

Late Holocene $\delta^{13}\text{C}$ and pollen records of paleosalinity from tidal marshes in the San Francisco Bay estuary, California[☆]

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Abstract

Records of stable carbon isotopes ($\delta^{13}\text{C}$) are presented from cores collected from four San Francisco Bay marshes and used as a proxy for changes in estuary salinity. The $\delta^{13}\text{C}$ value of organic marsh sediments are a reflection of the relative proportion of C_3 vs. C_4 plants occupying the surface, and can thus be used as a proxy for vegetation change on the marsh surface. The four marshes included in this study are located along a natural salinity gradient that exists in the San Francisco Bay, and records of vegetation change at all four sites can be used to infer changes in overall estuary paleosalinity. The $\delta^{13}\text{C}$ values complement pollen data from the same marsh sites producing a paleoclimate record for the late Holocene period in the San Francisco Bay estuary. The data indicate that there have been periods of higher-than-average salinity in the Bay estuary (reduced fresh water inflow), including 1600–1300 cal yr B.P., 1000–800 cal yr B.P., 300–200 cal yr B.P., and ca. A.D. 1950 to the present. Periods of lower-than-average salinity (increased fresh water inflow) occurred before 2000 cal yr B.P., from 1300 to 1200 cal yr B.P. and ca. 150 cal yr B.P. to A.D. 1950. A comparison of the timing of these events with records from the California coast, watershed, and beyond the larger drainage of the Bay reveals that the paleosalinity variations reflected regional precipitation. © 2004 University of Washington. All rights reserved.

Keywords: Late Holocene; $\delta^{13}\text{C}$; Pollen records; San Francisco Bay estuary

Introduction

Coastal and estuarine tidal marsh sediments contain evidence of paleoenvironmental conditions spanning several thousand years. The San Francisco Bay estuary (Fig. 1), extending from the Sacramento–San Joaquin Delta (“the Delta”) to the Pacific Ocean, is the largest urbanized estuary on the Pacific coast of North America, and for 6000 yr has supported highly productive tidal marshes (Atwater et al., 1977, 1979). The estuary attained its modern extent after initial rapid postglacial eustatic sea-level rise (average rate of rise was about 6 mm/yr) slowed to an average rate of between 1 and 2 mm/yr. This occurred approximately 6000 yr ago (Atwater et al., 1977; Fairbanks, 1989). Evidence

from the accumulating estuary and marsh sediments indicates that there have been alternating periods of fresher and more saline conditions in the Bay throughout the late Holocene (Byrne et al., 2001; Ingram and DePaolo, 1993; Ingram et al., 1996a, 1996b).

The principal control on salinity in the San Francisco Bay is the amount of fresh water passing through the Delta, a function of the amount of runoff from precipitation falling over 40% of the state of California (Peterson et al., 1989, 1995). Before artificial diversions of California rivers and upstream storage of water in reservoirs, the amount and timing of inflow depended upon, and so reflected, regional precipitation patterns (Cayan and Peterson, 1989). A record of paleosalinity in this estuary can be used to infer past climate conditions over a large region of the state. The relationship between tidal marsh plant distributions and adjacent estuarine salinity has been previously documented (Atwater and Hedel, 1976; Atwater et al., 1979; Goman, 2001) and this relationship has provided the basis for paleoclimate research in the Bay (e.g., Byrne et al., 2001; Goman and Wells, 2000).

[☆] A supplementary table entitled “Stable carbon isotopic compositions and calculated C_4 fraction from four marsh sites in the San Francisco Bay estuary” is available in a data repository and may be found on (<http://www.sciencedirect.com>).

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Here, we present records of vegetation change in four tidal marshes in the San Francisco Bay estuary, based on stable carbon isotopic compositions ($^{13}\text{C}/^{12}\text{C}$ or $\delta^{13}\text{C}$) and pollen contained in the organic sediments from a total of six sediment cores. The two proxy methods complement each other well. Fossil pollen has been used as a proxy for paleovegetation change in tidal marshes successfully (e.g., Byrne et al., 2001; Cole and Liu, 1994; Davis, 1992; May, 1999). The advantages added using stable carbon isotopes include better identification of tidal marsh grasses and a method quantifying the relative proportions of C_3 and C_4 plants that occupied the study sites at times in the past. Together, these methods are used to infer past estuary salinity. The four marshes span a range of salinity conditions in the estuary so that the records of change produced can be used to infer system-wide changes related to paleoclimate variability.

Stable carbon isotopes in estuary tidal marshes as a paleosalinity proxy

The majority of terrestrial plants use the C_3 photosynthetic pathway; however, certain grasses have evolved the C_4 pathway as an adaptation to water stress in arid and semiarid environments. Two such grasses are found in the San Francisco Bay tidal salt marshes (i.e., *Distichlis spicata* and *Spartina foliosa*, Table 1), where in many respects high salinity poses a similar stress on plants. C_3 plants are isotopically lighter ($\delta^{13}\text{C} \sim -29\text{‰}$) than C_4 plants ($\delta^{13}\text{C} \sim -12\text{‰}$) and the reasons for this difference are described elsewhere (O'Leary, 1981; O'Leary et al., 1992). The carbon isotopic composition of plants is preserved in marsh sedimentary organic matter, with relatively little change due to decomposition (Ember et al., 1987; Malamud-Roam and Ingram, 2001). The contribution of various plant species to the surface sedimentary organic matter (SOM) of several marshes in the Mississippi deltaic plain was modeled using the $\delta^{13}\text{C}$ value of the organic sediments (Chmura et al., 1987). This model was later used to predict SOM isotopic values for several coastal marshes (Chmura and Aharon, 1995), demonstrating that past estuarine facies may be reconstructed from the marsh SOM. This model was used in the San Francisco Bay marshes and a good correlation ($R^2 = 0.75$) was seen between the $\delta^{13}\text{C}$ values of modern tidal marsh SOM and fraction of C_4 plants present in the surface plant assemblages (Malamud-Roam and Ingram, 2001).

Stable carbon isotopes have been used in tidal marsh settings as a proxy for changing salinity patterns. Delaune (1986) produced a record of changing salinity conditions in a marsh in the Mississippi deltaic plain. The results from the isotopic record were verified by historical records of changes in the local hydrology. In the San Francisco Bay, $\delta^{13}\text{C}$ values were measured in a core from a marsh in Suisun Bay and the inferred vegetation changes were in agreement with other proxy records from the same core (Byrne et al., 2001).

Stable carbon isotopes have several advantages as an environmental proxy: They provide a method of quantifying changes in vegetation composition (i.e., the relative input of C_4 plants to the marsh assemblage); the methodology requires relatively little labor (compared with pollen or diatom identification); and because there are only two C_4 grasses in the San Francisco Bay area, this method can supplement pollen records that cannot distinguish different grass types, including upland species.

Study sites

We selected four marsh sites along the northern reach of the San Francisco Bay estuary (Fig. 1) because of the natural salinity gradient that exists there. The adjacent estuary surface water salinity includes a range from relatively high salinity at China Camp (annual average salinity for period 1988–1994 was ca. 27‰, U.S.G.S., 2003) to relatively low salinity at Browns Island, near the mouth of the Delta (annual average salinity for period 1988–1994 was 4‰, U.S.G.S., 2003). Human disturbance has impacted all the tidal marshes surrounding the San Francisco Bay estuary; therefore, we selected sites that were minimally impacted and never diked.

The coring locations were all above mean higher high water for comparability and to ensure the longest records. Two sites were cored at China Camp, CC S1 and CC S2. The site CC S1 was surrounded by a mix of marsh species including *Scirpus* spp. (*californicus*, *robustus*), *Typha latifolia*, *Salicornia virginica*, *D. spicata*, and *Grindelia stricta*. The second site, CC S2, was primarily surrounded by *Sal. virginica* and some *D. spicata*. We have presented data from two coring sites at Benicia State Park, BSP 825 and BSP 3401. The site BSP 825 was surrounded by *Sal. virginica*, with some patches of *D. spicata* and scattered *Atriplex tridentata*. The site BSP 3401 was surrounded by a mix of *Sal. virginica*, *D. spicata*, and *Sc. americanus*. The plant cover surrounding the coring site on Roe Island included *D. spicata* and *Sal. virginica*, and the cover surrounding the Browns Island site included *Juncus* spp., *Triglochin maritima* and some *Sal. virginica* (May, 1999).

Methods

Field techniques

All cores were taken using a 1-m Livingston piston corer with a 5-cm diameter. The cores were collected in a series of meter-long “pushes” with some overlap between. A plastic (butyrate) sleeve lined the inside of the coring device to reduce compaction during extrusion and facilitate X-radiography. The sediments were kept in the butyrate liners and transported intact to laboratory facilities at the University of California, Berkeley.

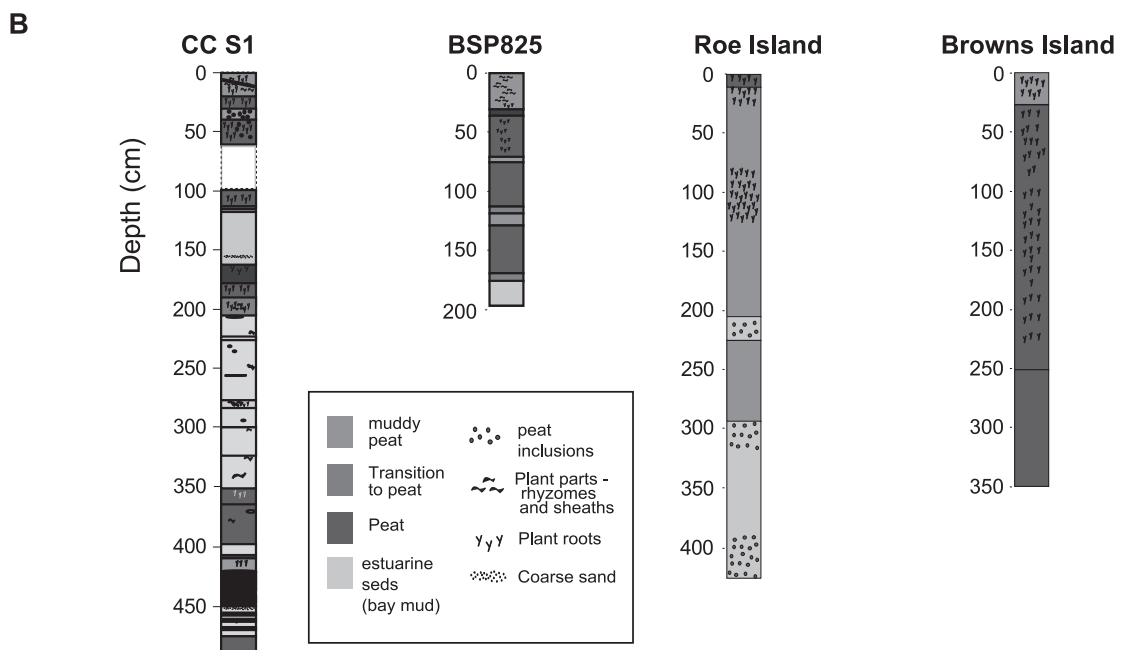
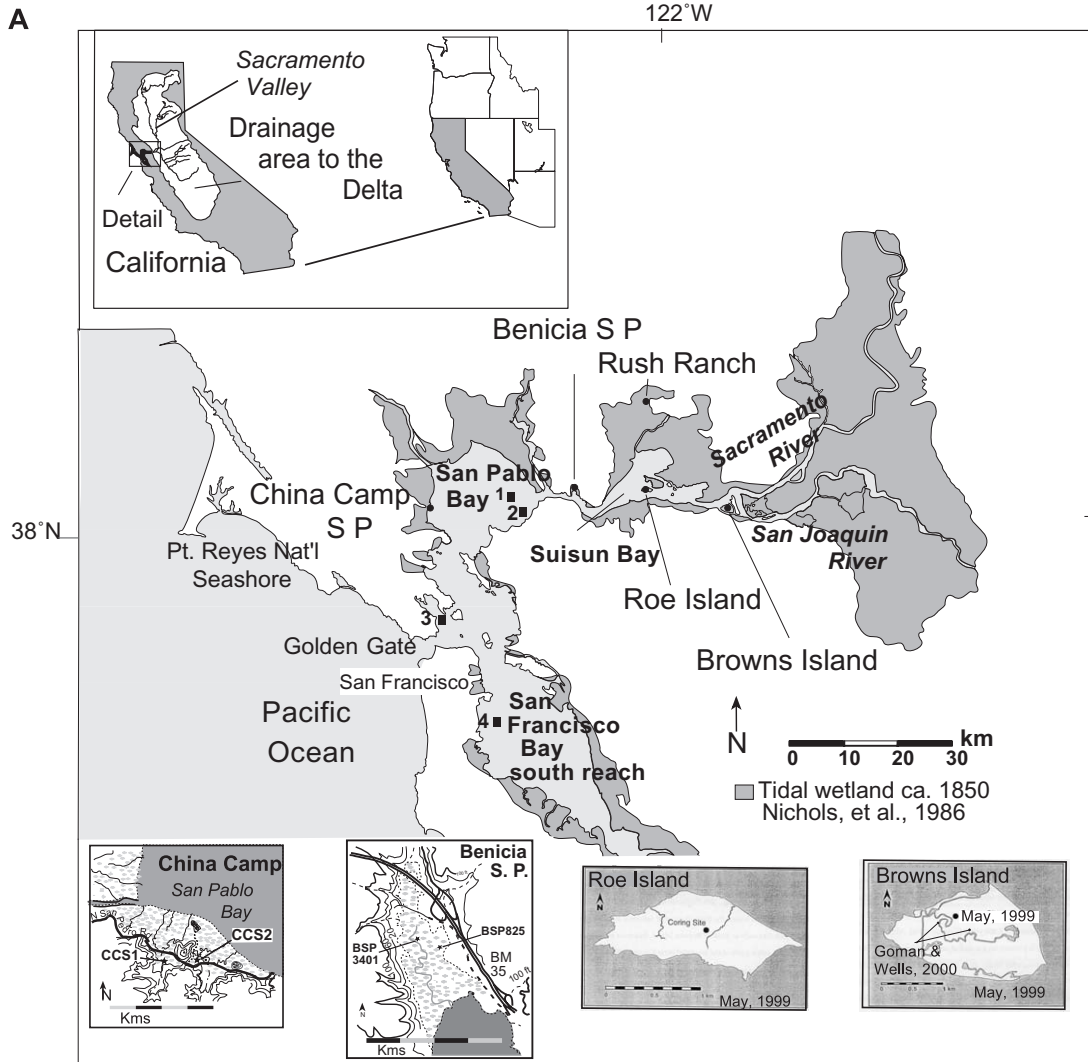


Table 1
 $\delta^{13}\text{C}$ values from multiple replicates of homogenized sediment samples from Roe Island and Browns Island

Depth (cm)	Mean $\delta^{13}\text{C}$ (‰)	Standard deviation	<i>n</i>
<i>Browns Island</i>			
15	−23.27	0.19	3
19	−23.56	0.11	4
106	−25.94	0.21	3
121	−26.95	0.13	3
146	−27.04	0.31	3
151	−27.00	0.14	4
156	−26.52	0.06	3
186	−26.06	0.22	3
74	−22.84	0.23	3
88	−26.51	0.10	3
104	−26.18	0.01	3
<i>Roe Island</i>			
26	−19.05	0.17	3
30	−17.29	0.55	3
38	−21.18	0.04	3
42	−22.03	0.38	3
46	−22.25	0.04	3
54	−22.83	0.30	3
62	−19.65	0.12	3
70	−20.22	0.07	3
80	−19.23	0.01	3
96	−22.76	0.27	3
112	−25.09	0.03	3
116	−24.61	0.04	3

Stable carbon isotopic analysis

Carbon isotopic ratios ($^{13}\text{C}/^{12}\text{C}$) were measured on a total of 36 samples from one core at Roe Island; 37 samples from one core at Browns Island; 68 samples from two cores at Benicia State Park; and 103 samples from two cores at China Camp State Park. Samples were collected at intervals of 3–5 cm on average. Sediments were sieved at 125- μm ; the fraction above 125- μm was assumed to represent locally derived plant detritus. The >125- μm fraction was then washed in 0.1 N HCl to remove carbonate material, and rinsed with distilled water. The residue was dried in a 100 °C oven for 24 h, and allowed to cool in a desiccator for at least 1 h. Samples were crushed using a mortar and pestle, then (if the sample size was sufficient) placed in a jar with two steel rods and rolled on a rolling mill for at least 24 h to produce a homogenous powder. Depending on the organic content of the sediments, 7–20 mg of the powdered samples was weighed into small aluminum capsules. Sediments with very low organic content, determined by loss on ignition (LOI), required greater amounts of material per sample. The capsu-

les were placed on the carousel of an automated 20/20 Europa mass spectrometer. The capsules were automatically dropped into the mass spectrometer, and combusted at approximately 1000 °C. The CO_2 produced was purified through several traps before introduction into the mass spectrometer. Stable isotopic measurements are calibrated against the standard NIST 1547 peach leaves. Precision for the $\delta^{13}\text{C}$ analyses is 0.1‰ or better, for five replicates.

Stable carbon isotope ratios are reported in the conventional delta notation in per mil (‰) relative to the PeeDee Belemnite standard (PDB), where

$$\delta^{13}\text{C}_{\text{sample}} = \left[\frac{^{13}\text{C}/^{12}\text{C}_{\text{sample}}}{^{13}\text{C}/^{12}\text{C}_{\text{standard}}} - 1 \right] 1000$$

(Craig, 1957).

The C_4 contribution to the organic sediments at each sample depth was calculated using an average non- C_4 $\delta^{13}\text{C}$ value of −27.04‰ and an average C_4 value of −13.1‰ (based on frequency distributions of C_3 and C_4 plants in the Bay estuary; Malamud-Roam and Ingram, 2001) and the following mixing equation:

$$-13.1\text{‰}(X) + [-27.04\text{‰}(1 - X)] = \delta^{13}\text{C}_{\text{sediment } i} \quad (1)$$

The $\delta^{13}\text{C}_{\text{sediment } i}$ is the observed isotopic value of the sediment sample from depths i and X is the fraction of plant cover represented by C_4 vegetation at the given depth i .

Pollen

We analyzed 27 pollen samples from one of the cores from Benicia State Park (BSP 825) and 33 samples from one of the cores from China Camp State Park (CC S1). Preparation followed standard extraction procedures (Faegri and Iversen, 1989), with *Lycopodium* spores added to all but six samples to allow calculation of concentrations. Samples were sieved through a 125- μm mesh after the KOH treatment to remove excess coarse organics (Byrne et al., 2001).

Chronology

We dated marsh plant seeds (when present) or identifiable rhizomes from the cores, using radiocarbon methods, at the Center for Accelerator Mass Spectrometry (CAMS), Lawrence Livermore National Laboratory. When rhizomes were sampled, we used the $\delta^{13}\text{C}$ value of −13‰ in the analysis for radiocarbon dating. Rhizomes also include errors associated with the depth to which they extend in the sediments. Error estimates include the potential depth that the rhizomes could extend (U.S.D.A., 2002). We used CALIB rev. 4.3 to convert ^{14}C years to calibrated ages (Stuiver and Braziunus, 1993;

Fig. 1. (A) The San Francisco Bay estuary with prehistoric tidal marsh extent. Research sites for this study are indicated (solid circles). Locations for the San Francisco Bay cores are also indicated (squares) and numbered: (1) San Pablo Bay core (Ingram and DePaolo, 1993); (2) San Pablo Bay core (Ingram et al., 1996a); (3) Richardson Bay core (Ingram and DePaolo, 1993); and (4) Oyster Point (Ingram et al., 1996b). Bay estuary water salinity decreases with distance from the Golden Gate. Location maps for each marsh site are included. (B) Core stratigraphies. Depth (cm) is shown on the left. Core descriptions for Roe Island and Browns Island are adapted from May (1999).

Stuiver et al., 1998). The reported age ranges and sedimentation rates reflect 2σ error range.

Results

A total of six cores were collected from the four tidal marshes. The cores predominantly consisted of peat and a

mix of silt and clay (mud), with some incompletely decomposed plant fragments (Fig. 1B).

Replicate isotopic analyses of sediment samples ($n = 3$) demonstrate that the samples are well homogenized (Table 1). While there is some variability in the standard deviation of these replicates, the average deviation for all the replicates is 0.16, and none have a standard deviation exceeding 0.55. By contrast, the minimum difference between the $\delta^{13}\text{C}$

Table 2
Core chronologies and sedimentation rates for the four tidal marsh sites

Site—depth (cm)	Material	Lab no (CAMS)	Age ^{14}C year B.P.	$2\text{-}\sigma$ Age range cal year B.P.	Sed rate (cm/year)	
					Max.	Min.
<i>Browns Island</i> ^a						
BI—59–60	Sc ^b seeds	46,627	240±80	3–424	20.0	0.14
BI—88–89	Sc and unk seeds	40,777	810±80	667–789	0.08	0.04
BI—128–129	Sc and unk seeds	40,778	1050±40	930–972	0.22	0.15
BI—148–49	Sc and unk seeds	46,624	1210±50	1060–1227	0.15	0.08
BI—178–179	unk seeds	46,625	1270±80	1072–1287	2.50	0.50
BI—208–09	Sc and unk seeds	46,626	1560±50	1390–1523	0.13	0.09
BI—239–40	Sc and unk seeds	40,780	1780±80	1571–1819	0.17	0.10
BI—322–23	Sc seeds	40,781	2330±70	2213–2357	0.15	0.13
<i>Roe Island</i> ^a						
RI—59–60	Sc and unk seeds	46,615	190±50	2–294	30.0	0.20
RI—99–100	Sc seeds	46,616	390±60	323–510	0.19	0.12
RI—139–40	Sc seeds	46,617	820±140	654–917	0.12	0.10
RI—179–80	Sc seeds	46,618	1300±60	1173–1289	0.11	0.08
RI—199–200	Sc seeds	46,619	1370±40	1272–1305	1.25	0.20
RI—229–30	Sc seeds	46,620	1720±80	1532–1712	0.12	0.07
RI—254–55	Sc and unk seeds	46,621	1830±60	1634–1860	0.25	0.17
RI—279–80	Sc and unk seeds ^c	46,622	1800±60	1627–1819	–	–
RI—344–45	woody frags	46,623	2110±90	1950–2299	0.28	0.21
<i>Benicia State Park (BSP825)</i>						
BSP—50–51	rhizome ^d	75,223	130±40	1–274	51.0	0.19
BSP—71–72	rhizome ^d	78,177	270±130 ^e	2–482	21.0	0.10
BSP—78–79	Sc seeds	86,845	1175±40	1010–1170	0.01	0.01
BSP—85–86	Sc seeds	75,224	1170±40	1007–1170	–	–
BSP—98–99	Sc seeds	78,178	1200±40	1060–1174	5.0	0.40
BSP—150–51	Sc seeds	63,815	1290±50	1173–1285	0.47	0.46
<i>Benicia State Park (BSP3401)</i>						
BSP—123–26	Sa seeds	99,101	1870±50	1698–1900	0.07	0.07
BSP—240–41	Sa and unk seeds	97,906	2985±40	3059–3269	0.08	0.08
BSP—351–52	Sa seeds	97,907	3345±40	3473–3642	0.30	0.27
<i>China Camp State Park</i>						
CC—54–55	seeds	80,346	405±40	344–509	0.16	0.11
CC—153–54	seeds	80,347	915±40	763–915	0.24	0.24
CC—182–83	rhizome ^d	80,349	930±40	788–923	3.63	1.16
CC—254–55	seeds	80,348	520±640 ^f	–	–	–
CC—368–69	seeds	80,350	2540±40	2511–2743	0.11	0.10
CC—450–51	seeds	75,225	3330±40	3474–3633	0.09	0.09
CC—470–71	seeds	75,226	3430±60	3591–3816	0.17	0.11

Two cores were dated for Benicia State Park.

^a Dates have been reported elsewhere (May, 1999) and have been recalibrated here using CALIB 4.3.

^b Seeds are abbreviated as follows: Sc = *Scirpus* (species not known); Sa = *Salicornia virginica*; unk = unknown.

^c This date is questionable—possible contamination during sampling.

^d *Distichlis spicata* rhizomes, implying that date is younger than surrounding material at that depth. Dates are corrected for the larger $\delta^{13}\text{C}$ value of this C4 plant.

^e The large uncertainty is due to very small sample size after chemical pretreatment (0.04 mg C).

^f The large uncertainty is due to very small sample size; calibration not done for this sample.

values of common C_4 plants and C_3 plants in the Bay estuary is 9‰ (if the algae *Ulva* sp. is included) and the common difference is greater than 12‰ (Malamud-Roam and Ingram, 2001).

Chronology

The cores are well dated. With the exception of BSP3401, each core has an average of two dates per meter of sediment and each meter represents approximately 1000 yr. All but four samples dated consisted of seeds that were deposited on the marsh surface by local plants. Seeds were collected for analysis from samples of 1-cm thickness, which represents approximately 5–10 yr based on average sedimentation rates. One of the four non-seed samples

consisted of “woody fragments” that were never identified (May, 1999). The other three samples were rhizomes of *D. spicata*. Roots of *D. spicata* grow to a minimum depth of approximately 5 cm (U.S.D.A., 2002), and so we include error bars of up to 10 cm. The calibrated radiocarbon dates indicate that the marsh sediments represented in our cores began accumulating 2000–4000 yr ago. Sediment accumulation rates varied over the length of the cores (Table 2), although the overall average rate of sedimentation for each core corresponds well with the calculated average rate of sea-level rise of 1.2 mm/yr for the south San Francisco Bay (Atwater et al., 1977). The variability may reflect local or estuary-wide fluctuations in sediment supply, local processes resulting in compaction or subsidence, and the inherent error involved with sampling and radiocarbon dating.

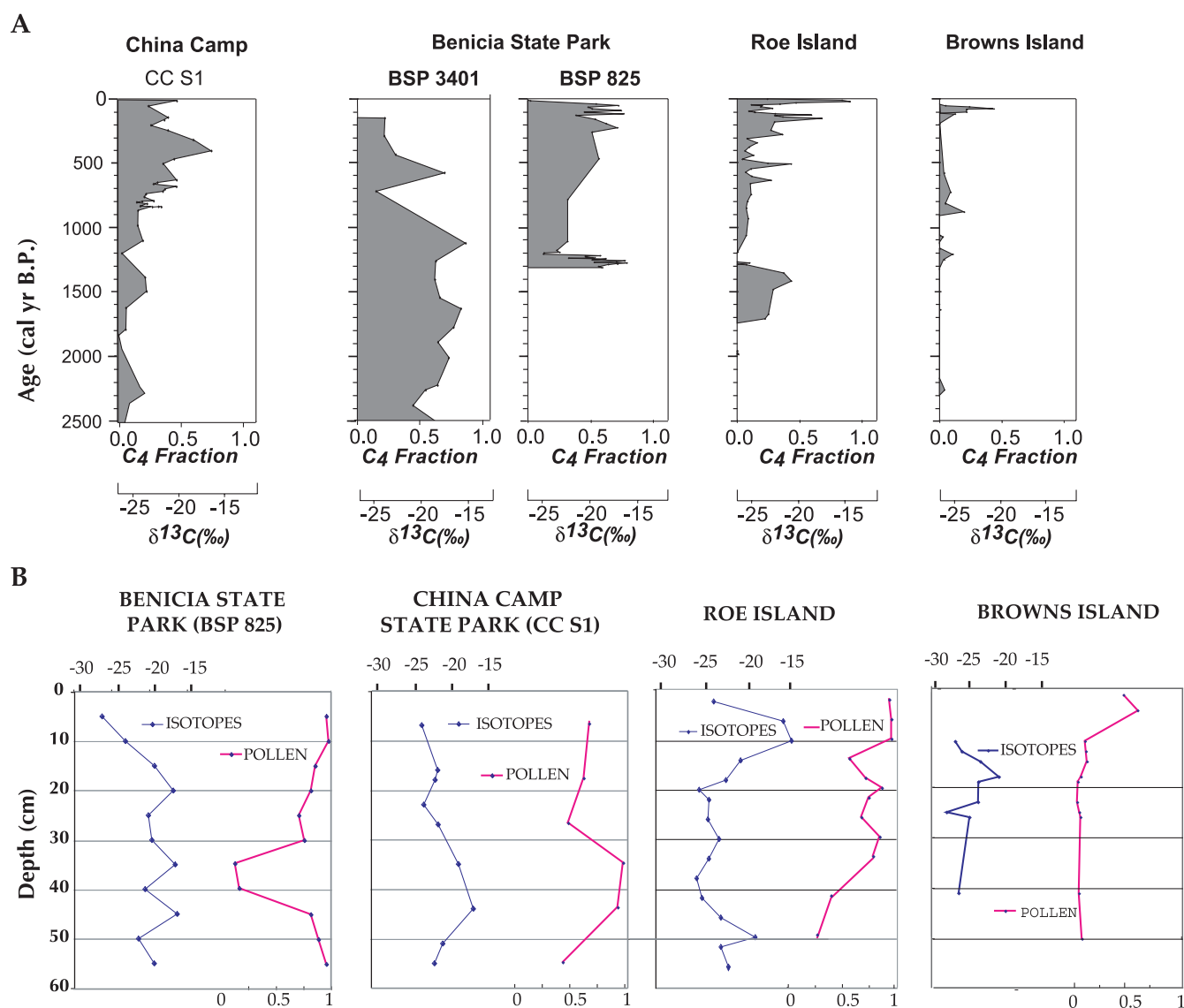
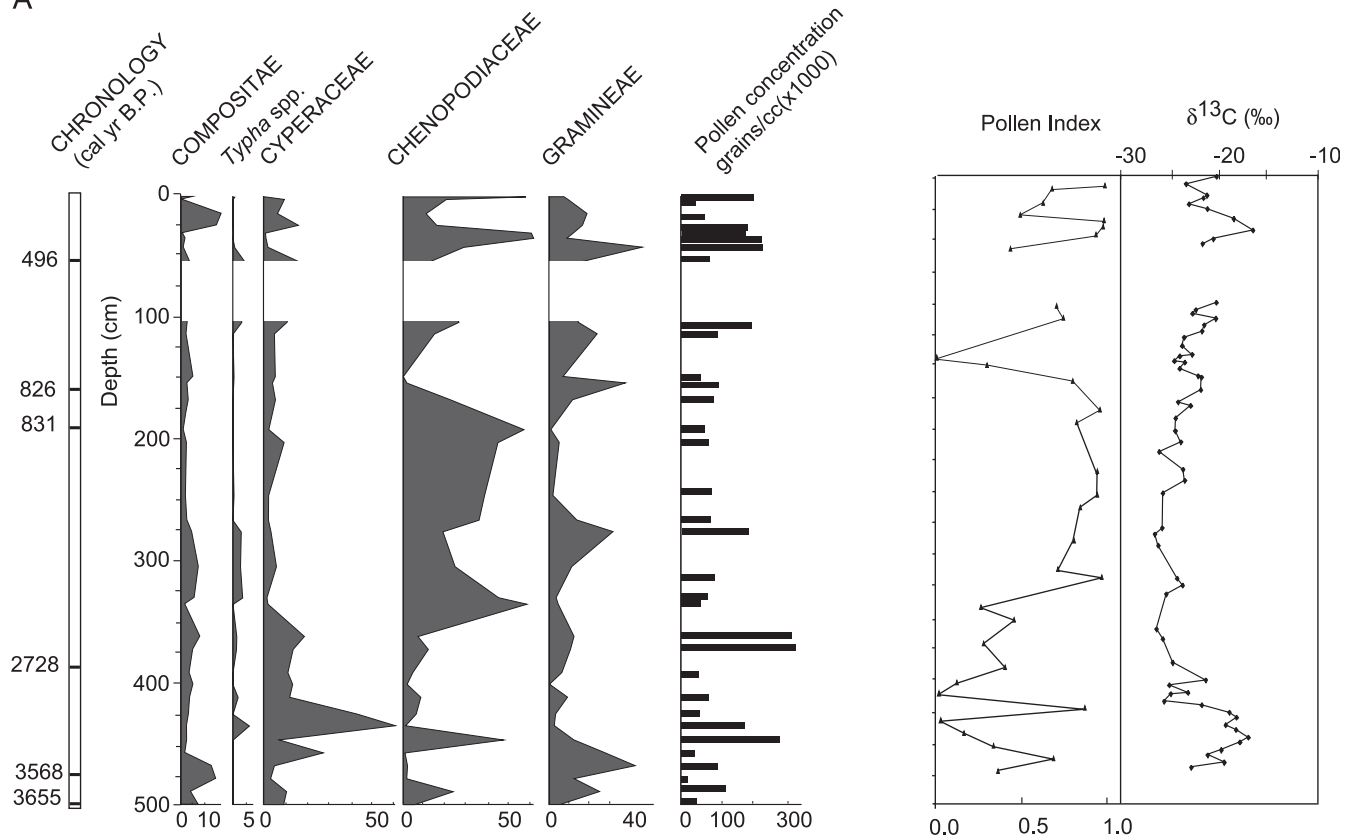
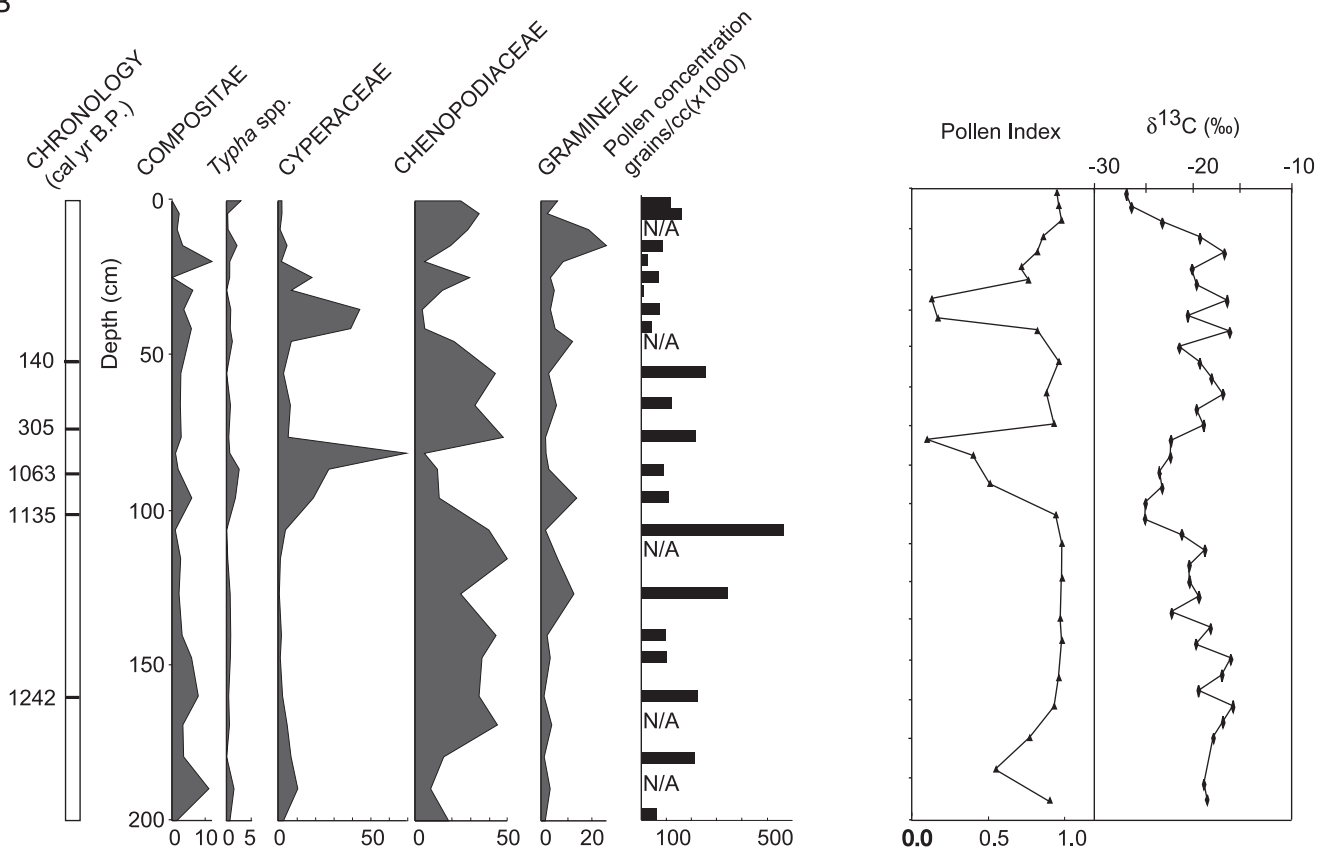


Fig. 2. (A) Changes in the $\delta^{13}C$ values (‰) and the fraction of C_4 input to the organic sediments in cores collected from four tidal marshes in the San Francisco Bay estuary. Ages (cal yr B.P.) are on the y axis. Sites are ordered geographically from west to east (see Fig. 1). (B) $\delta^{13}C$ values (‰) for top 60 cm of cores. Vertical axis is depth (cm); see text for explanation.

A



B



Carbon isotopes in marsh cores

The carbon isotopic compositions of the Browns Island core have a mean $\delta^{13}\text{C}$ value of $-26.18\text{‰} \pm 1.58$. This site exhibits the least variability of $\delta^{13}\text{C}$ with depth of the four marsh sites (see Supplementary data). The mean carbon isotopic value for Roe Island was $-23.80\text{‰} \pm 2.81$. Two cores were sampled from Benicia State Park: The mean for BSP 825 was $-20.04\text{‰} \pm 2.82$ and the mean for BSP 3401 was $-19.14\text{‰} \pm 3.48$. The means of the cores collected from China Camp were $-22.43\text{‰} \pm 2.51$ and $-21.92\text{‰} \pm 2.98$. Figure 2A presents the calculated C_4 fractions from all four marshes are compared on a common time axis and Figure 2B presents the $\delta^{13}\text{C}$ values of the top 60 cm from each core. These top sediments have high mineral content, possibly reflecting input from the hydraulic gold mining that took place between 1852 and 1864 in the Sierra Nevada foothills (Gilbert, 1917). The chronology for this part of the core is difficult to determine (see below), so we present the data with depth as the vertical axis.

Pollen

The pollen results from one core each collected from China Camp State Park and Benicia State Park are presented in Figure 3. The top panel presents the pollen data from China Camp, with that data summarized in a normalized pollen index described below and by the stable carbon isotope composition of the core, all relative to depth. The bottom panel presents these data from Benicia State Park core. Pollen analyses on the two cores from Browns Island and Roe Island were previously reported (May, 1999) and the data are not presented here. Pollen has been used as a salinity proxy in tidal marsh settings (Byrne et al., 2001; Davis, 1992; May, 1999). To compare the results from pollen analyses of all the cores with the stable carbon isotope results, we follow the format used by May (1999) and present the pollen data as an index of total Chenopodiaceae pollen (representing *Sal. virginica*) to the sum of Chenopodiaceae and Cyperaceae pollen (representing primarily *Scirpus* spp. in the estuary). This index serves as a summary of pollen data that primarily reflect salinity on the marsh surface.

Discussion

Applicability of $\delta^{13}\text{C}$ to core sediments

The isotopic values of common marsh species of the San Francisco Bay estuary have been published elsewhere (Byrne

et al., 2001; Malamud-Roam and Ingram, 2001), and there is an approximately 10‰ difference in $\delta^{13}\text{C}$ values between C_3 and C_4 plants. The two C_4 plants found in San Francisco Bay tidal marshes occupy different niches, but both are adapted to relatively high salinity. *S. foliosa* dominates the intertidal “low marsh” zone (i.e., at mean tide level and below, see Atwater and Hedel, 1976), at locations within the estuary where the mean surface water salinity is greater than 15‰. It is replaced by *Sc. californicus* in locations where the mean salinity ranges do not exceed 12‰ (Byrne et al., 2001). *D. spicata* is common on the “middle” and “high marsh” (i.e., mean tide level to above mean higher high water; Atwater and Hedel, 1976) of brackish and salt tidal marshes. Diversity is greater in brackish marshes with a mix that can include *D. spicata*, *Sal. virginica*, *Sc. robustus*, and *Sc. americanus* (Atwater and Hedel, 1976; Atwater et al., 1979), in contrast to tidal salt marshes with nearly continuous stands of *Sal. virginica* (Josselyn, 1983; Mahall and Park, 1976a, 1976b, 1976c). Several methods can be used to determine which of the two grasses (*S. foliosa* or *D. spicata*) comprise the C_4 fraction. In some cases, direct identification of rhizomes is possible. Inferences can be made based on the core stratigraphy (e.g., if clay-rich estuarine sediments immediately underlie the sample depth, the C_4 contribution is likely to be *S. foliosa*, a mudflat colonizer). Inferences can also be made using the macrofossil (seeds) and pollen data, as certain plant assemblages are associated with low vs. high marsh zones.

While the $\delta^{13}\text{C}$ method can benefit reconstructions of past marsh conditions, inferences about paleosalinity changes in San Francisco Bay marshes can be problematic if solely based on the isotopic values of core SOM. While the majority of C_3 plants occupying the tidal marshes prefer relatively fresh conditions, *Sal. virginica* (average $\delta^{13}\text{C}$ value = -27.21‰ ; Malamud-Roam and Ingram, 2001) tolerates a wide range of salinity, including mean annual soil salinities of 19–81‰ (Mall, 1969). Decreased isotopic values in a sediment core could thus indicate an increase in fresher tidal marsh plants such as *Sc. californicus* or *Scirpus acutus*, or an increase in the more salt-tolerant *Sal. virginica*. An independent means of detecting *Sal. virginica* is therefore important for interpretations of the isotopic record. Fortunately, *Sal. virginica* pollen is easily identified in marsh sediments, so that the two methods produce complementary records (Byrne et al., 2001). Finally, when interpreting these two proxy records together, we consider that changes in the $\delta^{13}\text{C}$ record may lag those in the pollen record as pollen is laid down on the marsh surface while the isotopic value of the SOM reflects contributions from the whole plant, including below-ground biomass.

Fig. 3. (A) Pollen results from China Camp State Park CC S1 core as percent of total pollen. Note variable scale. Depth (cm) is shown on the y axis with calibrated radiocarbon dates indicated. Sediment loss from 50 to 100 cm was due to friction in the corer. The pollen data are summarized in an index (described in text) with greater salinity approaching unity. Carbon isotopic compositions are also shown for comparison. (B) Pollen results from Benicia State Park (BSP 825 core) as percent of total pollen. Pollen concentrations at each sample depth are shown except for samples that *Lycopodium* spores were not available (N/A).

Nonclimatic causes for vegetation changes

Byrne et al. (2001) described some of the primary potential nonclimatic influences on changing tidal marsh paleosalinity, including plant succession, tectonic movement of the marsh surface, and human modifications of the hydrological system. In addition, sea-level rise over the period represented by these cores (ca. 2000–3000 yr) has caused increasing salinity in the estuary, a trend seen in several Bay core records (Ingram and DePaolo, 1993; Ingram et al., 1996a, 1996b). As direction and rate of sea-level rise in the San Francisco Bay has remained fairly constant over the last several thousand years (Atwater et al., 1977), the expected change in plant distributions would be toward ever-more saline assemblages. Changes in vegetation patterns on the marsh surface due to plant succession as elevation of the marsh surface increases over time through accretion would also show a trend toward more salt-tolerant plants (Cuneo, 1987; Mahall and Park, 1976b; Mitch and Gosselink, 1993). An overall trend has been seen in several records. For example, B.F. Atwater and D.F. Belknap (1979, unpublished data) and Goman and Wells (2000) both noted a shift from *Phragmites* dominated peat to *Scirpus* dominated peat in cores from Browns Island. A more recent appearance of *D. spicata* rhizomes (ca. 540 ± 120 ^{14}C yr B.P.; Atwater, 1982, plate 21) supports this trend. Despite this trend, we find evidence of paleosalinity changes in the opposite direction, i.e., indicating freshening conditions on the surface (Figs. 2 and 3).

The last century and a half has seen major human-induced modifications of the hydrology of the Bay estuary (Gilbert, 1917; Nichols et al., 1986). In particular, large-scale diversions of water away from the Sacramento–San Joaquin Delta for agriculture and flood control have reduced the net river inflow to the San Francisco Bay by more than 60% (Nichols et al., 1986). The timing of Delta outflow has been altered to alleviate the consequences of these diversions, with reduced flows during the period February to mid-June for reservoir storage that result in lowered surface water elevation and reduced flushing of the marsh surface, leading to increased monthly mean salinity levels during spring months in the northern reach of the estuary (Knowles, 2000), a critical time in plant life cycles. We consider that the recent shift to more salt-tolerant plants apparent in the core records presented in Figure 2B may reflect the ecological response to upstream water management practices of the last century, although the depths vary from site to site. High-resolution chronologies for the historical period (post A.D. 1850) are needed to precisely determine the consequences of human activities in the Bay–Delta hydrological system, but are unfortunately difficult to obtain using ^{14}C dating (Stuiver et al., 1998). Figure 2B, therefore, shows isotopic and pollen records as a function of depth for the top sediments of the cores rather than age. The calculated sedimentation rates in Table 2 show a wide range because of this dating issue and

actual rates are not likely consistent between sites and over time.

Paleosalinity

Combined isotopic and pollen results support a pattern of vegetation change that reflects changing estuary paleosalinity over the last 3000 yr (Figs. 2 and 3). These changes are superimposed upon the more long-term trends of increased salinity due to sea-level rise and succession on the marsh. In particular, the periods 1600–1300, 1000–800, and 300–200 cal yr B.P. appear in the marsh records to be times when the estuary salinity was relatively high. In addition, the results indicate periods when the estuary paleosalinity was lower (higher fresh water inflow); including the period before 2000, 1300–1200, and ca. 200 cal yr B.P. This timing compares well with the record from Rush Ranch, where Byrne et al. (2001) found three major periods of increased salinity at 3000–2500 cal yr B.P., 1750–750 cal yr B.P., and A.D. 1930 to the present, and a period of reduced salinity from 750 cal yr B.P. to A.D. 1930 (the recent dates were augmented by the identification of non-native pollen in the Rush Ranch core).

There are some differences in the timing of events between the marsh cores. Given the area of the estuary included in this study, these differences may, in part, be a reflection of the response time needed for changes in fresh water inflow to affect the salinity at locations throughout the estuary. This may be particularly true for Rush Ranch, which is located further from the main channel of the Bay estuary than any of the other sites (Fig. 1). Site-specific factors such as elevation and distance from channels can also impact vegetation patterns at each coring site. And while evidence is difficult to find in these marsh records for significant tectonic activity, even relatively minor movements along the faults crossing the estuary may change the local elevation of a site enough to affect the frequency and duration of tidal flooding. Finally, the differences may, in part, reflect the errors inherent in the sampling process and the radiocarbon dating process.

Our results are in general agreement with the paleosalinity records derived from stable carbon and oxygen isotopes in estuarine sediment cores collected from sites in the Bay although with some differences (Ingram and DePaolo, 1993; Ingram et al., 1996a, 1996b) (Fig. 4). Late Holocene $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records from the Bay cores indicate several periods of higher salinity, lasting from decades to more than 100 yr, interspersed with periods of reduced Bay salinity and these changes may occur cyclically with periodicities of ca. 60, 90, 150, and 200 yr (Ingram and DePaolo, 1993; Ingram et al., 1996a,b). The chronologies from the Bay cores, however, differ from the marsh records, most likely because they were based on radiocarbon dating of carbonate shells from a variety of organisms (clams, mussels, oysters,

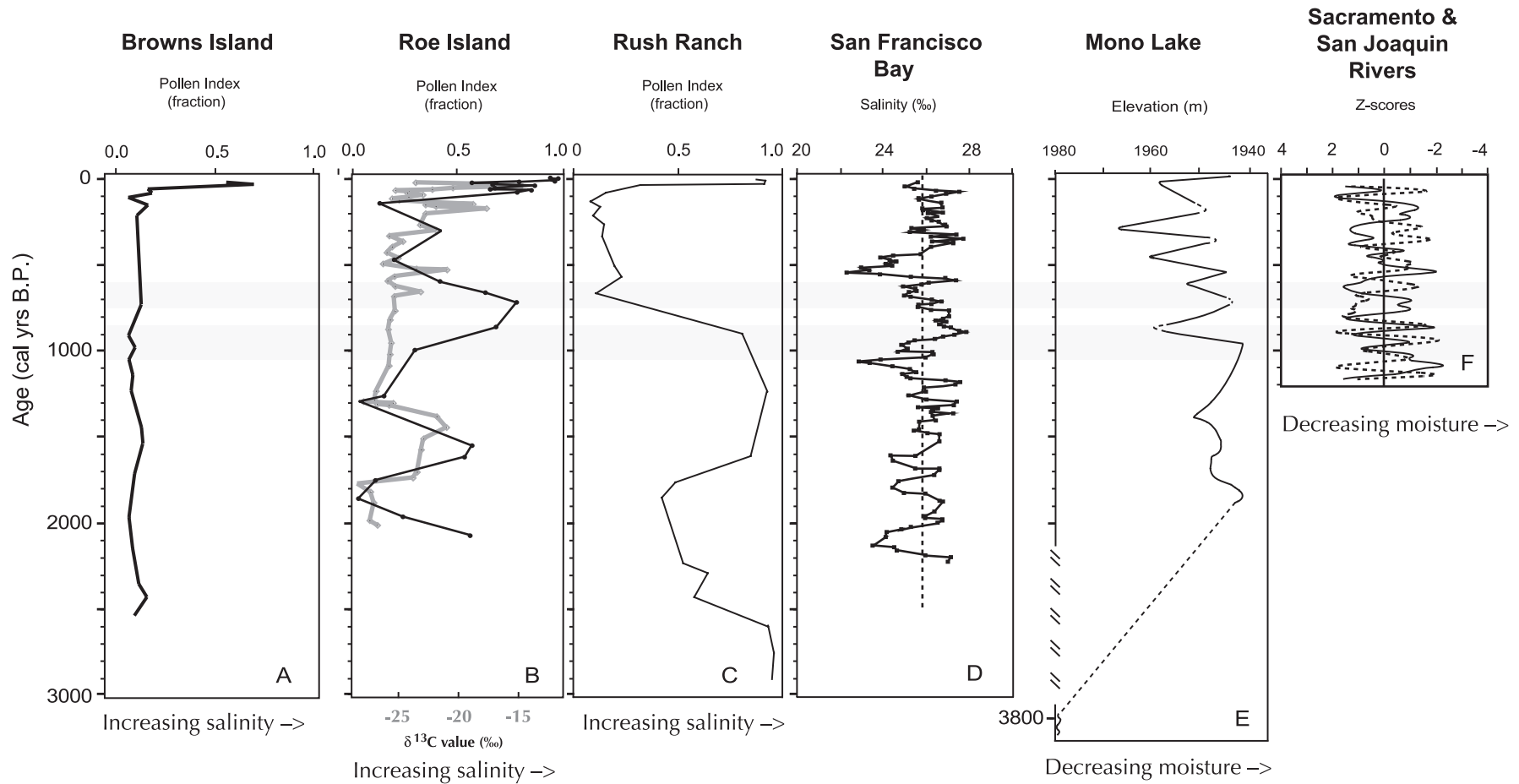


Fig. 4. Comparison of marsh sediment records with other paleoclimate records from the Bay and California. Vertical axis is time (cal yr B.P.). Periods of Mono lake droughts are indicated by shading (Stine, 1994). (A) Browns Island pollen index representing the Chenopodiaceae pollen as a fraction of the sum of Chenopodiaceae and Cyperaceae pollen (May, 1999). (B) Roe Island pollen index (May, 1999). The inferred salinity experienced at the site increases as the ratio approaches 1.0. (C) Rush Ranch (in Suisun Bay) pollen index, representing the Chenopodiaceae plus Poaceae as a fraction of the sum of Chenopodiaceae, Poaceae, Cyperaceae and Asteraceae (Byrne et al., 2001). The inferred salinity experienced at the site increases as the ratio approaches 1.0. (D) Oyster Point, San Francisco, paleosalinity (‰, three-point running mean) plotted for the last 2700 cal yr B.P. Salinity values derived from measured $\delta^{18}\text{O}$ values of carbonate shells. The dashed line is the estimated prediversion flow (Ingram et al., 1996b). (E) Fluctuations of Mono Lake shoreline elevation (Stine, 1990). Lower elevation for the shoreline indicates reduced water inflow to the lake. Note change in time scale. (F) Reconstructed river flows for the San Joaquin river (solid line) and the Sacramento river (dashed line) based on tree ring chronologies (Meko et al., 2003). Negative Z scores indicate reduced river flows.

and foraminifera) rather than terrestrial carbon. The errors inherent in radiocarbon dating of carbonates in coastal waters have been discussed elsewhere (Ingram, 1998; Ingram and Southon, 1996; Kennett et al., 1997; Stuiver and Braziunus, 1993) and involve ^{14}C turnover in the world's oceans as well as temporal and spatial variability. Thus, it is not surprising that inconsistencies in the timing of paleosalinity patterns exist between the Bay and tidal marsh cores.

Paleoclimate

Figure 4 presents a comparison of records from throughout the Bay estuary and includes sites outside the Bay, including lake elevation at Mono lake (Stine, 1990) and river flow reconstructions from the Sacramento and San Joaquin rivers. We include these records to show the regional trends in climate variability. Changes in estuarine salinity are primarily due to changing freshwater inflow through the Delta (Peterson et al., 1995), reflecting reduced precipitation experienced over much of the state. River flow reconstructions based on tree-ring records for the Sacramento and San Joaquin rivers going back 1200 yr (Meko et al., 2003) suggest that climate patterns over California have not been spatially uniform, and the Bay estuary appears to correspond better with changes occurring in the San Joaquin drainage. This appears to be supported by the correspondence between our Bay records, the reconstructed San Joaquin river flows (Meko et al., 2003), and the Mono lake record (Stine, 1990). Changing lake levels and submerged tree stumps at Mono Lake suggest at least two pronounced droughts that occurred on the east side of the Sierra Nevada at ca. 1050–850 and 750–600 cal yr B.P., both terminated by periods of high precipitation (Stine, 1994). In addition, tree ring data spanning the last 1000 yr from southern Sierra Nevada conifers indicate periods of anomalously dry conditions, the longest of which occurred from ca. 700 to 585 cal yr B.P. (Graumlich, 1993).

Conclusion

The stable carbon isotopic compositions of sedimentary organic matter in San Francisco Bay tidal marshes is used as a proxy for vegetation change on the marsh surfaces, which in turn change over time in response to changing estuarine salinity. The data suggest that paleosalinity of the San Francisco Bay estuary has been variable over the last 2000–3000 yr, corresponding to paleoclimate variability over a large region of California. During modern times, average precipitation in California has not been anomalously low, yet the salinity in the Bay estuary has been relatively high. This may reflect the impacts that human diversions of fresh water for agriculture and development (as well as upstream storage and timed releases) have had upon the San Francisco Bay estuary system.

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