Coupling of Sea Level Rise, Tidal Amplification, and Inundation

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ABSTRACT

With the global sea level rising, it is imperative to quantify how the dynamics of tidal estuaries and embayments will respond to increased depth and newly inundated perimeter regions. With increased depth comes a decrease in frictional effects in the basin interior and altered tidal amplification. Inundation due to higher sea level also causes an increase in planform area, tidal prism, and frictional effects in the newly inundated areas. To investigate the coupling between ocean forcing, tidal dynamics, and inundation, the authors employ a high-resolution hydrodynamic model of San Francisco Bay, California, comprising two basins with distinct tidal characteristics. Multiple shoreline scenarios are simulated, ranging from a leveed scenario, in which tidal flows are limited to present-day shorelines, to a simulation in which all topography is allowed to flood. Simulating increased mean sea level, while preserving original shorelines, produces additional tidal amplification. However, flooding of adjacent low-lying areas introduces frictional, intertidal regions that serve as energy sinks for the incident tidal wave. Net tidal amplification in most areas is predicted to be lower in the sea level rise scenarios. Tidal dynamics show a shift to a more progressive wave, dissipative environment with perimeter sloughs becoming major energy sinks. The standing wave southern reach of the bay couples more strongly back to the central portion of the bay, in contrast to the progressive wave northern reach of the bay. Generation of the M₄ overtide is also found to vary between scenarios and is a nonnegligible contributor to net changes in high water elevation.

1. Introduction

Among the many concerns related to recent and predicted climate change is the trend of rising sea levels. At global scales, studies such as Douglas (1997) show global sea level rising approximately 0.20 m in the past 100 yr, and predictions for sea level rise in the twenty-first-century range from 0.2 to 2.0 m (Parris et al. 2012). Adding to trends in the global-mean sea level, observations in some regions also show a pattern of increasing tidal amplitudes and increasing nontidal variations in sea surface height. The combined effects of sea level rise and potentially increasing tidal ranges will have far-reaching impacts on coastal inundation as many low-lying areas either become uninhabitable or require massive mitigation measures to fend off higher sea levels. Regional studies such as Grinsted et al.

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(2013) also point toward increases in inundation due to more frequent and more extreme weather events. As inundation is a consequence of peak sea level, not mean sea level, it is essential to consider both coastal ocean trends in mean sea level and how those trends will couple with local tidal dynamics to affect the peak sea surface height adjacent to areas in danger of flooding. At the same time, inundated areas add to the available tidal prism and the overall tidal energy dissipation, such that one must consider the whole system in order to accurately capture the coupling of tidal dynamics and inundation.

Nearshore regions are also influenced by management decisions, which in turn rely on predictions of flooding and sea level rise. Relevant management actions fall into two main categories. The first category, shoreline "hardening," describes the construction of hydrodynamic barriers such as concrete sea walls or levees. These projects may be motivated by flood risks, "reclamation" of shallows into dry land, or creation of ponds for salt harvesting. In many areas shoreline hardening is widespread and significantly alters the dynamics of the basin, such as in San Francisco Bay, California, where upward of 85% of historic marshlands have been filled

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or fundamentally altered (Collins and Grossinger 2004). Shoreline hardening decreases the tidal prism and often leads to greater tidal amplification. The second, generally opposite, category of shoreline modifications could be labeled shoreline "softening," but because it is often attempting to reverse the effects of earlier hardening projects, these actions may also be termed restorations. Typical restoration projects include breaching old levees or dredging new channels. Returning tidal action to these areas serves a number of purposes including reestablishing highly productive marsh ecosystems, improving water quality, and even mitigating flood risks. These projects often increase the area available to tidal action and introduce softer, natural shorelines and slough networks that are effective at dissipating tidal energy. Between the growing number of restoration projects and the potential for widespread sea level rise mitigation efforts, it will be important in the coming century to quantify the range of shoreline modifications and the effects those changes will have both on localized inundation and basinwide tidal dynamics.

a. Tidal amplification

Variations in tidal range within a basin come primarily from four physical processes: standing wave resonance from the reflection of the incident tidal wave, frictional effects, converging geometry (i.e., a landward decrease in the cross-sectional area), and inertial effects (van Rijn 2011). Resonance and converging shoreline geometry lead to an increase in tidal amplitude away from the open-ocean boundary of a basin, while diverging shorelines and friction lead to attenuation. Inertial effects are typically negligible and are ignored in most analyses that do not target shallow macrotidal systems.

Standing wave resonance is easily understood in terms of a simple prismatic channel, in which the tides can be described by the superposition of an incident wave and a reflected wave. Standing tidal waves occur when the incoming tidal wave is fully reflected, such as in a nonfrictional basin with a nondissipative landward boundary, and in this case the superposition of the two waves simplifies to

$$\eta = \frac{\eta_0}{2} \cos kx \cos \omega t, \qquad (1)$$

where η is the space- and time-varying free surface perturbation, *x* is the distance from the close end of the basin, *k* is the wavenumber, *t* is time, ω is the angular frequency of the tidal forcing, and η_0 is the tidal range at the closed end of the estuary. Given the length *L* from the closed, reflective landward boundary of a basin to the open, ocean-forced mouth, the amplification is simply $\alpha = \sec(kL)$. In systems where *L* is near a quarter wave node, the amplification approaches a resonant peak, such as the famous tides of the Bay of Fundy. The resonant period of a basin depends on the phase speed of the tidal wave; if the resonant period is altered toward the dominant period of the tidal forcing, one would expect that the net result would be an increase in tidal range. A fundamental parameter characterizing the degree to which a standing wave is present is the velocity phase lead ϕ , which we define here as the phase offset between peak flood velocity and peak high water for a specific tidal constituent. A progressive wave tide in which reflected energy is vanishingly small will have a phase lead approaching zero, while a standing wave system will see $\phi \approx 90^\circ$. The velocity phase lead provides a useful local estimate of basinwide tidal energy dynamics. Aside from diagnosing standing wave or progressive wave dynamics, ϕ is also relevant for residual scalar transport and sediment dynamics, because Stokes transport is greatest for $\phi = 0$ and negligible when $\phi = 90^{\circ}$.

The geometry of a basin can lead to amplification or attenuation through converging or diverging shorelines. van Rijn (2011) investigated the competing roles of convergence (both in depth and width), friction, and reflection. He found that in sufficiently long, deep, and converging estuaries, the amplifying effects dominate and tidal amplitude increases toward the head of the estuary. Shallow converging channels are dominated by friction, resulting in an attenuated tidal range landward of the mouth. In broad terms, he found that the reflected wave, if one exists, affects roughly the landward third of the basin, as the reflected wave is both dissipated by friction and attenuated by diverging shorelines as it travels seaward. In strongly converging channels, the phase lead of the peak flood velocity ahead of high water approaches 90°, independent of the presence of a closed landward boundary. Savenije et al. (2008) and van Rijn (2011) term this condition an apparent standing wave.

Cai et al. (2012) derived an analytic model of basin amplification applicable to basins with varying depth, convergence, friction, tidal forcing, and off-axis storage. The resulting expressions allow a classification of basins based on how the actual depth compares to the ideal depth (i.e., producing zero amplification) and the critical depth (i.e., producing maximum amplification). The model also includes prognostic equations for the phase lead and nondimensional amplification factor, as a function of basin geometry, tidal forcing, friction, and mean depth. The flexibility of the input parameters and wide range of behaviors that can be predicted make this model particularly relevant for sea level rise forecasts. In section 5a, we apply it to a portion of the study area and compare analytic and numerical predictions for the M_2 tide in order to understand the degree to which the analytic approach captures the necessary physics.

b. Inundation and tides

In regions with considerable inundatable area, the effects of inundation on tidal dynamics must also be considered. Higher sea surface heights allow the tides to access a greater tidal prism. The inundated areas are almost universally very shallow, and while the increased tidal prism may increase tidal velocities seaward of the inundated region, the shallow expanses have an overall dissipative effect on the tides. This additional dissipation tends to decrease reflection and mitigate some fraction of the sea level rise. While there is a well-established body of work on inundation resulting from storm surge, the literature on the energetics of inundation coupled with tides is relatively sparse. Depending on the characteristics of the newly wetted area, the amount of dissipation of tidal energy at the perimeter may decrease or increase. At one end of the continuum, one could imagine a basin with shear vertical walls at the original mean higher high water (MHHW) contour. As the sea surface rises, shallows that were originally intertidal become subtidal and less frictional. Overall, the perimeter becomes more reflective, leading to a greater tidal range. At the other end of the continuum, one can imagine that the region that was originally supertidal is instead flat and littered with drag-inducing features. In this case, the newly inundated areas are dissipative and tend to absorb the energy of the incoming tidal wave. Flows within the perimeter would shift toward a frictional regime, and flows in the interior of the basin would shift toward a progressive wave as incident tidal energy is absorbed in the intertidal areas.

Despite continual progress in analytic solutions to tidal propagation such as Lanzoni and Seminara (1998), Savenije et al. (2008), and van Rijn (2011), the complications of real world tidal basins limit the application of such models. Spatially varying friction and reflection, and geometries that do not fall cleanly into straight, exponentially converging, or steadily sloping beds, still frustrate analytic treatment and dictate the need for numerical approaches. Adding two-way coupling between inundation and tidal energetics, the problem is most thoroughly treated with numerical approaches. Recent work in tide inundation coupling includes Oey et al. (2007), who implemented a wetting and drying scheme in the Princeton Ocean Model (POM) that was then applied to modeling the dynamics of the wetting and drying on the extensive mudflats of Cook Inlet, Alaska. Their results showed up to a 20% increase in tidal range when wetting and drying were included, as well as a slowing of the tidal wave, reducing phase angles by up to 10%. The increase in tidal range with inundation appears to contradict expectations based on frictional dissipation in intertidal areas. The exact comparison in Oey et al. (2007), though, is between a case with wetting and drying allowed in the intertidal and a case where the would-be intertidal area is numerically "dredged" to become subtidal. Saramul and Ezer (2010), applying the same POM implementation to an idealized seamount, found that bottom stress and barotropic pressure gradients doubled when wetting and drying were allowed, further emphasizing the role of friction in inundation studies.

c. Present goals

We aim to investigate how sea level rise in the coastal ocean modifies the coupled tidal-inundation dynamics, in hopes of informing future mitigation and restoration efforts and anticipating their consequences. In an attempt to capture the complexities of a physical system, while maintaining broad applicability to other systems, San Francisco Bay, California, has been chosen as the domain for the numerical experiments. San Francisco Bay has moderate, mixed tides, representative of a wide area (Bromirski et al. 2003) and without particular anomalies that would make an analysis irrelevant for other basins. One advantageous feature of San Francisco Bay is its pair of dynamically distinct channels: a short, reflective, convergent channel to the south and a longer, progressive wave channel to the north leading to a dissipative inland river delta. With a single numerical experiment, we are thus able to see a wide variety of responses and interactions.

San Francisco Bay is also a prime example of the range of management actions that affect and are affected by inundation dynamics. Multiple large restoration projects will be returning previously nontidal salt ponds to tidal action, including a 61-km² project in south San Francisco Bay (SSFB) and a collection of smaller projects summing to a roughly similar area in the northern reach of San Francisco Bay. In addition to anticipated restoration efforts, sea level rise mitigation projects are also likely to alter significant reaches of shoreline in the next 50–100 yr, with two airports and numerous transportation corridors within reach of rising bay waters. An important question for planners is how far reaching the effects of a particular mitigation effort are. We wish to answer questions such as whether the hardening of a stretch of shoreline by additional levees will increase the inundation risk for neighboring soft shorelines. At larger spatial scales, we may ask whether hardening shorelines around one embayment alters the tidal signal in another embayment. To this end, multiple shoreline scenarios are modeled, with leveed reaches of shoreline



FIG. 1. Overview of model domain. Shaded depth contours are shown at -5- and -12-m NAVD88. The solid gray line outlines the present-day MHHW shoreline in SSFB and SPB. The solid black outlines the model domain including three false deltas. Dashed black line is the end-to-end thalweg, with the origin at the Golden Gate marked by a star and 10-km intervals marked by solid dots.

inserted into the model bathymetry to simulate shoreline hardening. Understanding the interplay between tidal dynamics, sea level rise, tidal marsh restoration, and the resulting inundation is essential for achieving the goals of these coastal engineering projects at the same time as predicting and mitigating inundation hazards.

2. Physical domain

San Francisco Bay has one of the longest continuous tidal records on the Pacific Ocean at 160 yr (Talke and Jay 2013), showing sea level trends of 0.22 m rise per century (Flick et al. 2003). Another recent analysis (Bromirski et al. 2011) has found a reversal in the trend of sea level off the coast of California since 1997, though they attribute this to the Pacific decadal oscillation and note that in due time the trend of increasing sea level is likely to return. A comprehensive evaluation of the projected change in ocean forcing for San Francisco Bay is detailed in Knowles (2010). Shifts in regional climate (as well as future management decisions) will undoubtedly affect river inputs as well. For the present purposes, though, we take freshwater forcing, nontidal sea level, and ocean tidal range as constants.

San Francisco Bay is a bifurcated, mesotidal estuary. Tides at the mouth are mixed diurnal and semidiurnal, with a great diurnal range at the mouth of 1.8 m (NOAA 2013). The mouth of San Francisco Bay is the Golden

Gate (star in Fig. 1), a 100-m-deep, constricted channel, connecting Central Bay to the Pacific Ocean. Alongchannel distances throughout this paper are referenced to the Golden Gate with negative distances denoting the southern transect and positive denoting the northern transect. The southern branch of the bay, typically referred to as SSFB, has little freshwater inflow, is roughly funnel shaped, and is characterized by a single channel 12-20 m deep and broad shoals tapering from 5m deep to intertidal. Tides in SSFB are close to a standing wave, with a velocity phase lead for the M₂ constituent in the channel of approximately $\phi_{M_2} \approx 75^\circ$ (i.e., peak flood velocity leads the peak sea surface elevation by 75°). The perimeter of SSFB, particularly to the south and southeast, is dominated by tidal sloughs and salt ponds. Ongoing restoration projects are returning many of the salt ponds to tidal action by levee breaches.

The northern branch of San Francisco Bay connects through two largely self-contained bays, consecutively San Pablo Bay (SPB) and Suisun Bay, before reaching the Sacramento–San Joaquin River Delta. A large fraction of the northern borders of both San Pablo Bay and Suisun Bay are tidal marshland joined to the respective bays via networks of tidal sloughs. A large number of wetland restoration projects, in various phases of planning or completion, are also targeted at the northern perimeter of San Pablo Bay.

3. Numerical model

To quantify the tidal processes and contributions to variation in high water across the domain, we employ the Stanford Unstructured Nonhydrostatic Terrain-Following Adaptive Navier-Stokes Simulator (SUNTANS) hydrodynamic model (Fringer et al. 2006; Wang et al. 2009) to simulate a range of sea level and shoreline configurations in San Francisco Bay. While estuarine circulation and other 3D processes undoubtedly will be altered by sea level rise, we have chosen to run the model in the less computationally expensive depth-averaged mode. Recent modeling studies have shown sea level rise-driven changes in the salinity intrusion length of 10%-20% of seasonal changes (Delta Science Council 2013; Chua 2011). Given that seasonal variation in amplification, presumably driven by seasonal river flows, is on the order of 0.05 m, we expect that changes in stratification will have minimal effects on inundation and the barotropic tidal response. The model solves the shallow water equations,

$$\frac{\partial u}{\partial t} + \mathbf{u} \cdot \nabla u - fv = -g \frac{\partial \eta}{\partial x} - \frac{1}{2h} C_D u ||\mathbf{u}||, \qquad (2)$$

$$\frac{\partial v}{\partial t} + \mathbf{u} \cdot \nabla v + fu = -g \frac{\partial \eta}{\partial y} - \frac{1}{2h} C_D v ||\mathbf{u}||, \text{ and } (3)$$

$$\frac{\partial \eta}{\partial t} = -\nabla \cdot (h\mathbf{u}), \qquad (4)$$

where u is eastward velocity, v is northward velocity, \mathbf{u} is the vector-valued horizontal velocity, g denotes gravitational acceleration, and $f = 2\Omega \sin \Phi$ is the Coriolis parameter with angular velocity of the earth Ω and latitude Φ . Elevation of the free surface is given by η , measured as a departure from the North American Vertical Datum of 1988 (NAVD88) geopotential surface, and h gives the total height of the water column. These equations are discretized on a prismatic, finite-volume grid comprising unstructured triangles in the horizontal. The drag coefficient C_D is derived from a bottom roughness z_0 that in turn is calculated from the water column height based on relationships from previous modeling efforts in San Francisco Bay (Gross 1997). At each time step, the bottom roughness for each cell is found by linear interpolation over the values in Table 1. Following MacWilliams et al. (2008), roughness throughout the false delta regions is set to a minimal 10^{-5} m. At each time step *n*, an effective drag coefficient $C_{D,i}^n$ is calculated from the edge-local bed roughness $z_{0,i}^n$ as

$$C_{D,j}^{n} = \left[\frac{\kappa}{\log\left(\frac{h_{j}^{n}}{2z_{0,j}^{n}}\right)}\right]^{2},$$
(5)

TABLE 1. Bottom roughness as a function of water column height [values taken from Gross (1997)].

<i>h</i> (m)	$z_{0}\left(\mathrm{m} ight)$
0.1	0.001 23
0.6	0.00229
2.0	0.001 16
6.5	0.000 25
8.5	0.00010

where h_j^n is the height of the water column on edge *j* at time step *n*, and $\kappa = 0.42$ is the von Kármán constant.

Wetting and drying in the model is handled by deactivating cells for which the water column height falls below a threshold height. We choose a threshold of 5 mm, as it is sufficiently small to avoid significant artificial storage, but large enough that the model is stable at a reasonable time step of 10 s.

The ocean boundary of the model domain is approximately 50 km beyond the Golden Gate, coinciding with a long-term tidal gauge at Point Reyes (see Fig. 1). The focus of the present study is SSFB and San Pablo Bay. Within these basins, the model domain extends up to the 3.5-m NAVD88 contour to accommodate tidal amplification and sea level rise. Upstream of San Pablo Bay and seaward of the Golden Gate, the model extends up to the present-day MHHW shoreline. Beyond Suisun Bay, the Sacramento-San Joaquin Delta is represented by a pair of hypsometry-matched false deltas, as is the slough and marsh network north of Suisun Bay. For each of the three false deltas, hypsometry (the relationship between planform area and free surface elevation) is extracted from a 10-m digital elevation model (DEM), excluding areas already accounted for in the original grid. A length for each false delta is determined by the along-channel length of the primary channel in each region. The width is determined by this length and the maximum area found in the hypsometry calculations. A regular triangular grid is constructed to match these dimensions, with a resolution of approximately 400 m. The hypsometry is then binned by depths such that each successive depth bin corresponds to an increase in the planform area equivalent to the area of one cell. These ordered depths are assigned to cells starting at the seaward end of the false delta, proceeding first along a strip of cells to the landward end, and then proceeding laterally to the next strip of cells until all cells have been assigned a depth. This simple approach ensures that a subtidal channel extends the length of the false delta (consistent with known river features), with a bed that slopes up in the landward direction and has a degree of lateral bathymetry variation. By matching hypsometry and length, the false deltas approximate the tidal response of the more complex physical channel network with a substantial reduction in the number of grid cells compared to a fully resolved delta.

For the purpose of this study, the portion of the domain at elevations between 0 and 3.5 m NAVD88 is the most relevant, as this is the intertidal range across the imposed sea level rise scenarios of 0-1.0 m. The grid resolution in the area between these contours is set to a uniform nominal length scale of 50 m (resolving the broad intertidal areas at finer resolution than this runs into practical computational limits). Additionally, five dynamically important channels outside this region are also given increased resolution: the Golden Gate (resolved at a scale of 125 m), Carquinez Strait and Suisun main channel (100 m), Suisun Cutoff (100 m), and New York Slough (200 m). In all other areas, the grid resolution is allowed to increase at a rate of 10%, up to a maximum grid scale of 3 km at the open-ocean boundary. The resulting grid has 937759 cells.

Bathymetry data are derived from a range of sources covering subtidal, intertidal, and supertidal areas up to the 3.5-m NAVD88 contour. The base elevation data source is a 10-m seamless topography-bathymetry product designed for inundation studies (Carignan et al. 2011). Bathymetry at 10-m resolution for the Sacramento-San Joaquin Delta is taken from Foxgrover et al. (2013). Given the impact of small levee and slough features on inundation and hydrodynamic connection, special care is taken to assemble up-to-date and high-resolution intertidal topography. This includes the preprocessed 2-m bathymetry from Foxgrover et al. (2011), as well as gridded bare earth lidar datasets from Foxgrover and Jaffe (2005), NCALM (2003), and NOAA (2012). Missing data in the lidar datasets in small regions are filled via interpolation from nearby lidar data, or, in cases where the lidar was missing data over a span greater than 10 m, data are filled in from Carignan et al. (2011).

Though the intertidal regions are resolved at 50 m, the length scale of many channels and levee features, essential for the inundation characteristics of the marshes, is 5–25 m. As in previous studies, such as Bates et al. (2003), we have found that simple averaging of the DEM along each edge was insufficient to resolve either channels or levees robustly. Given the importance of narrow channels and levees in quantifying inundation, and the difficulty in applying the method of Bates et al. (2003) to an orthogonal finite-volume grid, we instead have developed a method that calculates the overtopping elevation for each edge. The algorithm and comparisons between the simple averaging of bathymetry and this connectivity-preserving method are described in Holleman (2013).

Calibration of the model (see the appendix) has been performed with observed tides and winds s_{Obs} . Periodic tides are used for all subsequent analysis in order to

TABLE 2. Naming convention for numerical experiments: Obs denotes observed tides and M_2 denotes 12.42-h periodic tides with amplitude matched to spring range of observed tides. (The *N* denotes hardening in the northern reach, *S* denotes hardening in the southern reach, and NS indicates hardening in both the northern and southern reach.)

	Sea level	Hardened shorelines				
Tides	Rise (m)	None	SPB	SSFB	Both	
Obs	0.0	SObs	_	_		
M_2	0.0	<i>s</i> ₀	h_{N0}	h_{S0}	$h_{\rm NS0}$	
M_2	0.6	s ₆₀	h_{N60}	h_{S60}	$h_{\rm NS60}$	
M_2	1.0	s_{100}	h_{N100}	h_{S100}	$h_{\rm NS100}$	

avoid the need for spring-neap duration runs of each scenario and to allow the analysis to focus on the individual effects of a single tidal constituent. The ocean-free surface is forced with an M_2 period (12.42 h) sinusoidal signal with a peak-to-peak amplitude of 1.64 m. The amplitude was chosen to match the great diurnal range observed at the Golden Gate. The imposed M_2 amplitude, larger than the observed M_2 constituent, allows for a range of inundation similar to the combined tides. This avoids the complications of nonlinear interactions between constituents, but retains and resolves the interplay between tidal dynamics and inundation regions, which is not affected by these interactions, and allows the development of higher-frequency harmonics.

The numerical experiments cover three variations in sea level rise and four shoreline configurations, with a naming convention outlined in Table 2. The range of ocean boundary conditions comprises (i) present-day mean sea level, (ii) an increase of 0.6 m, and (iii) an increase of 1.0 m. These values were chosen to roughly bracket the middle of the range of predictions for conditions in 2100 (Parris et al. 2012). Multiple shoreline configurations are used to simulate the effects of mitigation efforts such as the construction of levees at presentday MHHW shorelines and how shoreline hardening in one portion of the domain affects tidal range in other portions of the domain. The first shoreline scenario, "soft" (s), does not include any explicit shoreline protection, only present-day topography and bathymetry. The completely hardened scenarios h_{NSx} limit flows in both San Pablo Bay and SSFB to present-day MHHW shorelines. Two additional scenarios represent shoreline hardening limited to either San Pablo Bay (h_N) or SSFB (h_S) . In all periodic cases, the model is allowed to spin up for 4 days before the data are extracted for a single M₂ period.

4. Energy flux and tidal phase analysis

Energy flux and tidal phasing provide fundamental information about the spatially variable dynamics in each basin. While the one-dimensional tidal propagation problem is well described by the velocity phase lead ϕ , in two dimensions the direction of wave propagation and the sense of land- or seaward can become ambiguous. For a more robust description, we switch to a comparison of depth-averaged tidal energy flux and an effective tidal phase lag derived from the energy flux. The east- and northward energy fluxes are defined as the pressure work done by the flow on an imaginary surface normal to the respective coordinate direction, averaged over the mean depth of the water column (Kundu and Cohen 2004):

$$F_i = \left\langle \int_{-h}^{0} p u_i \, dz \right\rangle,\tag{6}$$

where u_i is the velocity component in the *i*th coordinate direction (assumed constant in the vertical), $p = -\rho gz$ is the hydrostatic pressure due to fluid density ρ , *h* is the time-varying height of the water column, and angle brackets denote a tidal average. For a tidal constituent with frequency ω_c , vertical amplitude η_c , easting and northing velocity amplitudes U_c and V_c , respectively, and velocity phase shifts $\rho_{x,c}$ and $\rho_{y,c}$ relative to the free surface phase, the harmonic velocity and depth are given by

$$u_c = U_c \cos(\omega_c t + \phi_{x,c}), \qquad (7)$$

$$v_c = V_c \cos(\omega_c t + \phi_{y,c}), \quad \text{and}$$
 (8)

$$h_c = H + \eta_c \cos\omega_c t, \qquad (9)$$

where *H* is the mean depth. Substituting (7)-(9) into (6), the energy flux for constituent *c* is then

$$F_{x,c} = \frac{1}{2}\rho g U_c H \eta_c \cos\phi_{x,c} \quad \text{and} \tag{10}$$

$$F_{y,c} = \frac{1}{2}\rho g V_c H \eta_c \cos\phi_{y,c}.$$
 (11)

Consistent with the energy flux estimates, a measure of the tidal phase can then be determined as the inverse cosine of the ratio of the energy flux to the maximum possible flux for a fully progressive wave with the same H, U_c, V_c , and η_c :

$$\phi_c = \cos^{-1} \left[\frac{\sqrt{(U_c \cos\phi_{x,c})^2 + (V_c \cos\phi_{y,c})^2}}{\sqrt{U_c^2 + V_c^2}} \right].$$
 (12)

In the case of unidirectional tidal flow aligned with the x coordinate, (12) simplifies to $\phi_c = \phi_{x,c}$. However,

compared to considering the tides to be unidirectional along the principal axis, (12) is robust to amphidromes and rotary tides, where the principal axis is poorly defined.

Energy in higher harmonics is relatively small in the majority of the domain, and our initial analysis is focused on the M_2 constituent. The numerical experiments do predict a significant M_4 overtide, which is later considered in section 6. From each of the periodic scenarios in Table 2, the M_2 phase and amplitude of the sea surface height, eastward velocity, and northward velocity were extracted by a least squares approach over exactly one tidal period. Changes in M_2 energy flux between pairs of scenarios elucidate how the tides change in response to shoreline hardening and sea level rise and how this response differs between the two bays.

a. South San Francisco Bay

Figure 2a shows the M₂ energy flux and tidal phase for SSFB. The dominantly standing wave tidal dynamics are clear, with the majority of the embayment showing $\phi_{M_2} > 80^\circ$. The channelized portion of the bay to the southwest is slightly more progressive, but still close to a standing wave, especially in comparison to the channel in San Pablo Bay (discussed below). Though the tidal phase is generally close to standing throughout the bay, the energy flux magnitudes are still substantial due to the large tidal range.

Interestingly, in much of the bay the easternmost portion of the shoals shows slightly "overstanding" tides with a seaward energy flux. At the most eastward margins of the bay, the seaward-directed progressive component is sufficient to reduce ϕ_{M_2} below 80°. This seaward energy flux has been observed in SSFB (Lacy et al. 2014) and is reminiscent of the seaward residual transport in Li and O'Donnell (2005). The intratidal analytic model of Li and Valle-Levinson (1999), when evaluated with a bathymetry profile extracted from SSFB [detailed in Holleman (2013)], closely mimics the overstanding tides observed in the present model data. Qualitative agreement in ϕ between that analytic model and the present 2D model is a good indicator that a channel-shoal bathymetry profile and a reflective basin are sufficient to drive the overstanding wave in the shoals.

Having discussed the present-day M_2 dynamics, we now move to how these dynamics are altered with sea level rise and inundation. To approximately separate the effects of deepening from inundation, we consider first the changes due to sea level rise with hardened shorelines throughout the domain. The change in M_2 phase and energy flux between the h_{NS0} and h_{NS100} scenarios is shown in Fig. 2b. The choice of the 1.0-m scenarios is motivated by the characteristics of inundation in south



FIG. 2. SSFB: (a) ϕ_{M_2} and tidal energy flux for scenario $h_{\rm NS0}$. Incremental change in ϕ_{M_2} and tidal energy flux are shown for (b) $h_{\rm NS100} - h_{\rm NS0}$ and (c) $s_{100} - h_{\rm NS100}$.

San Francisco Bay, where a majority of the inundation occurs above a sea level rise of 0.6 m. The changes in phase in the bulk of the bay are minimal. The landward end of the bay has a convergent geometry and is nearly closed, such that, absent any dissipation from inundation, tidal energy has nowhere to go, and the landward energy

flux is constrained to be near zero. The southern half of the bay shows only small and scattered changes in the energy flux, while the northern half shows a distinct seaward shift in the energy flux. This shift is consistent with a deeper SSFB, which is less frictional and more reflective. Portions of the eastern shoals become more progressive, departing slightly from the bulk of the bay, but notably the change in energy flux is actually seaward, showing that the overstanding tidal phasing of the shoals is accentuated by the deeper sea level and greater tidal range in the far south end.

The effects of inundation on M₂ dynamics are shown in Fig. 2c, comparing scenarios $h_{\rm NS100}$ and s_{100} . SSFB is ringed by numerous tidal sloughs, connecting pond and slough networks to the main body of the bay. The greatest changes are at the mouths of sloughs, which function as gateways to the increased tidal prism when inundation is permitted. The sloughs are typically small, but when considered in the aggregate they are a significant sink of tidal energy in the M₂ band. The change in energy flux is everywhere landward, consistent with a basin transitioning toward a progressive wave. The eastern shoals actually show an increase in ϕ_{M_2} , toward a standing wave. In the scenarios with hardened shorelines, these shoals had a seaward energy flux, such that as the basin as a whole becomes more progressive, those areas that were originally overstanding (i.e., a seaward tidal energy flux) shift toward a standing wave. The flux changes of Fig. 2c are quite uniform across the width of the bay, especially compared to the baseline energy flux of Fig. 2a, which has significant lateral variation. This stems from the fact that in SSFB the majority of the inundation occurs in the southern portion of the bay. The primary effect of inundation for most of the bay is directly related to the amplitude of the reflected wave, rather than local or lateral dynamics.

b. San Pablo Bay

Figure 3a shows ϕ_{M_2} throughout the interior of San Pablo Bay, under present-day conditions (scenario h_{NS0}). San Pablo Bay has somewhat progressive tides along the main channel in the south and a partially standing wave across the shoals. Separated from the main channel by the broad shoals, the Petaluma River (to the northwest) and Sonoma Creek (to the north) connect to the northern shore. Along with the Napa River connecting to Carquinez Strait, these features are local sinks of tidal energy.

As with south San Francisco Bay, we first compare scenarios $h_{\rm NS0}$ and $h_{\rm NS100}$ in order to isolate the effects of a deeper basin interior, without significant change in inundation or tidal prism. Figure 3b shows the change in energy flux and phase between these two scenarios. The



FIG. 3. SPB: (a) ϕ_{M_2} and tidal energy flux for the h_{NS0} scenario. Incremental change in ϕ_{M_2} and tidal energy flux are shown for (b) $h_{NS100} - h_{NS0}$ and (c) $s_{100} - h_{NS100}$.

landward energy flux shifts from the channel to a proportionally greater flux in the shoals. The present-day mean depth of the off-channel areas of San Pablo Bay is quite shallow, making it highly frictional and a high impedance path for tidal propagation. A 1 m increase in mean sea level has a proportionally greater effect on the role of friction in the shoals, allowing a greater fraction of the tidal energy to propagate via the shoals. Another feature of the change in energy flux is the significant increase of energy leaving San Pablo Bay by way of Carquinez Strait to the east. This is likely due to a combination of less energy being lost in San Pablo Bay and greater dissipation in the false deltas beyond Carquinez Strait. In terms of the tidal phasing, the trend is clear that most of the bay shifts toward a progressive wave.

Figure 3c shows the incremental change in energy flux and phase between scenarios $h_{\rm NS100}$ and s_{100} . The bulk effect in San Pablo Bay is an increase in tidal energy entering the bay from the south and a decrease in energy leaving the bay in the east (note that Fig. 3c shows depth-averaged fluxes; in the south, the incoming flux is in twice the depth as the outgoing flux). Based on these changes in energy fluxes, it is clear that the inundation of the soft shorelines leads to greater dissipation, and the bay has become a greater sink of tidal energy. The hot spots of energy flux and progressive phase at the mouths of all three rivers show that the bulk of the newly inundating areas is not directly connected to the main body of the bay but are instead connected via river and slough features.

5. Tidal amplification and damping

So far we have examined only the M_2 wave, but to more concisely address the potential for inundation, we now turn to a direct measure of the peak sea surface height. The high water elevation includes the combined effects of the M_2 wave and its harmonics, as well as constant offsets due to, for example, Stokes transport increasing mean free surface setup. To separate changes in dynamics within the basin from changes in ocean forcing, we use the relative high water,

$$\Delta \eta(\mathbf{x}) = \max_{t} \eta(\mathbf{x}) - \max_{t} \eta_{\mathrm{BC}}, \qquad (13)$$

the height by which the high water level at a point in the domain exceeds or falls shy of a reference high water level at the coastal ocean boundary. Lateral variations in $\Delta \eta$ are typically small, allowing comparisons between scenarios simply along the central thalweg of the bay, as shown in Fig. 4. The highly reflective and converging SSFB is evidenced by the tidal amplification on the left side of the plot. At its most extreme, high water at the far south end of the bay exceeds coastal ocean high water by nearly 0.6 m in the s_0 scenario. Interestingly, the highest rate of amplification in the southern reach occurs in the first 20 km, which is still essentially part of Central Bay. The amplification rate slumps for the diverging reach between 20 and 40 km, with a slight uptick south of 40 km as the shorelines converge.

The more transmissive and dissipative northern reach of the bay sees mild amplification for the first 35 km, up



FIG. 4. Transect of relative high water for soft shoreline scenarios s_0 - s_{100} . Negative longitudinal distance indicates SSFB, positive is SPB and beyond.

to the transition from San Pablo Bay to Carquinez Strait. The seaward half of that stretch has particularly complex geometry and bathymetry, leading to greater variability over short length scales, up to the transition into San Pablo Bay proper, at 25 km from the Golden Gate. The greatest amplification occurs in the middle of San Pablo Bay. In a sense, San Pablo Bay can be considered a "leaky" reflective basin, particularly at present-day sea level. The landward outlet for tidal energy, Carquinez Strait, is relatively small; the ϕ_{M_2} within the strait itself dips below 30° (Fig. 3a), but this depression extends only partially into San Pablo Bay, approximately to the point of maximum amplification at 25 km from the Golden Gate. The northern shoals are far enough from the main channel to see additional, minor amplification relative to the main channel (not shown). This effect can be traced to the wide aspect ratio of San Pablo Bay and how the northern shoals behave somewhat like an offaxis standing wave basin.

The progression of tidal amplification from scenario s_0 to s_{100} in Fig. 4 demonstrates the combined, attenuating effect of inundation and sea level rise. Beyond Central Bay (from -20 to 15 km), the attenuation due to inundation more than offsets the amplification expected from a deeper basin. The locations at which the scenarios begin to diverge roughly correspond to where the inundatable regions occur, notably south of -40 km and north of 25 km. Figure 5 shows the incremental extent of inundation for each soft shoreline scenario. The most marked change in inundation in SSFB occurs when sea level rise approaches 1.0 m, compared to a relatively small change in the inundated area between the s_0 and s_{60} scenarios. Consistent with the inundation distribution in the south, the greatest change in amplification is between the s_{60} and s_{100} scenarios. San Pablo Bay has a more even distribution of inundated area, both in



FIG. 5. Extent of inundation across scenarios, progressing from least to most inundation.

terms of where these areas are located and at what rise in sea level they become inundated. There, the incremental difference in attenuation between s_0 and s_{60} is similar to the difference between s_{60} and s_{100} .

The comparisons between soft and hard shorelines approximately separate the effects of deepening from the effects of inundation, but also allow a comparison of local versus remote effects by selectively hardening only a subset of the shoreline. This demonstrates the dynamic interactions of the basins and at the same time informs practical management decisions regarding the degree to which mitigation efforts must be coordinated throughout a basin. The local and remote effects of shoreline hardening are quantified in Fig. 6, where the change in relative high water is shown for the four shoreline configurations. The baseline amplification of scenario s_0 has been subtracted out, as the changes are small relative to the baseline tidal amplification (i.e., the solid line of Fig. 4). When all shorelines are allowed to inundate, the model shows that a small portion of Central Bay is essentially unchanged, but everywhere else the tides are





FIG. 6. Longitudinal transect showing change in $\Delta \eta$ for s_{100} , h_{N100} , h_{S100} , and h_{NS100} , relative to the baseline $\Delta \eta$ of scenario s_0 . Negative longitudinal distance indicates SSFB, positive is SPB and beyond.

attenuated. With a maximum change in $\Delta \eta$ of approximately -0.13 m, the attenuation is notable, though small compared to baseline tidal amplification. Hardening only the shorelines of SSFB adds 0.06 m to the 0.60-m baseline amplification of the far southern reach, and also affects the high water level in Central Bay. Hardening and sea level rise both shift SSFB toward a more reflective, standing wave environment, and it is apparent in Fig. 6 that the reflected tidal wave couples back into Central Bay, which in turn alters the seaward boundary condition for the northern reach of the bay.

Modifications to San Pablo Bay have similar local effects as in SSFB. With soft shorelines, the broad inundatable regions of San Pablo Bay and its adjacent marshlands become a greater sink of tidal energy and tidal amplitudes decrease. Hardening these shorelines leads to a minor increase in tidal amplitudes. In contrast to SSFB, though, hardening the shorelines of San Pablo Bay has a negligible effect on SSFB, as the progressive wave dynamics of the northern reach reflect little energy back to Central Bay.

Analytic approach for converging basin

Analyzing the whole of San Francisco Bay through the analytic lens of converging estuary hydraulics such as Savenije et al. (2008), van Rijn (2011), and Cai et al. (2012) is frustrated by the various branching, diverging, and reconverging features. Nonetheless, analytic approaches aid in identifying the dominant factors controlling the tidal response and can quickly predict the general response of a system without detailed observation or involved numerical approaches. Although the complex geometry of much of San Francisco Bay makes a largescale application of analytic theory difficult, the central portion of SSFB has a smoothly convergent geometry. In this section, we apply the methods of Cai et al. (2012, hereafter CST) to this reach, between 30 and 55 km south of the Golden Gate (roughly the widest point of SSFB to the point at which the bay transitions to a broad slough), with a goal of understanding the predictive skill of the analytic model and its capacity to include inundation effects.

The formulations of CST include the following parameters in predicting the behavior of a basin: the length scale of the basin convergence a, the mean depth \overline{h} , the inverse of the Manning-Strickler friction coefficient $K = n^{-1}$, and the relative width of off-channel storage r_s . The ocean boundary condition is described by the tidal amplitude η_0 and angular frequency ω . These parameters are combined as

$$c_0 = \sqrt{\frac{g\overline{h}}{r_s}},\tag{14}$$

$$\zeta = \frac{\eta_0}{h},\tag{15}$$

$$\gamma = \frac{c_0}{\omega a}$$
, and (16)

$$\lambda = \frac{c_0}{c},\tag{17}$$

such that c_0 is the effective celerity including off-channel storage, ζ is a nondimensional tidal amplitude, γ is an estuary shape factor relating convergence and tidal wavelength, and λ is the ratio of the frictionless celerity to the frictional celerity. The tidal response of the system is described in terms of the nondimensional numbers:

$$\delta = \frac{1}{\eta} \frac{d\eta}{dx} \frac{c_0}{\omega},\tag{18}$$

$$\chi = \frac{r_s g c_0 \zeta}{K^2 \omega \bar{h}^{4/3} [1 - (4\zeta/3)^2]}, \text{ and } (19)$$

$$\mu = \frac{1}{r_s} \frac{v\bar{h}}{\eta c_0},\tag{20}$$

where δ is the amplification factor, χ is the friction number, v is the velocity scale, μ is the velocity number, and additionally $\epsilon = \pi/2 - \phi$ describes the velocity phase lead. CST then derives the system of equations:

2

1

$$\mu = \frac{\cos\epsilon}{\gamma - \delta},\tag{21}$$

$$\tan\epsilon = \frac{\lambda}{\gamma - \delta},\tag{22}$$

TABLE 3. Application of method of Cai to numerical scenarios.

	\overline{h}	n_0	а		Model		Analytic	
Scenario	(m)	(m)	(km)	r_s	ϕ	δ	ϕ	δ
$h_{\rm NS0}$	3.88	1.19	11.5	1.06	83.1°	0.237	79.7°	0.165
$h_{\rm NS100}$	4.81	1.21	12.1	1.05	85.2°	0.273	84.7°	0.240
s_{100}	4.74	1.16	12.2	1.33	79.4°	0.236	80.2°	0.214

$$\lambda^2 = 1 - \delta(\gamma - \delta), \text{ and}$$
 (23)

$$\delta = \frac{\gamma}{2} - \frac{4}{9\pi} \chi \frac{\mu}{\lambda} - \frac{1}{3} \chi \mu^2, \qquad (24)$$

the solution to which is a prediction of the tidal response. Application of CST to real bathymetry and shorelines, even with idealized tidal forcing, leaves considerable room for interpretation. While CST focuses on channels with a constant depth and converging width, the basis for the exponential convergence is in terms of the crosssectional area, such as in Savenije (1992). In the present case, we have found that the cross-sectional area demonstrates a significantly smoother trend and bases the length scale a on this quantity rather than the width. In all cases, K was taken to be 36.9, based on the range of \overline{h} and corresponding roughness values in Table 1. We define the in-channel region as a contiguous set of cells on either side of the thalweg with a mean depth of at least 1.0 m and a tidal amplitude of at least 0.2 m. The across-channel distance between the bounds of this region then define the channel width B (as a function of the along-channel distance), and similarly \overline{A} is defined as the cross-sectional area between these bounds. The storage ratio r_s is calculated as the ratio of the planform area of all tidally active cells (tidal amplitude greater than 0.2 m) to the planform area of all in-channel cells. Both the shallow intertidal margins of the main basin and all slough or pond features off the main channel are counted as off-axis storage. The open boundary tidal amplitude η_0 is taken from the model at a single cell along the thalweg. The mean depth at each cross section $\overline{h}(x)$ is the equivalent depth of a rectangular channel with area A and width \overline{B} . While CST assumes a constant depth, SSFB has a trend of decreasing depth toward the head of the estuary, and \overline{h} is taken as the mean over all cross sections.

The input parameters for each scenario and a comparison of phase and amplification between the numerical and analytic models are shown in Table 3. The range of η_0 is fairly small, showing that, according to the model, dynamics seaward of this section account for about 0.05 m of change in tidal amplitude. The greatest differences in the inputs parameters are the variation of \overline{h} with sea level and the variation of r_s with inundated



FIG. 7. Comparison of amplification factor δ in scenario $h_{\rm NS100}$, as predicted by the model and by the analytic approach of Cai et al. (2012). The longitudinal transect is limited to the convergent portion of SSFB.

area. All scenarios fall within the amplified classification of CST, meaning that the net amplification is positive, and incremental increases in the depth would result in additional amplification. While the velocity phase lead is predicted well by CST, the amplification factor is uniformly underpredicted (e.g., Fig. 7). The results from the analytic model reinforce the idea of competing effects of deepening and inundation, where an increase in \overline{h} due to the sea level rise is partially offset by an increase in r_s due to inundation. In terms of the amplification, these competing effects are nearly in balance in the numerical model, while the analytic approach is more sensitive to deepening than inundation, suggesting that inundation is not fully captured by considering storage alone.

6. Overtides

The total change in high water between $h_{\rm NS100}$ and s_{100} , at -60 km from the Golden Gate, is approximately 0.16 m (Fig. 6), but the M₂ amplitude explains only about 0.07 m of this difference. While the ocean boundary is forced only with an M2 tide, local generation of overtides leads to nonnegligible M₄ amplitudes within the domain, shown in Figs. 8a-c. Previous analysis of the nonlinearities in the shallow water equations (Parker 1991) has shown that M_4 overtides are predominantly generated by the depth dependence of the friction term, depth dependence in continuity, and the nonlinear advection term. The depth-dependent generation mechanisms are likely significant throughout much of San Francisco Bay, given the O(1 m) tides and O(2 m)depths prevalent in shoals throughout the domain. Though the mean M₄ amplitude is small (up to about 0.1 m), the differences across scenarios of the M₄ amplitude is of the same order as the differences in M_2



FIG. 8. M_4 amplitude for scenarios (a) h_{NS0} , (b) h_{NS100} , and (c) s_{100} . M_4 generation in 10^6 W, integrated by region, for scenarios (d) h_{NS0} , (e) h_{NS100} , and (f) s_{100} . Arrows in (d)–(f) show the direction of M_4 energy flux between regions.

amplitudes. In addition to varying amplitudes, the distribution of M_4 generation and the resulting phase relationships between the M_2 and M_4 vary greatly between scenarios.

The panels of Figs. 8d–f show that in all cases Central Bay is a significant source of M₄, but areas in the north and south may be sources or sinks of M₄ depending on the scenario. In all cases, M₄ appears to be generated in shallow, off-axis portions of the domain, and in most cases propagates seaward. Variation in M₄ generation appears to be driven by three factors: change in mean depth (i.e., h_{NS0} vs h_{NS100} or s_{100}), change in M₂ amplitude from which M₄ can be extracted, and local dissipation of the M₄. The M₄ dynamics are further complicated by the shorter wavelength that allows for standing wave nodes to exist within the basins, such as in h_{NS0} at the widest point of SSFB and in the middle of San Pablo Bay.

Taking into account the amplitude as well as the phase relative to the M_2 phase, we estimate that in $h_{\rm NS100}$ the M_4 adds roughly 0.04 m to high water in most of SSFB, compared to s_{100} in which M_4 actually decreases high water by up to 0.04 m. Of the original $h_{\rm NS100} - s_{100}$ difference of 0.16 m (at -60 km), the combined M_2/M_4 wave then accounts for roughly 0.15 m.

7. Discussion

Within a particular estuary or bay, the dominant factors controlling the tidal and inundation response to sea level rise include geometric factors like aspect ratio, the baseline phasing of the tidal wave, and the spatial distribution of inundated areas.

The aspect ratio determines the relative importance of longitudinal versus lateral variation. In the longer, high

aspect ratio SSFB, changes due to sea level rise were relatively consistent across lateral transects, and lateral dynamics appeared secondary. In contrast, San Pablo Bay, with a round, low aspect ratio footprint, showed significant shifts of tidal propagation from the channel to shoals.

Tidal phasing is important both in terms of local tidal amplification and how much the tides in one part of a basin feedback to other parts of the system. A standing wave system such as SSFB appears more sensitive to sea level rise, in both the case of deepening only and deepening with inundation. Additionally, standing wave systems tend to have greater tidal range such that even small changes in phasing or dissipation lead to large changes in energy flux and net amplification.

The quantity and relative location of inundatable areas also affects the tidal response. Greater expanses of inundatable areas relative to the subtidal area lead to greater attenuation of the incident tidal wave. The location of inundatable areas, along with the tidal phasing within a bay, affects the spatial extent of the attenuation due to inundation. In a purely progressive wave system, these effects are limited areas landward of the inundation/ attenuation. In a reflective, standing wave system, though, inundation even at the head of the estuary can attenuate the tidal range throughout the bay and even in adjacent tidal basins.

In addition to the incoming tidal wave constituents, overtides generated within a basin may add to or subtract from the high water elevation and appear to be very sensitive to shoreline conditions and incident tidal wave amplitudes. Depth-dependent M_4 generation mechanisms are of particular interest in sea level rise scenarios as the change in mean sea level can drastically change overtides in shallow basins. Large tidal ranges and shallow depths at the head of an estuary can generate seaward-propagating M_4 combined with a landward-propagating M_2 , along with the potential for M_4 resonance, modulation of overtides by sea level rise is a nonobvious but important aspect of predicting peak sea level within tidal basins.

The net physical response to coastal sea level rise clearly depends on a broad set of factors. We have considered only the M_2 forcing, but diurnal tides and interactions between diurnal and semidiurnal tides are likely significant. The long wavelength of diurnal tides leads to phasing closer to a standing wave, though net amplification is typically smaller at longer wavelengths (e.g., the analysis of section 5a when applied to the s_0 scenario yields 24% less amplification when the tidal period is doubled). Similarly, we expect that the dissipative effects of inundation are also less important for diurnal tides.

Perhaps the largest uncertainty in predicting what will happen in a particular estuary is the unknown evolution of morphology, whether by natural or managed actions. Understanding how basins respond to sea level rise when morphology is kept static is the first step toward understanding what natural changes are likely to occur and what management decisions may be deemed necessary.

8. Conclusions

Utilizing numerical experiments with a variable coastal sea level rise and varying shoreline configurations, we approximately separated the effects of deepening from inundation. Comparisons of phase and energy flux of the M_2 tidal wave show that deepening decreases the influence of friction, while inundation adds considerable dissipation in the perimeter areas.

Deepening allows additional tidal amplification [consistent with an amplified estuary in the parlance of Cai et al. (2012)], which was observed in all hard shoreline cases with sea level rise. The long, convergent southern arm becomes more reflective when deepened, while the shorter, transmissive northern arm shows the landward energy flux shifting from the channel to the shoals.

In both branches of the bay, inundation introduced large energy sinks at the bay margins. Most inundation occurred off perimeter sloughs and rivers, causing the most drastic changes in tidal phasing and tidal prism at points where these features join the larger bays. Energy sinks in newly inundated regions caused a progressive shift in tidal phasing, a decrease in tidal amplification, and an increase in the landward tidal energy flux.

In the case of SSFB, local changes in the shoreline alter both the tidal range within the basin and also the magnitude of the reflected wave. The reflected wave subsequently affects tidal range in other parts of the domain. In contrast, local changes in the shoreline of San Pablo Bay have limited effects on tidal range downstream of San Pablo Bay, because a smaller fraction of tidal energy is reflected.

A one-dimensional analytic model has been applied to a reach of SSFB, with moderate success. Predictions of amplification from this approach qualitatively agree with model output, though the hydrodynamic model shows a much greater effect of inundation than is captured in the analytic model. Other parts of the domain are likely too irregular and two-dimensional to be reasonably treated with a one-dimensional analytic model.

While the M_2 amplitude is much larger than the amplitudes of overtides, the variation in M_4 amplitude across numerical experiments is comparable to the variation in M_2 amplitude. Together with variation in the phase of the M_4 wave relative to the M_2 , we conclude

TABLE A1. Velocity and stage comparisons, forced with observed tides and winds. The *r* denotes the Pearson correlation coefficient. Lags are computed as the time offset that maximizes the correlation coefficient. The rms ratio is the ratio of model rms amplitude to observed rms amplitude. Bias is not reported at -40 km due to the lack of a reliable vertical datum.

	r	Bias	Lag (s)	Rms ratio
η at $-40 \mathrm{km}$	0.997	n/a	92	1.009
u at $-40 km$	0.983	$0.02{ m ms}^{-1}$	-151	0.985
η at 40 km	0.996	$-0.076{ m m}$	-146	1.016
u at 53 km	0.948	$-0.06{ m ms}^{-1}$	-1517	0.826

that overtides are an important component of the variation in high water. Depth-dependent nonlinearities in the shallow water equations are the most likely M_4 sources, consistent with extensive shoals in which the depth is of the same order as the tidal range.

Overall, the coupling between sea level rise, tidal amplification, and inundation is important and must be taken into account for accurate assessment of future restoration and mitigation questions. In many estuaries and bays, rising sea level in the coastal ocean will lead to newly inundated areas. To a degree this inundation can mitigate sea level rise by decreasing tidal amplification within the basins. Reinforcing and hardening impacted shorelines can increase flood risks in adjacent areas, and in highly reflective basins the effects can be far reaching. Restoration of tidal marshland and construction of new low-lying tidal areas offer significant protection from rising tides by dissipating incident tidal energy, and these benefits may extend well beyond the areas directly sheltered by marshland.

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APPENDIX

Model Validation

The tidal boundary condition is calibrated to match phase and amplitude of the sea surface height at the Golden Gate over the period from 24 February to 15 March 2009, by scaling measured tidal amplitude by 0.931 and adding a 120-s lag. The model has been validated against the observed tidal stage at two locations and depth-averaged velocity at two locations, over the period from 26 February to 9 March 2009 (with the exception of the velocity validation in SSFB, for which



FIG. A1. Observed and predicted sea surface height and alongchannel velocity. (a) Sea surface height at San Mateo Bridge (SSFB), relative to local MSL; (b) depth-averaged along-channel velocity at San Mateo Bridge; (c) sea surface height at Mare Island (SPB); and (d) depth-averaged velocity at Martinez (Carquinez Strait).

observations are truncated at 6 March 2009). Model forcing for the validation run was taken from the observed coastal ocean sea level as measured at Point Reyes and observed winds from Point Reyes, Port Chicago, Alameda, Redwood City, Richmond, and Union City. River flows were included for the San Joaquin and Sacramento Rivers, where the net delta outflow index (California Department of Water Resources 2011) was apportioned 25% and 75%, respectively. Table A1 summarizes the comparison between observations and model predictions.

Figure A1a shows time series comparisons in SSFB near the 40-km mark in Fig. 1 and laterally situated at the eastern edge of the channel at the foot of the slope leading into the shoal. A storm system passed through between 2 and 4 March 2009. Uncertainty in the distribution of wind stress is the likely cause of the trend of overpredicted sea surface height during this period. Depth-averaged currents at the same location are shown in Fig. A1b.

Sea surface height in San Pablo Bay is validated against observations at Mare Island, immediately west of the mouth of the Napa River (Fig. A1c). Long-term measurements of velocity in San Pablo Bay during the validation period were not available. Velocity measurements at the other end of Carquinez Strait are available for a site near the southern shore (near 53 km along the thalweg shown in Fig. 1). Comparison at this location is shown in Fig. A1d. We note that this site is beyond the intended study area, and validation here is adversely affected by proximity to the false deltas, decreased grid resolution outside the study area, and the highly energetic and spatially variable flows in this constricted tidal strait. Nonetheless, velocity phase and temporal patterns of variation in current magnitude are reasonably captured by the model.

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