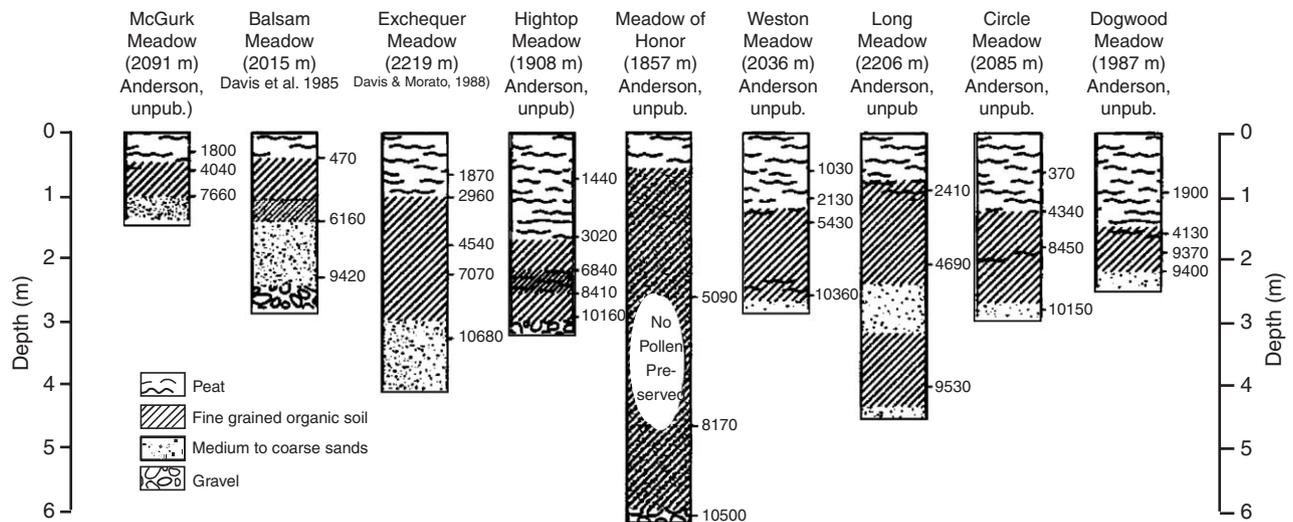


Fig. 6. Generalized stratigraphies from nine montane meadows, oriented from north (left) to south. Radiocarbon dates (in yr B.P.) are shown to right of each profile. (Anderson and Smith, 1994).

review of the application and theoretical underpinnings of this methodology are well discussed in Jackson (1994).

Several pollen and macrofossil studies have produced Holocene records of vegetation change and fire frequency in the mid and high elevations of the central and southern Sierra Nevada. These include high-elevation (Anderson, 1990; Edlund and Byrne, 1991; Edlund, 1996; Brunelle and Anderson, 2003) and mid-elevation lakes or meadows (Smith and Anderson, 1992; Anderson and Smith, 1994). A series of dated core stratigraphies from montane meadows from the Central Sierra Nevada shows changes in vegetation spanning the past 10,000 years (Fig. 6, Anderson and Smith, 1994).

2.2. San Francisco Bay estuary

Fine particles carried in by rivers, ocean sediments washed in with the tides, and the remains of plants and animals living in the estuary and surrounding marshes have been accumulating in the estuary sediments for millennia, providing the primary avenue for reconstructing climate variability over the Holocene, superimposed upon sea level rise. Sediment cores collected from the Bay (including Central Bay, South bay and San Pablo Bay, see Fig. 2) reveal that sedimentation rates throughout the Bay have been variable during this period and subject to varying degrees of bioturbation (Ingram and DePaolo, 1993; Ingram et al., 1996b, c). The Bay provides a depositional environment, with records extending to the last interglacial (Schweikhardt et al., 2002), but with hiatuses, possibly resulting from large floods and complications from old carbon, (Ingram et al., 1996b, c). Sediment cores from tidal marshes reflect changes occurring within the marshlands fringing the Bay estuary, which are less susceptible to erosion events from floods and accrete approximately 1 cm every 10 years on average, though sedimentation rates vary

within the cores (Goman and Wells, 2000; Byrne et al., 2001; Malamud-Roam and Ingram, 2004). Variable sedimentation rates, bioturbation, and compaction (Goman and Wells, 2000) limit the resolution of the tidal marsh core chronologies to decadal-scale at best. contain carbonate material that can be used to derive paleosalinity data (Ingram and Sloan, 1992; Ingram and DePaolo, 1993; Ingram et al., 1996a–c). The tidal marsh cores span the latter half of the Holocene; they generally provide reliable chronologies without hiatuses, and they provide information on how local ecosystems responded to variable salinity.

2.2.1. Stable isotopes

Sediment cores collected from the San Francisco Bay have provided records of past water salinities, derived from strontium isotope ratios and stable oxygen isotopes measured in fossil carbonate shells. Organisms in the Bay precipitate their calcium carbonate shells, recording the isotopic and elemental composition of ambient waters within the shells (Ingram et al., 1996a). Salinity does not affect these compositions, but can be inferred from calculations (mixing models) of the relative amounts of fresh and ocean water to the Bay. The two end-member sources of water for the Bay, saline ocean water and fresh river water, have distinct strontium, oxygen and carbon isotopic compositions, and strontium and oxygen isotopes have been shown to mix conservatively in the estuary. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of average river water entering the Bay is 0.7065, while that of ocean is 0.70918 (Ingram and Sloan, 1992; Ingram and DePaolo, 1993; Ingram et al., 1996a). The oxygen isotopic ($\delta^{18}\text{O}$) compositions of average river water entering the Bay is -11.6‰ , and of ocean water is 0‰ (Ingram et al., 1996a). Studies of the $\delta^{18}\text{O}$ of modern mollusks (*Mytilus edulis*) have demonstrated that total fresh water inflow is the primary control on the $\delta^{18}\text{O}$ of the

carbonate shell with temperature being only a minor effect (Ingram et al., 1996a). The relationships between Sr and O isotopes in the Bay estuary and salinity of Bay water were established with measurements of modern waters of known

salinity (Fig. 7). Due to differences in habitat, food source, and metabolism, it is important to select fossil shells from the same species in sediment cores (Ingram and Sloan, 1992; Ingram and DePaolo, 1993; Ingram et al., 1996b, c).

Carbon isotopic compositions have also been measured of the organic detritus found in sediment cores collected from adjacent tidal marshes (Fig. 7C, Malamud-Roam and Ingram, 2001, 2004; Byrne et al., 2001). The carbon isotopic composition of terrestrial organic matter is determined largely by the plants' photosynthetic pathway: C₃ ($\delta^{13}\text{C}$ about -29‰), C₄ ($\delta^{13}\text{C}$ about -12‰), or CAM (intermediate $\delta^{13}\text{C}$ values, none found in the tidal marshes of the San Francisco estuary); the biogeochemical basis for these differences has been established and reviewed elsewhere (Smith and Epstein, 1970; O'Leary, 1981; Ember et al., 1987). The C₄ photosynthetic pathway is less energetically efficient than the C₃ pathway, but more water efficient (Ziska et al., 1990). In the San Francisco Bay marshes, this evolutionary adaptation exists in two common salt marsh grasses: *Distichlis spicata* (salt grass) and *Spartina foliosa* (cordgrass).

Isotopic composition of marsh sediments, reflecting past plant assemblages, have produced paleosalinity records from Atlantic coastal sites (Delaune, 1986; Chmura and Aharon, 1995) and Bay estuary sites (Malamud-Roam and Ingram, 2004), based on modern studies of sedimentary carbon isotope signatures and salinity of surface sediments (Chmura and Aharon, 1995; Malamud-Roam and Ingram, 2001). Residual differences between expected and measured values of surface sediments in the San Francisco Bay marshes (RMS = 1.9, $n = 59$; Malamud-Roam and Ingram, 2001) had three potential sources: scatter in the isotopic compositions of individual plants of the same species; input of algal material, and effects of diagenesis. The effects of inter-species variability (ca. 5.5‰ maximum difference) seems to reflect differences between populations from different locations, and not between individuals within the same location (Malamud-Roam and Ingram, 2001). In the San Francisco estuary marshes, the salt-tolerant *Salicornia virginica* has a similar isotopic value ($\delta^{13}\text{C} = -27.21\text{‰}$; Malamud-Roam and Ingram, 2001) as the common fresh water plants, *Scirpus californicus* (-27.55‰) and *Jaumea carnosa*

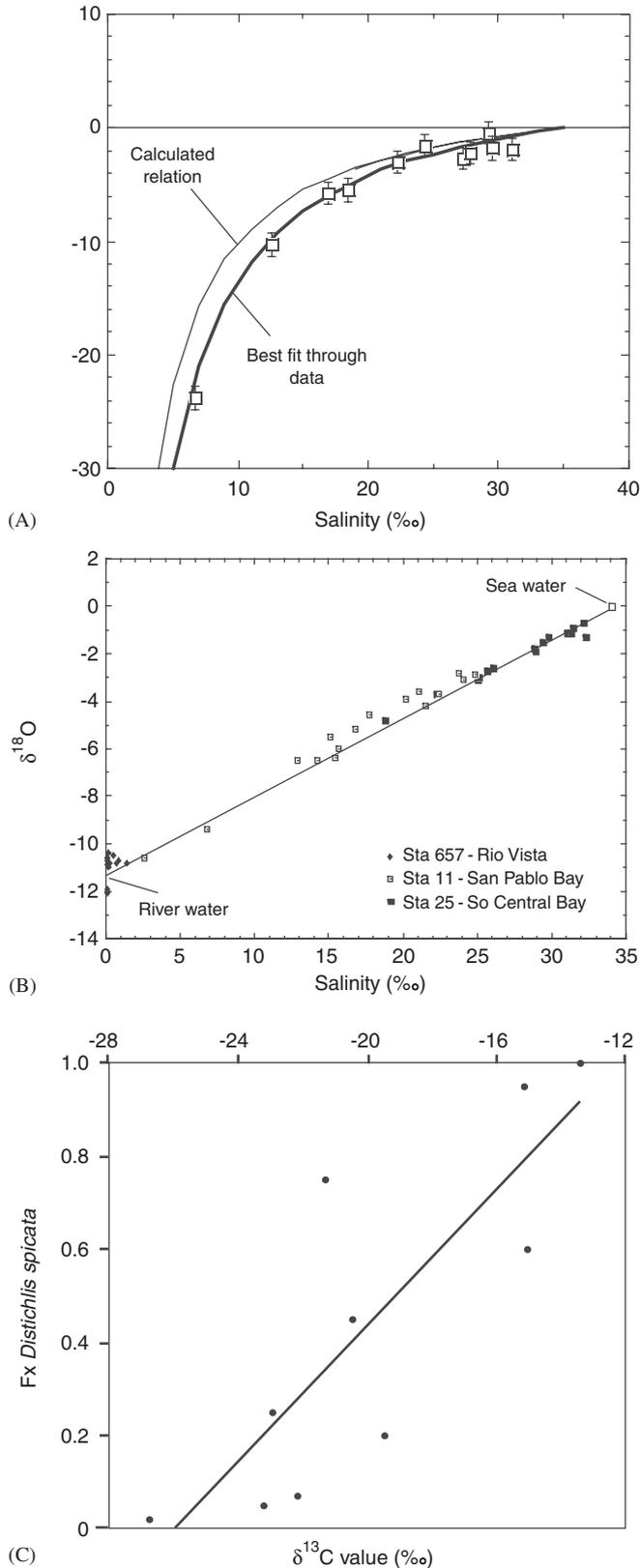


Fig. 7. Relationship between $\Delta^{87}\text{Sr}$ and salinity and $\delta^{18}\text{O}$ and salinity for estuarine waters. (A) Calculated relation uses seawater $\Delta^{87}\text{Sr}$ value of 0, and river water value of -262 . The $\Delta^{87}\text{Sr} = \left\{ \left(\frac{^{87}\text{Sr}/^{86}\text{Sr}}{^{87}\text{Sr}/^{86}\text{Sr}} \right)_{\text{sample}} - \left(\frac{^{87}\text{Sr}/^{86}\text{Sr}}{^{87}\text{Sr}/^{86}\text{Sr}} \right)_{\text{standard}} \right\} \times 100,000$. As the $^{87}\text{Sr}/^{86}\text{Sr}$ of seawater has remained the same as the modern seawater standard (0.70918; DePaolo and Ingram, 1985) for the past 200,000 years, variations in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of fossils from San Francisco Bay precipitates are believed to reflect changes in fresh water influxes and sea levels, and thus can be used as a proxy for salinity. (B) $\delta^{18}\text{O}$ and salinity of estuarine waters collected over a 3-year period (1991–1993) from south San Francisco Bay and San Pablo Bay show a linear correlation with the relation $\delta^{18}\text{O}_{\text{water}} = 0.34 \times \text{salinity} - 11.6\text{‰}$. (from Ingram et al., 1996a). (C) Relationship between the fraction of *Distichlis spicata* in plant assemblage and the measured $\delta^{13}\text{C}$ value of the surface sediments (Malamud-Roam and Ingram, 2001).

(−27.23‰) (Malamud-Roam and Ingram, 2001), posing another complication for paleosalinity reconstructions. Thus, researchers use this method in conjunction with other proxy records, such as fossil pollen (Byrne et al., 2001; Malamud-Roam and Ingram, 2004). *S. virginica* pollen is easily identified in the marsh sediments and the salt marsh grasses can be distinguished from other non-marsh grasses using the C isotopic signal.

2.2.2. Fossil evidence

Cores spanning the last 7000 years have been recovered from tidal marsh sites (see Fig. 2) and have been analyzed for plant macrofossils (seeds, leaves, rhizomes) and pollen to produce records of vegetation change (May, 1999; Byrne et al., 2001; Goman, 2001; Malamud-Roam and Ingram, 2004) and diatoms (Starratt, 2003, 2004) related to paleosalinity and to marsh elevation (Goman and Wells, 2000). Several authors have used normalized ratios of salt-tolerant to salt-intolerant marsh pollen as an indicator of changing salinity (Davis, 1992; May, 1999; Byrne et al., 2001; Malamud-Roam, 2002; Malamud-Roam and Ingram, 2004). These records show that the vegetation assemblages on the surface of these marshes have fluctuated between dominance by a few, highly salt tolerant species, and a more diverse mix of brackish and fresh plant species.

In general, based on modern plant surveys conducted in the San Francisco Bay marshes (Atwater and Hedel 1976; Malamud-Roam and Ingram, 2001), these changes in plant species compositions have been interpreted to reflect changes in estuarine salinity, and indirectly, changes in climate conditions (reflected in fresh water inflows). However, climatic inferences based on these studies have caveats, particularly as other factors can influence the plant assemblages (Byrne et al., 2001), including marsh maturation: as the marsh matures, the plants themselves elevate the surface of the marsh by trapping inorganic sediments and producing organic detritus. In a tidal environment, local changes in elevation relative to sea level have a significant impact on species compositions (Adam, 1990). Sea level has continued to rise steadily throughout the Holocene (Atwater, 1979), resulting in a long-term trend increasing the volume of the Bay, as well as its salinity (Ingram and DePaolo, 1993). Localized changes in marsh elevation may also have occurred as a result of past vertical movement along one of the many faults that cross the San Francisco Bay, though marsh core sediments do not contain evidence of substantial seismic events (Goman and Wells, 2000; Byrne et al., 2001). Plant interactions, including succession and competition also impact the plant assemblages (Adam, 1990).

2.3. Radiocarbon dating

Radiocarbon dating is the primary method for establishing chronologies in estuarine, marsh and lake sediments, as well as archaeological deposits. The uncertainties in radio-

carbon ages vary over time, in part because the production of ^{14}C in the atmosphere has not remained constant throughout time (Stuiver and Quay, 1980; Stuiver and Braziunas, 1992). Hence, chronologies based on radiocarbon dating of plant fragments, concentrated pollen, or bulk organic matter are necessarily of a coarser resolution than those based on annual growth (i.e. tree rings) or deposition (i.e., varves). However, because the radiocarbon method extends back to about 40,000 years (Bradley, 1999), it is ideal for the period of interest (the Holocene epoch), and is often the only method available for dating in many sedimentary environments (such as San Francisco Bay).

Cores from San Francisco Bay estuary were dated using carbonate shell material (Ingram and DePaolo, 1993; Ingram et al., 1996a, b), while cores from the estuary tidal marshes were dated using terrestrial plant matter, primarily seeds (May, 1999; Goman and Wells, 2000; Byrne et al., 2001; Malamud-Roam and Ingram, 2004). The northern hemisphere atmospheric calibration curve was used for these samples and the published radiocarbon dates (where available) have been recalibrated for this review using the online version of CALIB version 4.4 (Table 2; Stuiver et al., 1998a, b). The calibrated dates from estuarine carbonate material needed an additional correction to account for the reservoir age difference between ocean carbon and atmospheric carbon (Stuiver and Braziunas, 1993a, b). In the marsh sediments, seeds are preferred as they are resistant to decay and are deposited on the surface, whereas other plant parts (like rhizomes and roots) can extend into the soil to depths of up to 15 cm or more, complicating the interpretation of the ages of the sediments in which they are buried (Wells, 1995). Lake studies from the watershed region of the estuary also relied on plant material for dating, often by radiocarbon dating of bulk organic samples. The uncertainties inherent in radiocarbon dating, particularly when samples have mixed marine origin, have been mediated to a large extent by published calibration models (Stuiver and Braziunas, 1993a, b; Stuiver and Reimer, 1993).

3. Paleoclimates of the watershed and the San Francisco Bay estuary

3.1. The Sacramento River watershed

Paleoclimate records from the Sacramento River present the natural range of hydrological (i.e. stream flow) and climatic variations influencing the basin throughout the late Holocene. Sullivan (1982) analyzed the sediments (sand, silt lenses) and pollen of an oxbow lake in the Sacramento Valley, “Little Packard Lake,” producing an 800-year record of floods in the Sacramento River upstream of its confluence with the Feather River. Lens thicknesses suggest that very large floods, of a magnitude similar to the historic A.D. 1861 flood, occurred as often as ten times in the last 600 years (Table 3). Similarly, the US

Table 2
Recalibrated radiocarbon dates from sites in the San Francisco Bay estuary

Site–depth (cm)	Material	Age ¹⁴ C yr	2-σ Age range B.P.	Reported cal yr B.P.	Med. Prob. date
San Francisco Bay cores					
Richardson Bay ¹ (dates from Ingram and DePaolo, 1993, recalibrated here using CALIB 4.4, with ΔR value of 365 ± 35 years, obtained from Ingram and Southon, 1996)					
RBGC-3 (40)	Mollusk	<0	n/a	35 ± 5	
RBGC-5 (60)	Mollusk	830 ± 60	0–238	125 ± 60	98
RBGC-7 (80)	Mollusk	580 ± 70	n/a	0 ± 70	n/a
RBGC-10 (100)	Mollusk	690 ± 60	n/a	70 ± 60	n/a
RBGC-12 (130)	Mollusk	890 ± 70	0–279	225 ± 105	150
RBGC-14 (140)	Mollusk	850 ± 110	0–322	150 ± 110	139
RBGC-17 (170)	Mollusk	670 ± 110	n/a	20 ± 70	n/a
RBGC-18 (180)	Mollusk	1100 ± 100	133–538	440 ± 110	369
RBGC-20 (195)	Mollusk	1230 ± 70	328–612	570 ± 100	483
RB-2.1A (441)	Mollusk	780 ± 80	n/a	90 ± 80	n/a
RB-2.1b (460)	Mollusk	1070 ± 70	154–495	370 ± 70	353
RB-2.2A (526)	Mollusk	2690 ± 60	1782–2112	2210 ± 60	1940
RB-2.3A (666)	Mollusk	3680 ± 80	2919–3359	3410 ± 80	3151
RB-2.5b (815)	Mollusk	3720 ± 70	2984–3380	3510 ± 70	3205
RB-2.6A (849)	Mollusk	3790 ± 100	3000–3542	3590 ± 100	3284
RB-2.9A (1070)	Mollusk	4390 ± 80	3808–4293	4320 ± 80	4037
San Pablo Bay (dates from Ingram and DePaolo, 1993, recalibrated here using CALIB 4.4, with ΔR value of 365 ± 35 yrs, obtained from Ingram and Southon, 1996).					
SPB-3A-9 (150)	Mollusk	730 ± 70	n/a	130 ± 70	n/a
SPB-3A-11 (190)	Mollusk	740 ± 60	n/a	130 ± 60	n/a
HF-4-cc 270	Mollusk	1810 ± 120	739–1245	1180 ± 150	996
HF-5-2 (334)	Mollusk	2930 ± 60	2040–2361	2580 ± 80	2225
HF-6-1 (395)	Mollusk	2890 ± 90	1928–2393	2470 ± 90	2181
HF-9cc (850)	Mollusk	2820 ± 180	1671–2612	2435 ± 225	2101
HF-11-1 (1003)	Mollusk	3080 ± 80	2213–2696	2760 ± 90	2437
HF-11-2 (1034)	Mollusk	3070 ± 60	2266–2675	2770 ± 100	2417
San Pablo Bay (dates from Ingram et al., 1996c, recalibrated here using CALIB 4.4, with ΔR value of 365 ± 35 yrs)					
SPB (21)	Macoma	830	0–238	260	98
SPB (55)	Macoma	840	0–242	270	105
SPB (105)	Macoma	1260	402–631	610	509
SPB (127)	Mytilus	2320	1349–1676	1670	1504
SPB (207)	Mytilus	2420	1468–1802	1790	1624
SPB(327)	Mytilus	2600	1670–1998	1980	1833
SPB (378)	Mytilus	2720	1814–2139	2130	1974
SPB (486)	Mytilus	2690	1782–2112	2110	1940
SPB (551)	Mytilus	2690	1782–2112	2110	1940
SPB (583)	Mytilus	2650	1719–2056	2080	1893
SPB (676)	Mytilus	2710	1806–2131	2130	1962
Oyster Point (dates from Ingram et al., 1996b, recalibrated here using CALIB 4.4, with ΔR value of 365 ± 35 yrs, obtained from Ingram and Southon, 1996)					
OY (5)	Macoma	750 ± 60	n/a	160	n/a
OY (29)	Macoma	1110 ± 60	274–494	500	383
OYP (57)	Macoma	1410 ± 60	512–715	710	611
OYP (65)	Macoma	730 ± 70	n/a	120	n/a
OYP (269)	Macoma	2270 ± 70	1287–1620	1590	1450
OYP(380)	Macoma	3150 ± 70	2335–2706	2720	2521
OYP (402)	Macoma	3310 ± 70	2470–2879	2840	2715
OYP (523)	Macoma	5070 ± 60	4802–5143	5130	4939
Tidal Marsh Cores					
Browns Island ¹ (dates from May, 1999, recalibrated here using CALIB 4.4)					
BI-59-60	Sc ² seeds	240 ± 80	0–470	0–420	274
BI-88-89	Sc& unk sds	810 ± 80	573–920	660–780	743
BI-128-129	Sc& unk sds	1050 ± 40	917–1056	930–970	962
BI-148-49	Sc& unk sds	1210 ± 50	992–1262	1060–1170	1131
BI-178-179	unk sds	1270 ± 80	988–1310	1070–1280	1186
BI-208-09	Sc& unk sds	1560 ± 50	1333–1543	1360–1520	1452
BI-239-40	Sc& unk sds	1780 ± 80	1527–1877	1570–1810	1700

Table 2 (continued)

Site–depth (cm)	Material	Age ¹⁴ C yr	2-σ Age range B.P.	Reported cal yr B.P.	Med. Prob. date
BI-322-23	Sc seeds	2330 ± 70	2143–2709	2320–2360	2352
Browns Island (dates from Goman and Wells, 2000, recalibrated here using CALIB 4.4)					
BI 92 (12)	Sc seeds	430 ± 60	315–545	428	473
BI 92 (43)	Sc seeds	480 ± 120	326–645	518	517
BI 92 (160)	Sc seeds	1430 ± 120	1191–1510	1319	1335
BI 92 (272)	Leaf	1741 ± 120	1531–1816	1632	1652
BI 92 (339)	Sc seeds	2470 ± 200	2339–2753	2546	2545
BI 92 (523)	Sc seeds	3580 ± 420	3379–4497	3886	3898
BI 92 (547)	Leaf	3630 ± 120	3728–4144	3952	3943
BI 92 (605)	Leaf	3820 ± 120	4000–4412	4218	4218
BI 92 (687)	Sc seeds	4290 ± 120	4629–5040	4850	4856
BI 92 (747)	Sc seeds	4930 ± 120	5491–5886	5664	5666
BI 92 (758)	Bulk	5190 ± 120	5751–6175	5948	5948
BI 92 (779)	Leaf	5530 ± 180	6001–6525	6312	6324
BI 93 (409)	Sc Seeds	2920 ± 120	2878–3315	3077	3066
BI 93 (801)	Woody debris	4970 ± 120	5595–5889	5743	5704
BI93 (928)	Sc seeds	5390 ± 120	5994–6291	6192	6179
BI 93 (928B)	Bulk peat	5340 ± 120	5949–6278	6111	6109
BI 93 (1060)	Leaf	5880 ± 180	6450–6896	6710	6689
Peyton Hill (dates from Goman and Wells, 2000, recalibrated here using CALIB 4.4)					
PH (309)	Sa Sc seeds	2580 ± 120	2362–2840	2736	2632
PH (353)	Sc seeds	2920 ± 120	2878–3315	3078	3066
PH (406)	Sc seeds	3270 ± 120	3376–3635	3482	3503
PH (450)	Sc seeds	3410 ± 120	3478–3828	3637	3659
PH (590)	Sc seeds	4220 ± 120	4550–4867	4743	4730
PH (642)	Sc seeds	4420 ± 120	4859–5285	5041	5021
PH (705)	Sc seeds	4740 ± 120	5321–5591	5455	5480
Rush Ranch (dates from Byrne et al., 2001, recalibrated here using CALIB 4.4)					
RR (115)	Sc seeds	840 ± 120	669–911	674–785	760
RR (176)	Sc seeds	1840 ± 100	1626–1884	1706–1826	1772
RR (260)	Sc seeds	2390 ± 100	2333–2710	2344–2461	2445
RR (300)	Bulk Peat	2400 ± 120	2336–2711	2492–2748	2466
Roe Island ¹ (dates from May, 1999, recalibrated here using CALIB 4.4)					
RI-59-60	Sc& unk sds	190 ± 50	0–308	270	178
RI-99-100	Sc seeds	390 ± 60	312–518	470	434
RI-139-40	Sc seeds	820 ± 140	540–1046	720	768
RI-179-80	Sc seeds	1300 ± 60	1070–1306	1260	1220
RI-199-200	Sc seeds	1370 ± 40	1184–1347	1290	1289
RI-229-30	Sc seeds	1720 ± 80	1420–1821	1610	1633
RI-254-55	Sc& unk sds	1830 ± 60	1574–1916	1730	1761
RI-279-80	Sc& unk seeds ²	1800 ± 60	1568–1867	1710	1726
RI-344-45	woody frags	2110 ± 90	1896–2327	2060	2091
Benicia State Park (BSP825) (dates from Malamud-Roam and Ingram, 2004)					
BSP-50-51	Rhizome ³	130 ± 40	0–277		132
BSP-71-72	Rhizome ³	270 ± 130 ⁵	0–510		301
BSP-78-79	Sc seeds	1175 ± 40	973–1219		1095
BSP-85-86	Sc seeds	1170 ± 40	971–1175		1088
BSP-98-99	Sc seeds	1200 ± 40	991–1258		1120
BSP-150-51	Sc seeds	1290 ± 50	1076–1294		1218
Benicia State Park (BSP3401) (dates from Malamud-Roam and Ingram, 2004)					
BSP-123-26	Sa seeds	1870 ± 50	1634–1923		1802
BSP-240-41	Sa&unk seeds	2985 ± 40	3001–3322		3167
BSP-351-52	Sa seeds	3345 ± 40	3472–3685		3576
Petaluma Marsh (dates from Byrne et al., 2001, recalibrated here using CALIB 4.4)					
PET (42-48)	Bulk Peat	200 ± 120	0–424	0–296	184
PET (52-58)	Bulk Peat	280 ± 120	3–494	0–430	358
PET (60-70)	Bulk Peat	590 ± 140	514–662	529–648	594
PET (80-90)	Bulk Peat	480 ± 100	335–627	501–536	519
PET (90-100)	Bulk Peat	520 ± 100	485–647	511–548	541

Table 2 (continued)

Site–depth (cm)	Material	Age ¹⁴ C yr	2-σ Age range B.P.	Reported cal yr B.P.	Med. Prob. date
PET (105-15)	Bulk Peat	790 ± 120	572–906	664–733	717
PET (120-30)	Bulk Peat	880 ± 100	696–915	724–889	801
PET (148)	Seed	1000 ± 100	788–1050	800–946	914
PET (150-60)	Bulk Peat	1220 ± 100	994–1265	1064–1223	1142
PET (180-90)	Bulk Peat	1740 ± 140	1425–1863	1544–1716	1654
PET (210-20)	Bulk Peat	1900 ± 160	1613–2037	1720–1925	1837
China Camp State Park (dates from Malamud-Roam and Ingram, 2004)					
CC-54-55	seeds	405 ± 40	320–521		468
CC-153-54	seeds	915 ± 40	740–923		840
CC-182-83	Rhizome ⁴	930 ± 40	743–928		849
CC-254-55	Seeds	520 ± 640 ⁶	—		n/a
CC-368-69	Seeds	2540 ± 40	2471–2750		2598
CC-450-51	Seeds	3330 ± 40	3468–3680		3559
CC-470-71	Seeds	3430 ± 60	3480–3834		3685

Notes:

- 1 Seeds are abbreviated as follows: Sc = *Scirpus* (species not known); Sa = *Salicornia virginica*; unk = unknown.
- 2 This date is questionable—possible contamination during sampling.
- 3 *Distichlis spicata* rhizomes, implying that date is younger than surrounding material at that depth.
- 4 The large uncertainty is due to very small sample size (0.04 mgC).

Bureau of Reclamation (USBR) has investigated terrace surfaces delineated on the basis of geomorphic expression and soil development, near Fair Oaks on the Lower American River. Geomorphic, stratigraphic, and geochronological relations have provided preliminary evidence for numerous large paleo-floods. Archaeological artifacts present on and within the terrace are interbedded with sand layers and supply a general chronology of five flood events, including one between 1600 and 1400 cal yr B.P. (A.D. 350–550), one between 1125 and 650 cal yr B.P. (A.D. 825–1300), and three between 650 and 150 cal yr B.P. (A.D. 1300–1800) (USBR, 2002).

Flood plain sedimentary evidence can only describe the upper range of natural stream flows, and does not demonstrate the climate impacts on surrounding ecosystems. Tree-ring studies provide a more complete picture of moisture availability, and, by definition, the effects on ecosystems. Earle (1993) correlated tree-ring chronologies from 15 locations in California and Oregon to stream flow in four rivers in the Sacramento River Basin, and produced a stream flow reconstruction dating back to A.D. 1560 (390 cal yr B.P.). He found that sustained periods of low or high flow have generally been synchronous between the northern and southern parts of the basin, with some exceptions. Both the wettest and driest events Earle found in the tree-ring records occurred during the historical period: the wettest interval was A.D. 1854–1916, followed by the driest from A.D. 1917–1950. These periods correspond closely to a reversal in the conditions of the Pacific Decadal Oscillation from colder (negative PDO) to warmer (positive PDO) that occurred at A.D. 1924 (Mantua et al., 1997). However, a more recent and much longer tree-ring reconstruction of annual Sacramento River

stream flows completed by Meko et al. (2001) extends back to A.D. 869 (Fig. 8). In that reconstruction, prolonged periods (several decades) of high or low flows characterized the period before ca. A.D. 1400 (550 cal yr B.P.), and extended droughts may have been more common before that date.

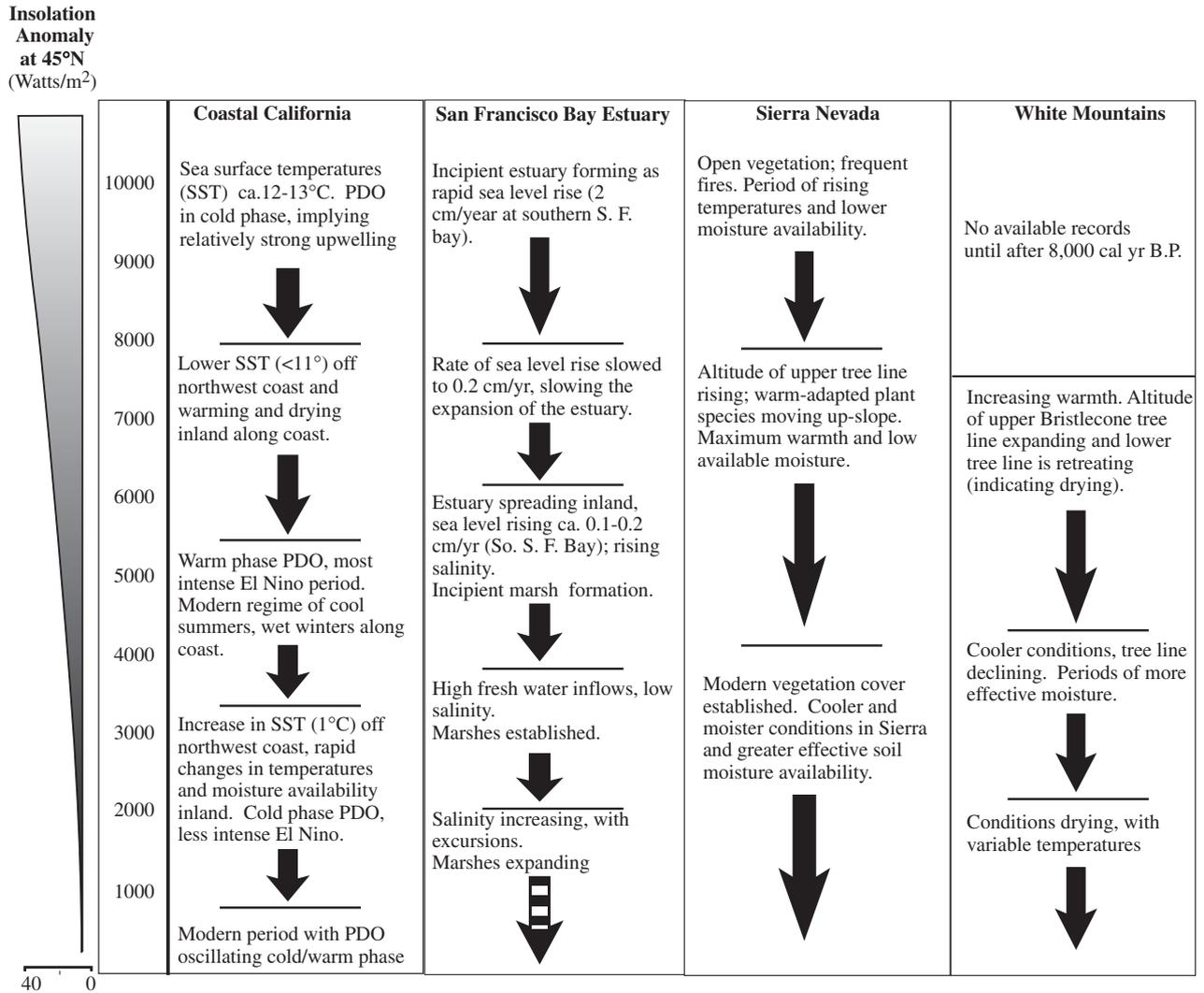
Meko et al. (2001) used the, 1930s drought, a brief, but extreme drought (seen across the coterminous US, Cook et al., 1999), as representative of short (less than ten years) droughts. On only one other occasion has reconstructed six-year mean flows been lower than the reconstructed mean for the period, 1929–1934: a seven-year drought ending in A.D. 984. The most extreme single-year drought in this reconstruction occurred in the year A.D. 1580 (370 cal yr B.P.), when flows almost certainly were less than in A.D. 1977 (Meko et al., 2001). This extreme drought year is seen in many tree-ring chronologies throughout the west, most notably in the Giant Sequoia groves of the western flank of the Sierra Nevada as far north as the American River (Brown et al., 1992; Hughes and Brown, 1992; Hughes et al., 1990, 1996). Throughout the Sierra Nevada, most sequoia had either an extremely thin ring, or the ring was absent for the year A.D.1580.

3.2. The San Joaquin River watershed and Great Basin

Paleoclimatic studies in the San Joaquin River watershed, and surrounding areas, provide long term perspectives on the climate of central California, derived from central and southern Sierra Nevada mountain lakes and meadows that go back to the early Holocene/late Pleistocene transition (e.g., Anderson, 1990; Anderson and Smith, 1994; Edlund, 1996; Brunelle and Anderson,

Table 3

General climatic trends recorded in Holocene paleoclimate records throughout California, organized geographically from the coast to the White Mountains



This is a summary from sources cited in this paper.

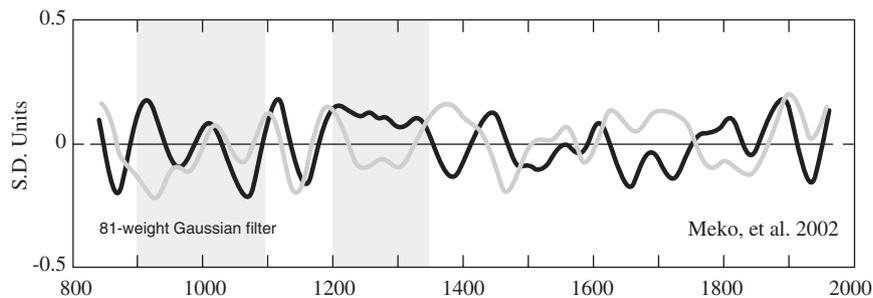


Fig. 8. 1200 year record of river flow for Sacramento River (dark line) and San Joaquin River (light line). Negative deviations indicate low flows. Shaded area reflects period of droughts recorded at Mono Lake (Stine, 1994). Source: Meko et al., 2002.

2003). Episodic events are recorded in the sediments of southwestern playas (e.g., Enzel et al., 1989; Ely et al., 1993), and studies of former lake levels from sites on the east side of the Sierra Nevada (e.g., Benson et al., 2002;

Mensing et al., 2004; Stine, 1990a, b, 1994) demonstrate that climate throughout the Holocene epoch has varied substantially. Providing greater detail, tree-ring studies from this region, particularly from Giant Sequoia groves,

but also from a variety of other tree species, describe both changing precipitation and temperature patterns for the Holocene on annual to decadal time-scales (e.g., Graumlich, 1993; Hughes et al., 1996; Hughes and Graumlich, 1996; Biondi et al., 1999; Hughes and Funkhouser, 2003).

Pollen assemblages from several mountain lakes and meadows in the central Sierra Nevada provide records of vegetation responses to changing climatic conditions from the late Pleistocene through the Holocene (Anderson, 1990; Smith and Anderson, 1992; Anderson and Smith, 1994; Edlund, 1996; Brunelle and Anderson, 2003), which have been supplemented by recent studies of fossilized midge remains in lake sediments (Porinchu et al., 2003). These studies indicate that the early Holocene (until ca. 7000 cal yr B.P.) was generally dry, with open vegetation following the recent retreat of mountain glaciers, and frequent fires (Edlund and Byrne, 1991; Brunelle and Anderson, 2003). Warming continued during the middle Holocene (7000–3500 cal yr B.P.) as solar insolation reached its maximum for the northern hemisphere. In the midlatitudes, soil moisture was greater, perhaps due in part to increased vegetative cover (Anderson and Smith, 1994; Edlund, 1996). These generally warm and moist conditions were seen throughout the American southwest, for example, in paleo-flood histories (Ely et al., 1993) and lake deposits from the Mojave Desert (Enzel et al., 1989). Recent studies have reconstructed changes in the lake level of Pyramid Lake, NV, as the lake has shifted from an open system to a closed basin system in response to changes in moisture regime. Total inorganic carbon, which detects abrupt changes from open to closed basin conditions and oxygen isotope measurements (Benson et al., 2002), and pollen records (Mensing et al., 2004) show a complicated early to mid-Holocene climate history, with lake level fluctuations indicating overall drought conditions punctuated by brief very wet periods from ca. 8000 cal yr B.P. to about 3800 cal yr B.P., when conditions became generally wetter. LaMarche (1973, 1974a), analyzing the altitudinal limits and tree-ring widths of bristlecone pines in the White Mountains, estimated that temperatures were about 1.9 °C warmer during this middle Holocene period. Conditions in the central Sierra Nevada became moist for a period ca. 4000–3500 cal yr B.P., and a cooling trend with higher variability began, which continued through to the modern period (LaMarche, 1974a; Anderson, 1990; Smith and Anderson, 1992; Anderson and Smith, 1994; Brunelle and Anderson, 2003).

Stine (1990a, b, 1994) used former lake level elevations of Mono Lake and submerged tree stumps from sites in the eastern Sierra Nevada as evidence of prolonged, severe droughts occurring twice in the last 1000 years. Both droughts occurred during the period from the 10th century A.D. to the 14th century, referred to as the Medieval Climate Anomaly (Stine, 1994). When the lake levels dropped during these droughts, trees grew at elevations in the basin (and in other river beds in the eastern Sierra) that are completely submerged today. By dating the submerged

tree stumps and counting the tree rings, Stine concluded that these mega-droughts prevailed in the region from ca. 1050–850 cal yr B.P. (A.D. 900–1100) and again from 750 to 600 cal yr B.P. (A.D. 1200–1350). In at least two cases, these eastern Sierran droughts appear to precede anomalously wet periods in the larger region (as evidenced by the submerged trees). A paleoclimatic reconstruction from bristlecone pines found at high elevations in the White Mountains (LaMarche, 1974a) corresponds to the Mono Lake droughts: from ca. 1100–900 cal yr B.P. (A.D. 850–1050), conditions were cool and dry, and from ca. 830–650 cal yr B.P. (A.D. 1120–1300), conditions were warm and dry. As in the Mono Lake record, these dry periods were terminated by periods of increased moisture, e.g., 800–730 cal yr B.P. (A.D. 1150–1220) was warm and wet in the White Mountains, and 650–300 cal yr B.P. (A.D. 1300–1650) was cool and wet. As severe as these droughts were—they also appear in tree-ring records from the region (Graumlich, 1993; Hughes and Funkhouser, 1998)—they were considerably less severe than the droughts detected in the early Holocene by Brunelle and Anderson (2003) based on charcoal remains indicating frequent and intense fires. Fire histories provide records of decadal/centennial variations in temperature and interannual fluctuations in precipitation (Swetnam, 1993; Brunelle and Anderson, 2003).

The late Holocene tree-ring studies from the San Joaquin watershed region and the mountain ranges of the Great Basin indicate that (1) the 20th century experienced relatively stable climate conditions, lacking major droughts when considered on a millennial timescale, though there have been brief and intense droughts such as in, 1977 (Hughes and Brown, 1992; Hughes et al., 1996) and (2) major, sustained droughts were more common before ca. 450 cal yr B.P. (A.D. 1500) than after (Hughes and Funkhouser, 1998). Several climate reconstructions from subalpine tree-ring records of in the San Joaquin watershed and the White Mountains (LaMarche, 1974a; Graumlich, 1993; Graybill and Funkhouser, 1999) have shown that unusually dry conditions prevailed during the Medieval Climate Anomaly, followed by wetter (and cooler) conditions during the 17th to the, 19th centuries, referred to as the Little Ice Age. For example, Graybill and Funkhouser (1999) describe temperature reconstructions for Grants Grove based on tree-ring chronologies that reveal a rapid warming trend from 950 to 800 cal yr B.P. (A.D. 1000–1150), and a cooling trend in the 17th and early, 19th centuries. Fire frequency (and severity), as seen in fire scars from giant Sequoia groves (Swetnam, 1993) also supports this pattern. Swetnam (1993) found more frequent fires occurred during a warm period from 950 to 650 cal yr B.P. (A.D. 1000–A.D. 1300), and fewer during the cooler periods from 1450 to 950 cal yr B.P. (A.D. 500–A.D. 1000) and after 650 cal yr B.P. Finally, Graumlich (1993) analyzed 20-yr and 50-yr precipitation and temperature means in the region for the last 1000 years and found a period of extended summer warmth at Grants

Grove in Sequoia National Park during the 12th century, with temperatures about 0.24 °C warmer than the, 1928–1988 instrumental mean from A.D. 1118–1167 (832–783 cal yr B.P.). The coldest reconstructed 50-year period was from A.D. 1595–1644 (355–306 cal yr B.P.), and was 0.45 °C cooler than the instrumental mean.

3.3. The San Francisco Bay–delta estuary

Unlike the watershed, most of the paleoclimatic information from San Francisco Bay and its immediate surroundings comes from sediment analyses. As noted previously, these paleoclimatic resources provide coarser temporal resolution, but they describe the estuarine variations and responses to climate changes “in the first person” (as the evidence of local ecological change are preserved in situ) rather than through inferences based upon distant climatic and hydrologic forcings seen in paleoclimatic resources of the Sierra Nevada. There is a need for more comprehensive core collection from the Bay and its surrounding tidal marshes. There is significant within and between marsh variability due to site specific factors that can pose challenges to paleoclimatic interpretations. However, considerable paleoclimate research has been underway in the estuary, and is described below.

3.3.1. Bay sediment records

Estuary sediment cores have been collected from four sites in the San Francisco Bay: two sites in San Pablo Bay (Ingram and DePaolo, 1993; Ingram et al., 1996a), one site in Richardson Bay (Ingram and DePaolo, 1993) and one site near Oyster Point in the Central Bay (Ingram et al., 1996b; Fig. 2). Strontium isotopic compositions in carbonate shells of fossil mollusks in the sediment cores were measured to reconstruct paleosalinity in Richardson Bay during the past 4300 yr, and in San Pablo Bay during the past 1100 yr (Ingram and DePaolo, 1993). Oxygen and carbon isotopic compositions were measured in carbonate shells from the clam species, *Macoma nasuta*, and the mussel *Mytilus edulis*, in another sediment core from San Pablo Bay to reconstruct a 2100-yr record of paleosalinity, though a hiatus occurred in that record between ca. 1670 and 750 cal yr B.P. (Ingram et al., 1996c). Carbon and oxygen isotopic compositions were measured in a core from Oyster Point to reconstruct a 5900 years record of paleosalinity, excluding the period from ca. 2700 cal yr B.P. to 3800 cal yr B.P., which lacked fossil material (Ingram et al., 1996b). These studies show salinity in the estuary increasing over the past 2500–4300 as sea level rose. Higher-frequency variations in salinity (and thus fresh water inflow) are superimposed upon the overall trend of increasing salinity.

Variations inferred from the Bay cores reflect timescales of about 90, 200 and 500 years (Ingram and DePaolo, 1993; Ingram et al., 1996a, b). These same timescales are found in the spectra of radiocarbon generation in the atmosphere (Stuiver and Quay, 1980; Damon et al., 1989;

Stuiver and Braziunas, 1992) and in other paleoclimate records (Beer et al., 1988; Halfman and Johnson, 1988; Anderson, 1991), implying control by a common mechanism. Correlations between the Bay cores are hampered by the error involved in the radiocarbon dating estuarine carbonates, and relatively coarse temporal resolution precludes detection of interannual fluctuations (less than 10 years). However, the core records are in general agreement and identify the following changes in fresh water inflows from the Delta. Relatively low fresh water inflow during the periods: ca. 1230–1150, 1400–1300, 2700–2600 and 3700–3450 cal yr B.P. (Ingram et al., 1996b, c) interspersed with relatively high fresh water inflow during the periods: ca. 1180–1100, 2400–2200, 3400–3100, and 5100–3800 cal yr B.P. (Ingram et al., 1996b, c). During the past 1000 years, five periods of lower than modern stream flow are detected in the cores, though at slightly different times (Ingram et al., 1996b, c). Four high-flow periods occurred in the last 700 years at Oyster Point (Fig. 9).

3.3.2. Tidal marshes

Tidal marshes began to form around the San Francisco Bay estuary about 6000 years ago, when the early Holocene rate of sea-level rise (following the melting of continental ice sheets) slowed enough to support marsh development (Atwater et al., 1979). Evidence from marshes that existed around the estuary from ca. 6200 to 3500 years (Goman and Wells, 2000) ago suggests that, initially, marshes were intermittent. Sediments deposited in incipient tidal marshes during this period alternated between marsh deposits and subtidal estuarine deposits (“Bay mud”) at Browns Island, Peyton Hill (Goman and Wells, 2000) and China Camp (Goman, 1996; Malamud-Roam, 2002). This intermittency has been attributed to short-term variations in the rate of sea level rise in the estuary, such that sediment supplies provided by inflows from the upstream watersheds were insufficient to keep pace during the periods of more rapid sea-level rise (Goman and Wells, 2000) but were sufficient to keep pace during periods of slower sea-level rise. The long-term trend since about 3500 cal yr B.P. has been toward shoaling, development and expansion of marsh deposits (Goman and Wells, 2000; Malamud-Roam, 2002). The marshes have been expanding since that time, reaching maximum extents by A.D. 1850, when human development and land-use practices reduced marsh areas around the San Francisco Bay estuary (San Francisco Estuary Project, 1992).

Carbon isotope, pollen and diatom records from tidal marshes around the Bay show evidence of increased salinity (as seen, for example by dominance of salt tolerant plants) over the last 3000 years (May, 1999; Goman and Wells, 2000; Byrne et al., 2001; Malamud-Roam, 2002; Starratt, 2003, 2004; Malamud-Roam and Ingram, 2004). This trend can be seen at Browns Island near the mouth of the Delta, where macrofossils of the fresh water plant *Phragmites communis*, which dominated the organic

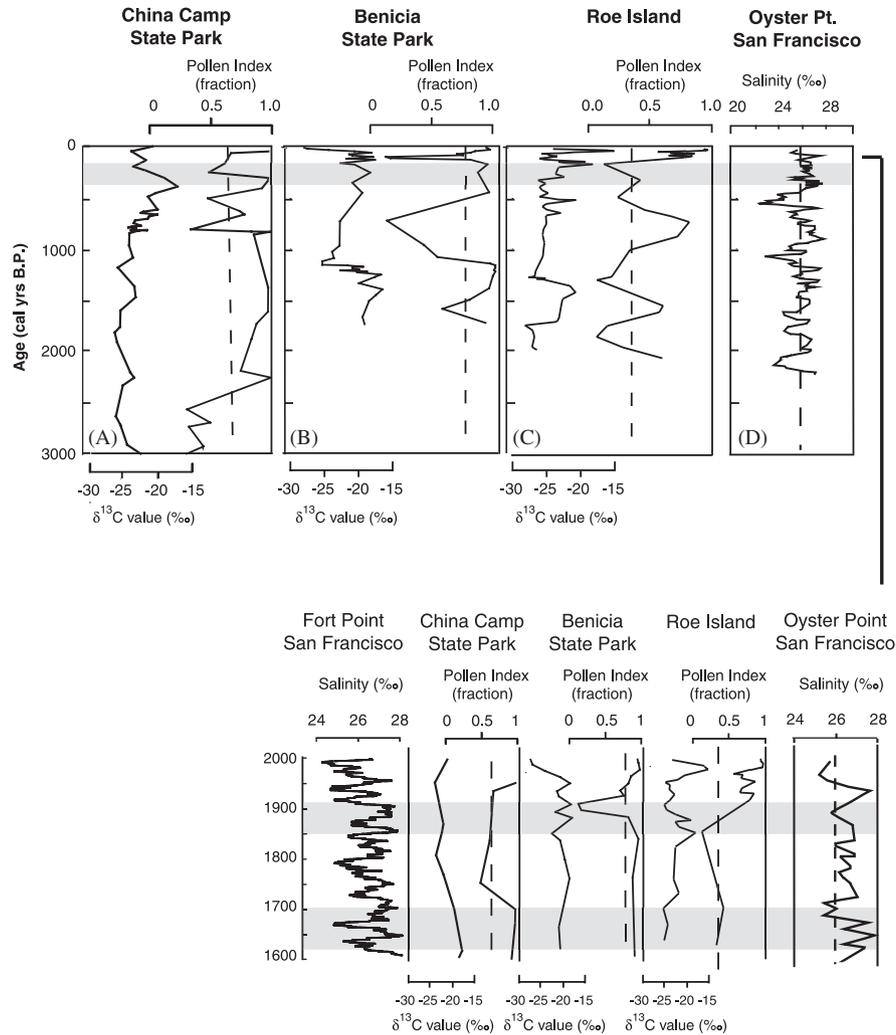


Fig. 9. San Francisco Bay estuary paleosalinity records. Top: sedimentary records covering the full history of the Holocene marshes. For marsh sites (A–C), the stable C isotope compositions are shown on the left, with $\delta^{13}\text{C}$ values increasing (less negative) to the right and indicating greater dominance of salt-tolerant grasses; the marsh pollen index is shown on the right, with *Chenopodiaceae* (pickleweed) pollen as a fraction of the sum of *Chenopodiaceae* and *Cyperaceae* (sedge) pollen, unity represents full dominance by pickleweed pollen and dashed line represents pre-European disturbance mean. (A) China Camp State Park in San Pablo Bay. (B) Benicia State Park in Carquinez strait. (C) Roe Island in Suisun Bay (source for pollen: May, 1999). (D) Oyster Point, San Francisco salinity derived from stable oxygen isotope compositions (Ingram, et al., 1996b). Bottom: the last 400 years compared against salinity at Fort Point reconstructed from Central Valley Oak tree-ring chronologies (source: Stahle et al., 2001). Shaded period ca. A.D. 1850–1900 indicates period of relatively fresh conditions seen in records and ca A.D. 1625–1675 indicates period of relatively high salinity seen in records.

sediments before 1000 cal yrs B.P., essentially disappeared and were replaced by *Scirpus americanus* (a fresh- to brackish water tulle species) (Atwater, 1980; Goman and Wells, 2000). Increasingly brackish conditions have also been indicated by the appearance of the salt tolerant *S. virginica* (pickleweed) pollen on the island (May, 1999) and Atwater (1980) found rhizomes of the salt grass *Distichlis spicata* that returned a radiocarbon date of 540 ± 120 ^{14}C yr B.P. (calibrated 2 sigma range is 310–690 cal yr B.P.).

Similar patterns of plant-species change over the past 3000 years have been detected at other marsh sites. At Peyton Hill, macrofossils of *Scirpus americanus* declined in abundance, to be replaced by *S. virginica* and *Triglochin maritima* (both salt-tolerant, high-marsh species) (Goman

and Wells, 2000; Goman, 2001). At China Camp, *Cyperaceae* pollen (probably representing species of *Scirpus*) and *Typha* pollen declined during the past 3000 years, again replaced by *S. virginica*, which has since dominated the assemblage (Malamud-Roam, 2002).

A comparison of paleoclimate records (pollen and stable carbon isotope compositions) from a series of local marshes (and salinity from Oyster Point) along the natural salinity gradient that spans the northern reach of the estuary shows substantial agreement between the records (Fig. 9). Plant assemblages at individual marsh sites respond to a host of physical processes, but the dominant physical influences are salinity and depth and duration of tidal inundation (Adam, 1990; Mitsch and Gosselink, 1993; Sanderson et al., 2000). The agreement among these

records implies salinity changes that were estuary-wide and due to changing inflow patterns. The differences between records may be attributed to a combination of methodological issues (sampling frequency, core chronology) and site-specific ecological differences (site elevation, location relative to channel, sedimentation rates over time). For example, both Roe Island (May, 1999) and Rush Ranch (Byrne et al., 2001; Starratt, 2004) are located in Suisun Bay (Fig. 2), but due to its location within the channel, Roe Island may be quite susceptible to estuary salinity fluctuations (May, 1999) perhaps more so than Rush Ranch, which is located inland and may receive runoff from surrounding uplands.

A salinity record at Fort Point, near the Golden Gate, as reconstructed from Central Valley Blue Oak tree-ring chronologies (Stahle et al., 2001), shows periods with relatively low salinity (e.g., ca. A.D. 1850–1900, Fig. 9, shaded) overlap with periods of less salt-adapted plant species, and extended (multi-decadal) periods of higher than average salinity at Fort Point overlap with increased domination of salt tolerant plants in several of the vegetation records, for example between A.D. 1625 and 1675 (Fig. 9, shaded). The tree-ring record is highly resolved, though short, and can be extended using the tidal marsh sedimentary records. Furthermore, the sedimentary records from the Estuary provide direct evidence of ecosystem responses to the climate variations that resulted in the changing salinity patterns at Fort Point.

San Francisco estuary tidal marsh paleoclimate records indicate that relatively fresh conditions existed in the Bay estuary from 3600 to 2000 cal yr B.P. and from ca. 750–300 cal yr B.P. (Byrne et al., 2001; Malamud-Roam, 2002; Starratt, 2004). This latter period of relatively fresh conditions overlaps with the period of relatively cool and moist conditions seen elsewhere in the paleoclimate record from the watershed region, particularly the San Joaquin drainage area, though interestingly, conditions appear to have become fresher in the Bay before the onset of cooling in the southern watershed (e.g., Mono lake record described by Stine (1990a) suggests a major drought occurred at 750 cal yr B.P.). These periods of fresh conditions in the estuary were interspersed with periods when more salt-tolerant plant species dominated the marshes, from 1650 to 1300 cal yr B.P., 1000–800 cal yr B.P., and over the last century, suggesting increased salinity in the estuary (i.e., reduced fresh water inflow). The second of these high salinity periods corresponds to the first of the megadroughts seen in the paleoclimate records from the eastern and central Sierra Nevada. Several marsh sites also show evidence of a period of increased estuary salinity occurring ca. 300 cal yr B.P. (Malamud-Roam and Ingram, 2004). The exact timing of this event is difficult because of variations in radiocarbon production occurring in recent centuries (Stuiver and Quay, 1980), though it may have been a response to the major drought described in tree-ring paleoclimate records from the Western US occurring

ca. 370 cal yr B.P. (A.D. 1580) (e.g., Meko et al., 2001; Brown et al., 1992; Hughes et al., 1990, 1996).

4. Discussion

4.1. Sub-decadal timescales of variability

Much progress has been made in recent years in reconstructing the past behavior of the El Niño-Southern Oscillation phenomenon, as recorded by the Southern Oscillation Index (SOI) and by sea surface temperatures at key locations in the tropical Pacific Ocean. These reconstructions have used a variety of proxy indicators, including proxies from remote regions that are impacted by ENSO (Stahle et al., 1998; Mann et al., 2000) and, increasingly, proxies from “core” ENSO regions in the tropical Pacific (Cobb et al., 2001; Evans et al., 2001). The location of the Aleutian-Alaskan Low pressure cell, which is partly modulated by ENSO, is the principal predictor for historical precipitation variations in California climate patterns, hence wintertime salinity variability in San Francisco Bay (Peterson et al., 1995). Now that more robust reconstructions of ENSO variations from the “core region” in the tropical Pacific are becoming available, the possibility exists of examining more fully the links between ENSO and subregional precipitation variations in the San Francisco Bay watershed.

A number of high-resolution proxies from California exhibit spectral power in the 3–7 year range that is typical of ENSO influences (Dettinger et al., 2001); however, connecting local climate variations to ENSO in either instrumental or proxy records is a complex undertaking (Schoner and Nicholson, 1989). Due to the state’s geographical location, spanning nearly 10° of latitude, ENSO events may result in either increased or decreased annual precipitation; albeit, in either case, often with extreme conditions (Peterson et al., 1995). ENSO affects precipitation more reliably in the Pacific Northwest and in the Southwest, including southern California, than in central California, with El Niños being associated with wetter conditions in the Southwest and drier conditions in the Pacific Northwest, north of California (Cayan and Webb, 1992). La Niñas bring the opposite precipitation conditions, and indeed are typically more reliable predictors of precipitation in the Northwest and Southwest than even El Niños (Dettinger et al., 2002). Cayan et al. (1999) write that “major floods in the Sierra Nevada appear to be more likely during La Niña than during El Niño”, whereas major floods in the coastal plain and coast ranges occur more frequently during El Niño years. Off the central coast, on average, El Niño years are associated with wetter years, particularly strong and very strong El Niño years (Jones and Kennett, 1999), though a 600-yr tree-ring reconstruction of Santa Barbara precipitation is due to a tendency for wet El Niño years to be extremely wet (Haston and Michaelson, 1994). Despite the occasionally unreliable linkage between ENSO and California’s climate,

a finding of potential relevance to the San Francisco Bay watershed is a reduction of Southern Oscillation Index (SOI) amplitude in the mid-19th century (Peterson et al., 1989). Such fluctuations in the strength or frequency of ENSO are part of its naturally chaotic character and might add considerably to fluctuations in the predictability and variability of California's climate (e.g., McCabe and Dettinger, 1999).

4.2. Decadal-to-multi-decadal variability

For much of the watershed, decadal (>7 years) variations account for 20% or more of total variance of annual precipitation (Cayan et al., 1998). Decade-long runs of dry and wet years may be linked to sea-surface temperatures over much of the Pacific Basin and to changing modes of circulation. Dettinger et al. (1998) mapped large-scale patterns of precipitation in Western North America for the past century, as well as patterns of tree-ring variability since the 18th century, and found them to strongly resemble one another. They confirmed that, at both ENSO and inter-decadal timescales, the total amount of winter "precipitation delivered to the cordillera of western North America has been remarkably stationary during the last 115 years", and perhaps back to the early eighteenth century. That finding applied to the full latitudinal range they studied, from, 25° to 55°N, so that an examination of a smaller region, such as the component parts of the San Francisco Bay watershed, could well be affected by changes in the distribution of the total precipitation within the general area. In particular, the "center of distribution" of precipitation moves north and south in association with warming or cooling of sea surfaces in tropical regions. Dettinger et al. (1998) also found evidence from moisture-sensitive tree rings that spatio-temporal patterns of variation in precipitation amounts across the region also showed considerable consistency back to A.D. 1710. They did, however, note a greater degree of multi-decadal variability between the mid- to late- 19th century and, 1960.

Perhaps the most important historical mode of multi-decadal climate variation in western North America is the Pacific Decadal Oscillation (PDO, Mantua et al., 1997), a multi-decadal variation of the air-sea climates of the North Pacific Ocean that is intimately connected with long-term variations in the strength and frequency of ENSO fluctuations (e.g., Dettinger et al., 2001). As with ENSO-related variability, multi-decadal climatic variations of precipitation associated with the PDO is strongest in the southern and southeastern parts of California (Biondi et al., 1999), potentially affecting the Transverse Ranges and southernmost Sierra Nevada. An index of the PDO has been reconstructed back to A.D. 1661 (Biondi et al., 1999, 2001) using moisture-sensitive tree-ring chronologies from Southern California and northern Baja California. The fluctuations of tree-ring indices associated with the PDO appear weaker in the period A.D. 1780–1880 than in the

centuries immediately before and after this period. Several other studies of PDO variation in recent centuries confirm this general finding (D'Arrigo et al., 2001; Gedalof and Smith, 2001; Gedalof and Mantua, 2002).

4.3. Century to multicentury-scale variability

As described above, intense droughts lasting over a century occurred in the region near Mono Lake, California, during the Medieval Climate Anomaly (Stine, 1994). Furthermore, paleo-precipitation records from the Methuselah Walk bristlecone pine tree-ring chronology (Hughes and Graumlich, 1996) and five similar paleoclimate records from sites in and close to the Great Basin (Hughes and Funkhouser, 1998) show a greater incidence of intense persistent droughts between A.D. 400 and A.D. 1500, at least in the Great Basin. Stine (1994) proposed that these droughts, which coincided in time with droughts at the same latitudes in South America, "may have been caused by reorientation of the mid-latitude storm tracks, owing to a general contraction of the circumpolar vortices and/or a change in the position of the vortex waves". To put these prolonged droughts into context relative to the modern "benign" climate era, Fig. 10 presents probability distributions of annual flows of the Sacramento and San Joaquin rivers, estimated for the periods A.D. 1052–1150 (898–800 cal yr B.P., spanning the first Mono lake mega-drought), A.D. 1200–1350 (750–600 cal yr B.P., spanning the second Mono lake mega-drought) and A.D. 1850–1986 (modern era). The reconstructed flows are likely skewed towards less variability and more normal distributions than historical and, presumably, actual past flows. These estimated probability distributions describe general tendencies and qualitative differences between the flows of the various periods. Overall, the distributions of reconstructed flows from each of the ca. 150-yr periods fall mostly within the range of 95% of random samples drawn from the 1000-yr records from each of the two rivers (shaded bands), indicating that the benign historical period and the medieval droughts were not completely different from the overall flow statistics of the past 1000 years. This probability distribution figure is presented elsewhere (Malamud-Roam et al., in review) and discussed in greater detail. We present it here to demonstrate that the prolonged droughts of the Medieval Climate Anomaly were not marked by extreme low flows: instead the absence of very high flow years and the reduction in frequency of moderately more-than-normal flows may have contributed more to the character of these droughts. The medieval droughts, as experienced within the Bay-Delta watershed, were century-scale periods of more frequent moderately dry years and less frequent (than modern) moderately wet years, but do not appear to include an unusual number of extremely dry years. The modern era has yielded more frequent moderately wet years and, in the San Joaquin River, a few notable extremely wet years.

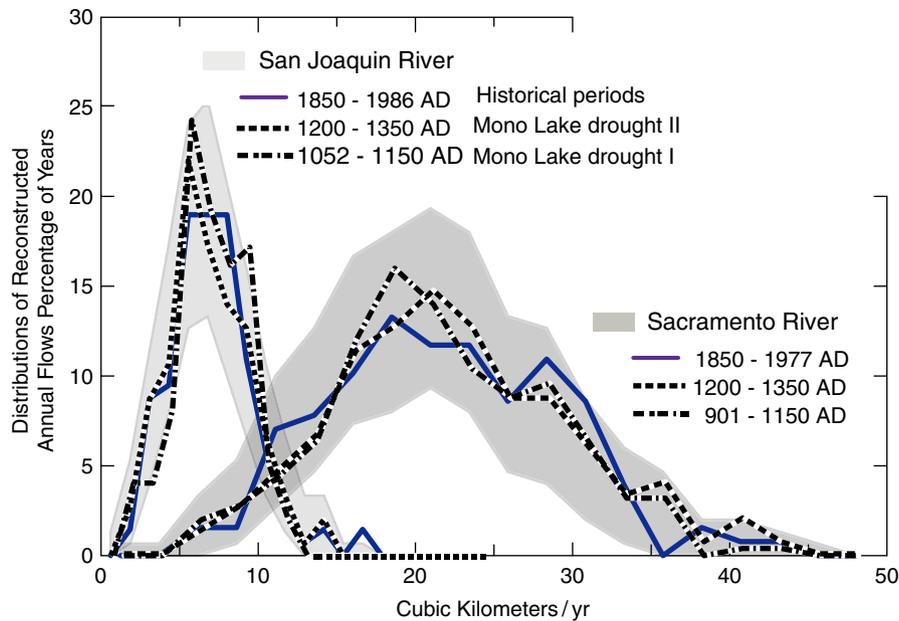


Fig. 10. Frequency (probability) distributions estimated from reconstructed annual stream flows (km^3/year) in the Sacramento and San Joaquin Rivers (reconstructed flows were provided by David Meko of the Laboratory of Tree-Ring Research at the University of Arizona, see Meko et al., 2001, 2002) for the historical period (1850 A.D. to end of reconstruction), the first (beginning of reconstructions to 1150 A.D.), and second medieval Mono Lake drought (1200–1350 A.D.); shaded areas indicate the 95% confidence intervals on distributions obtained by similarly analyzing 10,000 random samples of 150 years from the reconstructions. The confidence intervals shown here test whether or not the differences between the various distributions fall beyond the range of fluctuations that might be encountered in random subsamples of the overall reconstructions. This figure has been kindly provided by M. Dettinger at the USG.S./S.O.I.

4.4. Holocene climate variations in California

A summary of general climatic trends of the Holocene is presented in Table 4, organized geographically from coastal California (Barron et al., 2003) to the White Mountains and based upon paleoclimate studies throughout the region. Despite the different paleoclimate archives used, the major climate features seen in the records are in good agreement and are shown relative to the changes in solar insolation received at the northern latitudes. The early Holocene period was a time of rising temperatures and reduced moisture availability, with consequent changes in tree line elevations, vegetation assemblages and rising sea level. Fire apparently played a large role in the forests of the central Sierra Nevada during this period, until the vegetation cover filled in more and soil moisture increased (Edlund and Byrne, 1991; Brunelle and Anderson, 2003). The San Francisco Bay estuary was in the early stages of reforming as rapid sea level rise brought ocean waters into former river valleys (Atwater, 1979). The warming trend continued, peaking about 5000–6000 years ago, though fires were less frequent in mountain forests as they matured and conditions became more mesic. Tidal marshes were forming along the protected shores of the San Francisco Bay at this time (Atwater et al., 1979; Goman and Wells, 2000), as the post-glacial rapid rise of sea level slowed globally (Fairbanks, 1989), dropping locally to a rate of less than 2 mm/yr (Atwater, 1979). We can only speculate about sediment loads delivered to the Central Valley

lowlands during the early to mid-Holocene, though it is an area which deserves more detailed analysis. When Sierran forests were open and fires more frequent, it is likely that sediment deposition rates were high, declining as tree line moved upslope and the forests filled in, holding the soil. As sea level rose and the Bay estuary expanded inland, we can infer that deposition rates again increased due to the changes in lowland river gradients, and since then have varied largely due to climate variations, particularly increased storm activity (as seen in the flood records discussed above).

The period from ca. 4000 to 2000 cal yr B.P. stands out in every part of the study region as a time that was unusually wet, compared to the preceding millennia. The evidence for this is particularly strong in the San Joaquin watershed basin, where, for example, lake records from the Sierra Nevada suggest that the period was marked by gradual cooling and greater moisture availability (Anderson, 1990; Anderson and Smith, 1994), perhaps with intensified cooling from 3700 to 3000 cal yr B.P. (Smith and Anderson, 1992). In the Great Basin, tree line and tree-ring studies in the White Mountains (LaMarche, 1973, 1974a, b) indicate cooler and generally moister conditions (LaMarche, 1973) there, as well. Pyramid Lake had reformed and the lake level at Lake Tahoe was high enough to spill over, with some outflow received in Pyramid Lake (Benson et al., 2002). Further, a shallow lake existed in the Mojave Desert around ca. 3620 ± 70 yr B.P., suggesting very wet conditions at that time (Enzel et al., 1989).

Table 4
Reference list for climate events in California shown in Fig. 11

I.D.	Reference	Location	Year	Description
1	Ingram et al., 1996c	San Francisco Bay	1270–1380, 1675–1730, 1800–1860	High inflow
	Ingram et al., 1996c	San Francisco Bay	1200	Top of unconformity
2	Goman and Wells, 2000	San Francisco Bay	1420	Browns Is Flood
3	Malamud-Roam, 2002	San Francisco Bay	1090, 1645	China Camp flood, top of unconformity, Benicia core
4	Earle, 1993	Sacramento River	1597–1613, 1641–1657, 1664–1675, 1725–1735, 1741–1754, 1798–1821, 1854–1869, 1874–1887, 1891–1916, 1962–1973	High flow
5	Sullivan, 1982	Sacramento River	1235–1360	Large flood
			1295–1410	Flood
			1555–1615	Largest flood
			1750–1770	Flood
			1810–1820	Large flood
			1861	Historic flood
6	USBR, 2002	American River	350–550	1 flood larger than historic & gage records
			825–1300	1 very large flood
			1300–1800	3 floods larger than historic records
7	Graumlich, 1993	So. Sierra	1071–1090, 1478–1527	High precipitation
8	Stine, 1990a, b	Mono lake	1084	Post Office High Stand
			1270–1345	Rush Delta High Stand
			1400–1485	Danberg Beach H.S.
			1575–1650	Clover ranch H.S.
			1857–1919	Historic High Stand
9	Leavitt, 1994	White Mts	1080–1129	Abundant soil moisture
10	Enzel et al., 1989	Mojave Desert	1527	Silver lake deposits
11	Ely et al., 1993	U.S. southwest	1000–1200, 1400–1900	Period of numerous large floods
12	Schimmelmann et al., 2003	Sta Barbara Basin	212, 440, 603, 1029, 1418, 1605, 1840	Floods
13	LaMarche, 1974a	White Mts	1100–900, 830–650	Tree rings
14	Hughes and Graumlich, 1996; Hughes and Funkhouser, 1998	White Mts	1650–1350	Tree rings

The citation is given in the second column, geographic location of research in the third column, dates in cal yr B.P. in the fourth column, and the evidence is shown in the last column.

The San Francisco Bay appears to have been fresher during the period after 4000 cal yr B.P., as seen in marsh vegetation and sedimentary records (Goman and Wells, 2000; Goman, 2001; Byrne et al., 2001; Malamud-Roam and Ingram, 2004). Evidence of at least one extreme flood event is present in each of the stratigraphic records from Browns Island, Peyton Hill and China Camp in deposits from ca. 3600 cal yr B.P. (West, 1977; Goman and Wells, 2000; Malamud-Roam, 2002). Ingram et al. (1996b) measured isotopic compositions of fossil carbonates collected in Bay sediment cores and calculated Delta inflows of 1.5–2 times the modern, prediversion stream flow. Higher sedimentation rates in the cores from San Pablo Bay (Ingram and DePaolo, 1993; Ingram et al., 1996b,c) also support increased fresh water flow. And regional ^{14}C reservoir ages were at a minimum in Bay carbonate samples, which is consistent with greater fresh water input (Ingram, 1998).

Human occupation around the Bay was changing significantly during this period as well (Fagan, 2003). The lower rate of sea level rise after ca. 6000 cal yr B.P. allowed

the establishment of vast, contiguous tidal marshlands around the San Francisco Bay (Atwater, 1979; Atwater et al., 1979), providing abundant sources of food for the native populations. As settlements became more permanent, these early people built up mounds from their refuse, largely consisting of shells from mollusks and oysters they harvested locally. The earliest layers of excavated shell mounds date to ca. 5000 cal yr B.P. (Ingram, 1998), though earlier sites may be submerged under modern sea level, and they appear to have grown rapidly (Lightfoot, 1997; Nelson, 1909). The shell mounds provided raised platforms, offering protection from the continually rising seas as well as from higher inflows through the Delta.

This period of relatively moist conditions ended ca. 2000 cal yr B.P., and climate has since been cooling and becoming drier. The paleoclimate history of the last 2000 years is more detailed, given the availability of high-resolution proxy records (see Table 1), and several extreme climatic events, lasting decades to centuries, are seen. Fig. 11 shows extreme climatic events of the last 2000 years as recorded in paleoclimate records throughout the study

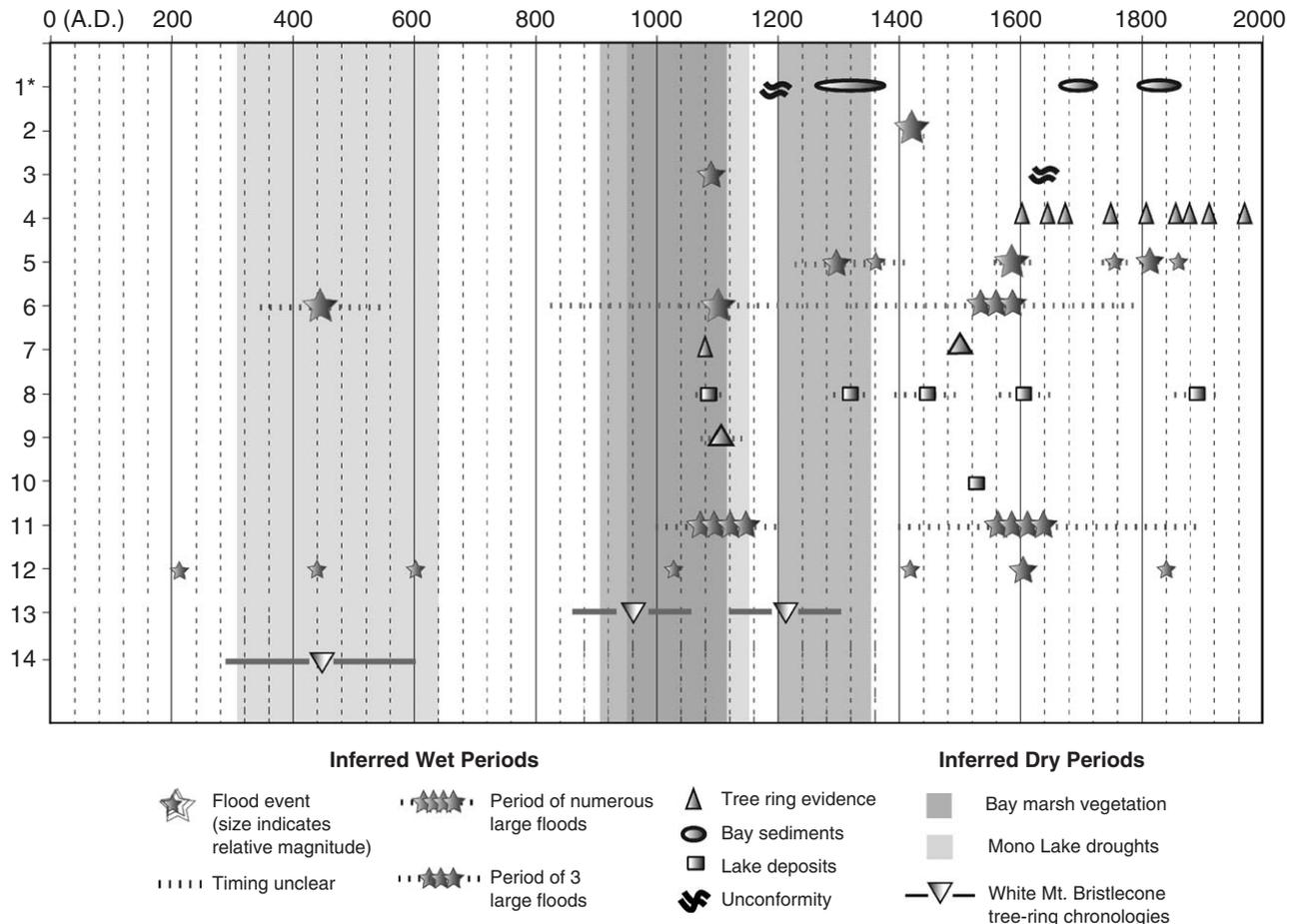


Fig. 11. Extreme climate events in California over the last 2000 years. Shaded areas indicate periods of drought and symbols indicate evidence of excess moisture from selected paleoclimate studies. Note that the unconformities identified in the Bay estuary cores reflect apparent hiatuses as inferred from the chronologies. See Table 3 for references.

area (references given in Table 3). Climate variability during the late Holocene may have played at least some role in sudden changes in indigenous societies in early California, particularly during the period of the Medieval Climate Anomaly (Jones et al., 1999; Jones and Kennett, 1999) which was associated with dry conditions in the San Francisco Bay estuary (Malamud-Roam and Ingram, 2004) and watershed region (e.g., Graumlich, 1993; Stine, 1990a, 1994; Hughes and Graumlich, 1996; Hughes and Funkhouser, 1998; Benson et al., 2002). Key wet events also stand out in multiple records, including the periods A.D. 1071–1090 (879–860 cal yr B.P.) and 1478–1527 (472–423 cal yr B.P.) in the central Sierra (Graumlich, 1993), the Great Basin (Hughes and Graumlich, 1996) and in the White Mountains (A.D. 1080–1129, Leavitt, 1994). Flood deposits and unconformities in cores collected from San Francisco Bay also date to these periods (Ingram et al., 1996c; Goman and Wells, 2000; Malamud-Roam, 2002).

4.5. Preferred timescales

Longer-term paleosalinity records from the southern reach of San Francisco Bay sediments reveal variability

with “preferred timescales” or quasi-periodicities of ca. 440, 220, 150, 90, and 60 years (Ingram et al., 1996b), and the San Pablo Bay core showed a strong period at 200 years (Ingram et al., 1996c). While uncertainties are associated with these periodicities due to the large uncertainties in the calibrated radiocarbon ages, the preferred timescales identified are similar to those found in other proxy records from the watershed and coastal marine sediments. These preferred timescales, or quasi-periodicities, appear in records from throughout California, and in some cases throughout the western hemisphere, as Schimmelmann et al. (2003) demonstrate in their summary of flood records from California, Meso- and South America, all showing timescales of 200 or 400 years. The lake-level record from Mono Lake also exhibits a strong preference for 200-year timescales (Stine, 1990a). For example, preferred timescales in several records from the Santa Barbara Basin reveal similar periodicities, including: fluctuations in thermal variability in the surface waters of the California Current centered around periods of approximately 55, 70 and 90 years over the last 900 years (Field and Baumgartner, 2000); variations in fish scale deposition of sardines and anchovies with periods centered around 55, 75, and 100

years (Baumgartner et al., 1992); radiolarian faunal changes with spectral peaks at 84, 89, and 104 years (Pisias, 1978); changes in sediment varve thicknesses with spectral peaks at 58 and 100 years over a 1000-yr record (Biondi et al., 1997), and evidence of major flood events with periods of about 200 and 400 years (Schimmelmann et al., 2003).

Hughes and Funkhouser (1998) point out the importance of apparent periodicities in climate signals in defining concepts such as “natural range of variability”, particularly in the context of managing aspects of the natural environment. The variability, on multiple timescales, seen in the paleoclimate records stresses the need to consider the possibility that the stability in precipitation that has been seen in the historic record (Dettinger et al., 1998) may not last into the future.

4.6. Spatial linkages

A 400-yr spatial reconstruction of precipitation in California from the latitude of San Francisco Bay south to the Mexican border (Haston and Michaelsen, 1997) illustrates California’s variable precipitation in recent centuries. The study found precipitation anomalies to be more common in the northern half of their study region during the period A.D. 1560–1760, and suggested that this might be the result of stronger meridional (north–south) flows in the atmosphere during that period compared to recent times. In addition, an unusual concentration of wet years and pentads occurred in the late 16th and 20th centuries, though the first of these periods also contains an extended drought. This drought (dated to the A.D. 1580s) has been detected in other records (Stahle et al., 2000) and appears to have preceded two decades of high precipitation (Haston and Michaelsen, 1997). The early 17th century “pluvial” may well be recorded in a high stand of playa lakes in southern California (Enzel et al., 1989; Enzel 1992; Enzel and Wells, 1997), the highest level of Mono Lake in the last 1000 years (Stine, 1990a), and a large flood ca. A.D. 1605 recorded in the varved sediments of the central Santa Barbara Basin (Schimmelmann et al., 1998). In a tree-ring reconstruction of temperature patterns across the western United States, Briffa et al. (1992) identified periods since A.D. 1600 with summer 6-month periods that were anomalously warm/cold. This study identified cool conditions during the first decade of the 17th century and the end of the century. Thus it appears that these periods show up in records as wet and cool (Briffa et al., 1992; Haston and Michaelsen, 1997).

The sequence of extended drought followed by an unusually wet period has been described as a “dry/wet knockout”, with major ecosystem effects, potentially mediated through fire and subsequent movement of sediment from terrestrial ecosystems into riverine and marine environments (Biondi et al., 2000). Schimmelmann et al. (1998) hypothesize that the shift from drought to wet is consistent with “an equatorward displacement of major

wind patterns and associated storm tracks”. As this sequence of events in California is part of a large-scale coherent pattern covering much of western North America, it seems very likely that it results from fluctuations in the large-scale atmospheric circulation, possibly resembling those that occurred from the 1940s through the 1990s, which have been associated with very large-scale multi-decadal sea-surface temperature anomalies linked to the PDO index (Latif and Barnett, 1994; Mantua et al., 1997). These changes have been (Mantua et al., 1997).

While these large-scale shifts in circulation patterns have definite impacts on California’s climate, it is interesting to note that the paleoclimate history of the San Joaquin watershed differs in significant ways from that of the Sacramento watershed. This is demonstrated in Fig. 8, showing Sacramento and the San Joaquin river flows reconstructed from tree-ring chronologies (Meko et al., 2002) where deviations from a generally good correspondence occur during the important periods of the 13th–14th centuries and the 17th–19th centuries. Specifically, the Mono Lake drought of 750–600 cal yr B.P. (A.D. 1200–1350), so pronounced in the San Joaquin watershed, was not seen in the Sacramento River reconstructions (Meko et al., 2001). And during the two centuries of the later “Little Ice Age”, the two river reconstructions reflect opposing conditions, so that during the 17th century the San Joaquin river flows were high, while the Sacramento river had low flows and the reverse was the case in the following century.

An intriguing question thus arises as to which river basin is most critical for determining the health of the San Francisco Bay estuary. Today it is estimated that approximately 85% of the water supplied to the Bay comes through the Sacramento drainage (SFEP, 1992); however it is also true that the San Joaquin river is largely fed by melting snow from mountains in the Central Sierra during spring/early summer (Dettinger and Cayan, 2003) and thus this river may actually play an important role in the local ecology of the Bay. A comparison of the records of vegetation response seen in cores from several marsh sites with smoothed Sacramento and San Joaquin river flows reconstructed from tree-ring chronologies (Meko et al., 2002) reveals that the ecosystems of the San Francisco Bay estuary are most sensitive to dry conditions affecting both major rivers concurrently (Fig. 12). However the relationship between the two rivers and ecosystems in the Bay tidal marshes is complex, as timing of river inflows can be critically important to the life histories of the marsh plants (May, 1999; Knowles, 2000). This is an area that needs more in-depth research, particularly as the health and maintenance of local wetlands is largely determined by the role vegetative cover has in sediment supply both in producing and trapping sediments (SFEP, 1992; Adam, 1990).

There are several key periods during the Holocene when paleoclimate records demonstrate the spatial climatic linkages between coastal and terrestrial sources. For

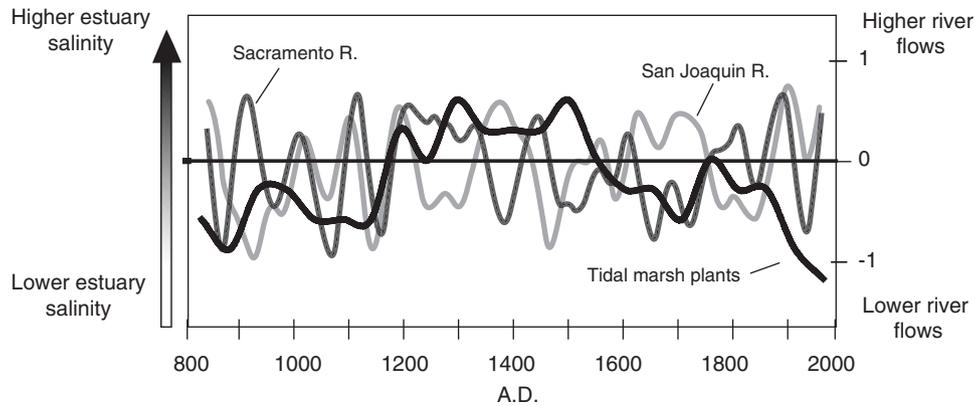


Fig. 12. Estuary salinity inferred from tidal marsh paleoclimate records (May, 1999; Byrne et al., 2001; Malamud-Roam and Ingram, 2004) compared against tree-ring reconstructed flows of the Sacramento and San Joaquin rivers from Meko et al. (2002). Salinity index is qualitative, based on vegetation patterns from five estuary marsh sites.

example, as described above, ca. 2000 cal yr B.P., terrestrial and tidal marsh records from the San Francisco Bay-Delta watershed region indicate a shift from cooler-wetter conditions to drier conditions. This shift coincided with significant changes in ocean circulation and upwelling patterns, when the ΔR values (the difference between the regional reservoir ^{14}C age and the average ocean surface water ^{14}C age) recorded in paired marine shells and carbonized plant remains in Santa Barbara Basin and Daisy Cave (Kennett et al., 1997; Ingram et al., in prep.), and in San Francisco Bay shell middens (Ingram, 1998) underwent rapid fluctuations, with the ΔR values declining by more than 500 years. Similarly, radiocarbon age differences between planktonic (surface dwelling) and benthic (bottom dwelling) foraminifers in Santa Barbara Basin sediments increased from an average of 440 years before 2000 years ago to an average of 620 years after (Roark et al., 2003). The authors note a decline in bioturbation recorded in the Santa Barbara Basin sediments (reported by Behl and Kennett, 1996) and posit an influx of older, oxygen depleted Pacific Intermediate water entering the basin at about 2000 cal yr B.P. One thousand years later, the intermittent mega-droughts of the Medieval Climate Anomaly (ca. A.D. 900–1350; Stine, 1994) coincided with a period of anomalously warm coastal ocean temperatures in the California Current, as indicated by oxygen isotope analyses of foraminifera from Santa Barbara Basin (A.D. 1200–1450; Field and Baumgartner, 2000). Oxygen isotope compositions of mussel shells from archaeological sites along the central coast also indicate that sea surface temperatures were slightly warmer than present, but, more importantly, during this time there existed a wide seasonal range in temperatures, greater than at any other time in the Holocene (Jones and Kennett, 1999). This period, in turn, was followed by cooler, and, in many places in California (Fig. 11), wetter conditions ca. 500–150 cal yr B.P. (A.D. 1450–1800, e.g., Scuderi 1993; LaMarche, 1973, 1974a), coinciding with changes in ocean surface water temperatures, inferred from positive

anomalies of $\delta^{18}\text{O}$ of planktonic foraminifera (Field and Baumgartner, 2000).

A measure of how intimately linked the Bay-Delta system is to its surrounding watershed regions is seen in the changes of unimpaired annual fresh water flows from the Sacramento River and San Joaquin River basins resulting from large-scale atmospheric changes in the ratios of precipitation falling in the northern and southern watershed regions. For example, the ratio of Sacramento River water (f_s) to San Joaquin river water (f_{sj}) was higher during the period, 1965–1976 ($f_s/f_{sj} = 2.1$) than during the period, 1977–1987 ($f_s/f_{sj} = 1.7$) (Knowles and Cayan, 2002), a change that contributed to differences in the timing of inflow to the San Francisco Bay between the two decades.

5. Conclusions

The examples given above illustrate the degree to which climate variability in the watershed, including its internal spatial structure, is coupled to the larger-scale climate system, and the extent to which paleoclimate research has already contributed insights on these connections. As the number of proxy records with high temporal resolution reflecting major features of the global climate increases (Bradley et al., 2003), and as they are complemented by multi-century runs of coupled ocean-atmosphere climate models with reasonable estimates of forcing (for example, Bolville and Gent, 1998), the possibility increases of better understanding the range of possible future natural contributions to the climate variability of the watershed. These various regional to global scale influences have resulted in a complex history of climate and hydrology for San Francisco and its watershed.

Climate over California during the last century has been relatively stable (Dettinger et al., 1998) when examined in the context of the past 2000 years, even with the severe droughts of the, 1930s and the mid-1970s. Several preferred timescales of interdecadal change are shared by

paleoclimate records from San Francisco Bay estuarine and marsh sediments, coastal sediments, and lake sediments in the watershed, and these timescales of climate variability are easily missed when only the climate of the last 100–150 years is considered. Longer term records, including tree rings, and sediments from lakes and from in and around the San Francisco Bay, provide a means of extending records to cover paleoclimate for much of the Holocene. When these longer paleoclimate records are considered, current drought conditions experienced in the US Southwest do not appear out of the range of natural variability. However, the warmer temperatures associated with anthropogenic global warming, may exacerbate such conditions. It is notable that while calculated stream flows through the Delta during the Medieval Climate Anomaly were approximately one-third lower than modern pre-diversion flows (Ingram et al., 1996b), they were still greater than modern flows unadjusted for diversion. In other words, actual fresh water inflow to the Bay is lower today than it has ever been during the Holocene, and this during a period when climate has been relatively stable.

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