Statistical analysis of vertical and alongshore temperature variability in the subtidal, diurnal, and semidiurnal frequency bands along the central California inner shelf

Jamie H. MacMahan a, Falk Feddersen* b, Thomas M. Freismuth a, Matt K. Gough a, and Michael Kovatch b

a Department of Oceanography, Naval Postgraduate School, Monterey, CA, 93940, USA
b Scripps Institution of Oceanography, University of California, San Diego, La Jolla CA 92093, USA

* Corresponding author
Abstract

Inner-shelf temperature $T$ over 50 km alongshore was observed with 20 moorings deployed in 9-16 m depth with high thermistor vertical density during Sept-Oct 2017 in central California, USA along a region with relatively straight shorelines interrupted by coastal headlands (Pt. Purisima and Pt. Sal). Temperature spectra had peaks in the subtidal (ST), diurnal (DU), and semidiurnal (SD) bands. The headlands influence temperature statistics in all bands. In each band, the vertical $T$ structure is given by the first EOF. The ST and DU first EOF are largely consistent across all moorings. The ST first EOF is mostly barotropic. The associated 1-1.5 °C variability is related to three northward propagating warm plumes with varied excursion distance. The ST stratification is vertically uniform. The DU first EOF is quasi-barotropic with some baroclinicity, and is not surface-intensified. The associated 0.2-0.4 °C variability was not modulated by ST stratification. North of Pt. Sal, the DU variability coherently propagates southward at nearly 2 m/s over 20 km, unrelated to the DU-winds. The SD first EOF shape varies significantly across moorings, from linear mode-1 baroclinic to bottom-intensified. The associated variability was incoherent, in contrast to locations farther offshore, was modulated by ST stratification, and decorrelated rapidly alongshore. The SD depth-averaged energy varies strongly with latitude and is enhanced at headlands, particularly Pt. Sal. Strong and weak SD depth-averaged energy are related to bottom-intensified and linear mode-1 vertical structure. This vertical and alongshore inner-shelf temperature variability is likely important for larval transport.
Plain Language Abstract

Twenty moorings with a high vertical density of thermistors were deployed in 9-16 m water depths along 50 km of the central California inner shelf to measure the temporal, vertical and alongcoast inner-shelf temperature variability. The coastline has regions of relatively straight beach shorelines interrupted by coastal headlands. The inner shelf is geographically defined as seaward of the surf zone, where depth-limited wave breaking occurs, out to about 20 m water depth. Temperature had significant variability in the subtidal (ST, 33hrs < time scales), diurnal (DU, 16 hrs < time scales < 33hrs ) and semidiurnal (SD, 10 hrs < time scales < 16 hrs) time scales motivating the evaluation of (co-)variability for each time scale. In each time scale, the vertical temperature structure at each mooring is characterized by the first empirical orthogonal function describing a high fraction of variance over the vertical. For each time scale, the temperature statistics differ spatially, and all are influenced by the presence of the headlands. The inner-shelf water temperature variability has important implications in the distribution, retention, and settlement of marine biota, the delivery of nutrients, and can be a measure of cross-shore exchange.

3 statements

ST temperature variability (1-1.5 °C ) is largely depth uniform and consistent across all moorings related to warm plumes
DU temperature variability (0.2-0.4 °C ) is quasi-barotropic with some baroclinicity, is not surface-intensified, and propagates southward
SD temperature variability (0.2-0.5 °C ) varies significantly across moorings, from linear mode-1 baroclinic to bottom-intensified
Keywords: temperature, variability, inner shelf, subtidal, diurnal, semidiurnal, internal wave, sea breeze

1. Introduction

Inner-shelf water temperatures can significantly vary spatially and temporally (e.g., Boehm et al., 2004; Tapia et al., 2014; Aristizabal et al., 2016, 2017). The inner shelf is geographically defined as seaward of the surf zone, where depth-limited wave breaking occurs, out to about 20 m water depth. The inner-shelf water temperature variability has important implications in the distribution, retention, and settlement of marine biota (O'Connor et al., 2007; Tapia et al., 2014), the delivery of nutrients (McPhee-Shaw et al., 2007; Lucas et al., 2011), and can be a measure of cross-shore exchange (Hally-Rosendahl et al., 2014).

Studies of inner-shelf temperature variability can be divided into two general categories: (1) heat budgets and (2) statistical analyses. A heat budget accounts for the total heat change of a control volume due to cross-shore, alongshore, and surface heat fluxes. Heat budgets have identified how processes such as upwelling, internal waves, and sea breeze driven transport contribute to inner-shelf water temperature variability at subtidal (ST, $f < 33^{-1}$ cph; Austin, 1999; Fewings and Lentz, 2011), diurnal (DU, $33^{-1} < f < 16^{-1}$ cph; Suanda et al., 2011; Herdman et al., 2015; Molina et al., 2014), and semidiurnal (SD, $16^{-1} < f < 10^{-1}$ cph) timescales. Estimates of cross- and alongshore heat fluxes require co-located measures of water temperature and currents over the vertical as well as adequate alongshore spacing to identify advective heat flux divergences (e.g., Austin, 1999), which can develop from coastal inhomogeneities, such as coastal topography or kelp forests (Suanda et al., 2011). Thus, heat budget studies are impractical over long stretches of inhomogeneous coastlines.
Statistical analysis of the temporal, vertical and alongcoast temperature (co-) variability can link magnitudes and scales of variability to processes and the effects of coastal heterogeneity. In addition, the spatial variability can be explored across different frequency bands (e.g. ST, DU, or SD) that differ in forcing and scale. For example, Tapia et al. (2014) found that, along 800 km of the central Chilean coast, nearshore ST-band temperature variability was related to the alongshore wind stress (i.e., upwelling and downwelling), and to coastal headlands that modified the local variability. On the central CA coast, ST-band temperature variability is primarily associated with northward propagating warm, buoyant (density changes $\Delta \rho$ of 0.1-0.8 kgm$^{-3}$) plumes originating from the Santa Barbara Channel (SBC), associated with wind relaxation events (Melton et al., 2009; Washburn et al., 2011). These warm buoyant plumes typically propagate to Pt. Sal, CA, but can extend even farther up-coast depending on the duration of the wind relaxation (Melton et al., 2009; Washburn et al., 2011; Suanda et al., 2016). Washburn et al. (2011) observed plume temperature gradients as large as 4 $^\circ$C with propagation speeds between 15-30 km day$^{-1}$. The warm plumes are generally attached to the coast, and tend to occupy the entire water column on the inner shelf (Washburn et al., 2011; Suanda et al., 2016). The changes in stratification induced by these warm plumes can modulate DU- and SD-temperature fluctuations (Aristizabal et al. 2017).

Inner-shelf DU-temperature variability can be related to a number of concomitant processes, such direct wind forcing (e.g., Kaplan et al., 2003), diurnal heating (e.g. Molina et al., 2014), and resonant forcing (e.g. Nam and Send, 2013). In the northern and southern portions of the SBC, near-bottom DU temperature and DU winds were coherent with no phase lag, suggesting direct wind forcing without any alongshore propagation (Aristizabal et al., 2016). Diurnal winds can also induce local coastal upwelling along coastal headlands (e.g., Woodson et al., 2007;
Suanda et al., 2011) or in coastal embayments (Walter et al., 2017). Diurnal winds can also drive
baroclinic modal (internal) oscillations that can be resonant when subtidal shear reduces the
effective Coriolis parameter (Lerczak et al., 2001; Nam and Send, 2013; Kumar et al., 2016) or
when the inertial and diurnal periods align (e.g. Lucas et al., 2014). Diurnal internal motions have
amplitudes that increase with larger background stratification and can be coherent over
approximately 50 km (Cudaback and McPhee-Shaw, 2009). At some SBC near-bottom locations,
DU-temperature variability can be linked to the barotropic tide (Aristizabal et al., 2016).

Inner-shelf SD-temperature variability tends to be influenced by internal waves (e.g.,
Pineda and Lopez, 2002). SD-band near-bed temperature variability was elevated on the mainland
(north) SBC channel and decreased north of Pt. Conception, CA (Aristizabal et al., 2016). This
SD-band variability was incoherent alongshore (Aristizabal et al., 2016) and presumably
incoherent with the barotropic tide. However, at locations near Santa Cruz island, SD-band near-
bed temperature variability was coherent due to the proximity to an SD internal tide generation
region (Aristizabal et al., 2016). Similarly, in a 6-km wide Chilean bay in 20-m depth, the SD-
temperature variability was coherent across the bay and with the barotropic tide (Bonicelli et al.,
2014), also likely due to proximity to a submarine canyon as a source region. Just offshore of Pt.
Sal, CA during summer 2015, Colosi et al. (2018) observed nonlinear internal waves with
significant energy in the SD-band from 50 m to 20 m water depth. In 50-m depth, the SD internal
tide was relatively alongshore uniform north and south of Pt. Sal and was approximately ½
coherent with the barotropic tide (Kumar et al., 2019). Modeling suggests the SD internal tide is
generated at least a few 100 km offshore (Kumar et al., 2019).

Previous statistical temperature studies that described inner-shelf temperature variability
over long-durations (O(10yrs)) were limited in the number of moorings or the total thermistors
resulting in poor vertical or alongshore spatial resolution. For example, Tapia et al. (2014) observed alongshore temperature variability with a single thermistor at 15 locations spanning ~600 km, with typical spacing of 50 km. Previous Santa Maria Basin (SMB) temperature observations also spanned multiple years, but were limited to a few locations on the south side of the topographic points, resulting in an alongshore mooring spacing of >17 km, and did not extend north of Pt. Sal (Cudaback and McPhee-Shaw, 2009; Melton et al. 2009; Washburn et al. 2011; Aristizabal et al. 2016, 2017). Furthermore, the 15-m depth temperature moorings were composed of three thermistors sensors located near the bottom, middle, and surface resulting in relatively coarse (~5 m) vertical resolution. Some analyses used only the near-bed thermistor. Thus, alongshore and vertical variation in inner-shelf temperature variability near alongshore heterogenous coastlines (such as Pt. Sal, CA) across different frequency bands is not known, as well as their linkage to processes such the winds and onshore propagating SD internal tide.

Here, a detailed statistical analysis is performed to describe the temporal, vertical, and spatial patterns of inner shelf (9 < h < 16 m, where h is water depth) temperature variability over 50 km alongshore in the heterogeneous Santa Maria Basin (SMB) region north and south of Pt. Sal, CA (Figure 1). The 20 temperature moorings have good alongshore resolution (~2 km separations) and high vertical resolution (0.5 m spacing) which allows for statistical description of variability in the ST, DU, and SD bands and linkage to processes driving the variability. The observations and the EOF-based methodology are given in Section 2. The vertical, alongshore, and temporal temperature variability in the ST, DU, and SD bands are described in Section 3. The variability in each band is contextualized in the Discussion (Section 4), and Section 5 summarizes the results.
2. Methods

2.1 Study Region

The study site is within the Santa Maria Basin (SMB) along the central California coast (Figure 1a) with coastline characterized by a headland-and-straight coastline topography with three rocky headlands (or topographic points): Pt. Sal (34.9N), Purisima Pt. (34.76N), and Pt. Arguello (34.58N). Note that Pt. San Luis (35.26N) to north of experimental site also influences the observations. The 10 – 20 km of coastlines located north of the three headlands are generally oriented ≈15° east of North (Figure 1b). The coastlines located to the south of the headlands are oriented ≈30° west of North for approximately 6 - 8 km (Figure 1b). In addition to the larger headland features, a small (~0.3 km) headland Mussel Pt. (34.93N) with a subaqueous rocky outcrop 2 km offshore (Figure 1c) is located 3 km north of Pt. Sal. Of note, San Luis Obispo (SLO) Bay is located at 35.17N.

2.2 Temperature Observations, Measured Winds, & Modeled Winds

Measures of temperature, \( T(y, z, t) \), were obtained over the vertical \((z)\) at 20 temperature-string (T-string) moorings (M) deployed in an alongshore array \((y)\) on the inner shelf from SLO Bay to Pt. Arguello, California (Figure 1b) for yeardays 244-287 (1 September to 14 October 2017). The N-S spacing between moorings ranged from 670 m to 50 km. A total of 15 moorings were deployed along the 9-m isobath (M1-5, M11-20). For 9-m moorings, measures were obtained from 1.5 m above the seabed to mean low-water with a vertical resolution of 0.5 m. Five moorings located near Pt. Sal (M6-10, Figure 1b), were deployed in 10-16 m water depths and obtained temperatures from 1.5 m above the sea bed to the free surface with a vertical resolution of 1.5 m. All thermistors were RBR soloT and sampled at 1 s. For analysis, data were averaged to 10
minutes. Winds were obtained hourly at four locations throughout the experiment site (Figure 1b, denoted at W1-W4). W1 is an offshore mooring site. Locations W2 and W4 are located very close to shore, and W3 is moored in 20 m depth. Modeled 10-m winds from a 2-km horizontal resolution Coupled Ocean/Atmosphere Mesoscale Prediction System (COAMPS) forecast model (Hodur, 1997) were linearly interpolated in three-dimensions (longitude, latitude, and time) to the temperature mooring locations. The easting and northing wind velocities $u_e$ and $u_n$ are defined in the toward-direction.

2.3 Temperature Filtering, Vertical EOF Analysis, and Alongcoast EOF analysis.

At each thermistor on each mooring, temperature $T(z,t)$ was decomposed into a mean$(T)$and four frequency bands, subtidal ($< 33^{-1}$ cph, $T^{(ST)}(z,t)$), diurnal ($33^{-1}$ to $16^{-1}$ cph, $T^{(DU)}(z,t)$), semidiurnal ($16^{-1}$ to $10^{-1}$ cph, $T^{(SD)}(z,t)$), and high frequency ($> 10^{-1}$ cph, not considered) with a frequency domain filter. Note the parenthetical superscript represents the frequency band of interest (i.e. ST, DU, and SD). Owing to differences in mooring water depths and thermistor locations over the vertical, vertical locations of temperatures observations are depth $(h)$ normalized, where $zh=0$ represents the sea surface and $zh=-1$ represents the ocean bottom. At each mooring the shape of the vertical temperature profile is isolated by a normalized mean temperature $\bar{T}(z/h)$, defined as

$$\bar{T}(z/h) = \frac{(T) - \langle T_b \rangle}{{\langle T_s \rangle} - \langle T_b \rangle}$$

(1)

where $\langle T_b \rangle$ and $\langle T_s \rangle$ are the mean near-bed and near-surface mean temperature, and the mean is a time-average over the experiment duration. Thus $\bar{T} = 0$ at the bed and $\bar{T} = 1$ at the surface. At each mooring, the dominant coherent vertical variability of temperature in each band $T^{(\text{band})}(z/h, t)$, where “band” is ST, DU, or SD, is decomposed into vertical EOFs
(eigenfunctions), $\phi_n^{(band)}(z/h)$, and time-varying amplitudes, $A_n^{(band)}(t)$, referred to as principal components (PCs) (Preisendorfer, 1988). Variability is separated into orthogonal modes, such that (for the subtidal band)

$$T^{(ST)}(z/h, t) = \sum_{n=1}^{N} A_n^{(ST)}(t)\phi_n^{(ST)}(z/h),$$

(2)

where $T^{(ST)}(z/h, t)$ is the subtidal (ST) temperature signal, $N$ is the total number of measurements over the vertical, and the first mode ($n=1$) describes the largest fraction of variance. For evaluating vertical structure across all moorings, the first EOF mode vertical structure, $\phi_1^{(band)}(z/h)$ is normalized in all 3 bands such that

$$\int_{-1}^{0} [\phi_1^{(band)}(z/h)]^2 d(z/h) = 1.$$  

(3)

### 3. Results

#### 3.1 Coastal Winds

The Fall 2017 modeled (COAMPS) wind patterns are consistent with previous wind observations in this region (Ohashi and Wang, 2004; Melton et al., 2009; Fewings et al., 2015). The easting winds, $u_e$, are dominated by persistent diurnal variability and some subtidal synoptic variability that is relatively consistent across latitude with weakening just south of the Pt. Sal (Figure 2a). The northing (or poleward) winds, $u_n$, are weaker than $u_e$ with both diurnal and subtidal variability (Figure 2b). Upwelling favorable southward ST $u_n$ events (blue colors in Figure 2b) persist for 3-7 days and are relatively coherent across latitude from 35.05N to the south. Relaxation or reversal events with northward $u_n$ winds (e.g., Melton et al., 2009; Washburn et al., 2011) also occur (green-yellow colors Figure 2b). The COAMPS modeled winds have variances that are similar to the observed winds (Figure 3a). The observed and modeled easting winds, $u_e$, have zero-lag correlation coefficients >0.5 (circles in Figure 3b). The observed and modeled
northing winds, \( u_n \), are also similarly correlated except at W2 (\( r=0.35 \), squares in Figure 3a). The similarity between model and observations indicate that using modeled winds at mooring locations is appropriate.

3.2 Vertical Water Temperature \( T(z,t) \) at Mooring 15 (M15)

An example of the M15 (at 34.78N) \( T(z,t) \) highlights the vertical and temporal variability (Figure 4) along the 50 km N-S array. M15 is located near the middle of the array between Pt. Sal and Pt. Purisima. The observed \( T(z,t) \) reveals a range of vertical and temporal variability (Figure 4a). The mean and ST temperature variability (Figure 4b) is characterized by colder water (< 14°C) upwelling events and warmer water (> 15°C) plume events (e.g., yeardays 246-260). The DU-temperature variability \( T^{(DU)}(z,t) \) occurs throughout the water column, and is ±1.0 °C (Figure 4c), with larger fluctuations occurring earlier in the experiment with intermittent events occurring later in the experiment. The SD temperature \( T^{(SD)}(z,t) \) also varies ±1.0 °C and is largest mid-water column, with strongest variability occurring earlier in the experiment (Figure 4d). Hourly tidal elevation is provided for context in Figure 4e. The tidal fluctuations on the west coast of the US are described by a mixed, semidiurnal tide with a range of ≈2 m, where the higher-high water precedes lower-low water (Friedrichs, 1995). The tidal amplitude is modulated by lunar (K1) and lunisolar (O1) diurnal tides with principal lunar semidiurnal tide (M2) (Godin and Martínez, 1994; Nidzieko, 2010) with maxima of the spring tides occurring around yeardays 284, 261, and 282 (Figure 4e).
3.3 Mid-water column temperature spectra at M1 and M7

Temperature spectra with 10 DOF and frequency resolution of 0.002 cph were computed in the mid-water column \(z=-4\) m at two contrasting sides: M1 (35.14N) to the north near SLO Bay and M7 (34.91N) at Mussel Pt. near Pt. Sal (Figure 5). At low frequencies in the ST band, the spectral energy is red with similar levels at M1 and M7. The ST band contributes 68% (M1) and 48% (M7) of the total temperature variance. At both M1 and M7, the DU and SD spectral peaks are similar per mooring. The DU and SD spectra peaks are broad and not at distinct tidal frequencies.

DU and SD spectral levels are elevated at M7 which has 3x more variance than at M1. Spectra levels decrease in the high frequency band (>0.1 cph). The spectral valleys between peaks are used to define the ST, DU, and SD band cutoffs (Section 2.3).

3.4 Vertical Temperature Statistics: Mean and subtidal, diurnal, and semidiurnal EOFs

Across all moorings, the normalized mean temperature \(\bar{T}(z/h)\) profile (Eq. 1) has a largely linear vertical structure with some mid-water column variation (Figure 6a). The mean (time-averaged) stratification \(\langle N^2 \rangle\) is estimated solely from the time-averaged temperature profiles (assuming salinity effects are weak) as

\[
\langle N^2 \rangle = -\frac{g \alpha}{\rho_o} \frac{\partial \bar{T}(z)}{\partial z}
\]  

(4)

where \(g\) is gravity, \(\rho_o=1025\) kg m\(^{-3}\), \(\alpha\) is the thermal expansion coefficient, and the mean temperature gradient is from the linear mean temperature profile (Figure 6a). The linear \(\bar{T}(z/h)\) profiles suggest that the time-averaged mean stratification \(\langle N^2 \rangle\) is largely depth-uniform across all moorings. In the subtidal band (ST), the first EOF represents between 95% to 99% of the variance, and across all moorings the \(\phi_1^{(ST)}(z/h)\) vertical structure is near one and mostly depth uniform (Figure 6b), consistent with a largely barotropic response. However, the mooring averaged...
\( \phi_1^{(ST)}(z/h) \) does have some vertical structure, varying by 0.18 top to bottom, which impacts the stratification on subtidal time-scales. The \( \phi_1^{(ST)}(z/h) \) variability across moorings is weak with small maxima near-surface and near-bed suggesting surface- or bottom-boundary layer processes. The weak \( \phi_1^{(ST)}(z/h) \) variability across moorings suggests the subtidal vertical variability can be well represented by a single mode across all moorings.

In the diurnal band (DU), the first EOF represents 75% to 95% (mean 87%) of the variance. Across all moorings, the mean vertical structure of \( \phi_1^{(DU)}(z/h) \) has a weak mid-water column maximum but is significant (⅓ of max) near surface and near bed (Figure 6c), indicating a mixed barotropic and first mode baroclinic response. This result is consistent with the mixed barotropic and baroclinic diurnal temperature structure observed in 8-m and 10-m depth off of Huntington Beach, CA (Kumar et al., 2016). The weaker near-surface first EOF \( \phi_1^{(DU)}(z/h) \) is not consistent with surface heating and cooling. Across all moorings, the \( \phi_1^{(DU)}(z/h) \) variability is weak mid-water column but increases near-surface and near-bed (dashed lines in Figure 6c), also suggesting that a single mode largely can represent the DU-vertical variability across all moorings. In the semidiurnal band (SD), the first EOF represents 76% to 95% (mean of 88%). Across all moorings, the mean \( \phi_1^{(SD)}(z/h) \) vertical structure has a clear mode-one like baroclinic response, although \( \phi_1^{(SD)}(z/h) \) is maximum in the lower-mid \((z/h=0.65)\) water column with asymmetric response above and below (Figure 6d). Similar \( \phi_1^{(SD)}(z/h) \) was also observed in 8-m and 10-m depth by Kumar et al. (2016). Across all moorings, \( \phi_1^{(SD)}(z/h) \) variability is elevated throughout the water column relative to other bands (Figure 6d -dashed lines), suggesting that a single vertical mode cannot represent the SD-vertical structure at all moorings.
Because the SD $\phi_{1}^{(SD)}(z/h)$ has the largest variability across moorings, the $\phi_{1}^{(SD)}(z/h)$ vertical structure is examined separately over all moorings relative to a linear mode 1 baroclinic structure (i.e., normalized $\sin(\pi(z+h)/h)$ for depth-uniform $N^2$ (Figure 7). At some locations (e.g., northern and southern M1-4, 13, 18, 19, 20 in Figure 7), the SD $\phi_{1}^{(SD)}(z/h)$ vertical structure is very similar to the linear baroclinic mode-1 structure, indicating largely linear SD dynamics. In contrast, locations near Pt. Sal and Mussel Point with the subaqueous rocky outcrop (Figure 1c) have a strongly bottom intensified $\phi_{1}^{(SD)}(z/h)$ (e.g., M6-9 in Figure 7) inconsistent with linear baroclinic mode-1 structure, indicating nonlinear dynamics and suggesting near-bottom cold bores. Particular locations such as M12 in the bay south of Pt. Sal and M16 just north of Pt. Purisima have a quasi-barotropic vertical structure with a depth-uniform lower layer and weakly decaying upper layer (Figure 7). Although clearly not a linear baroclinic mode-1, whether the SD processes inducing this vertical structure are more wave-like or advection is unclear.

3.5 Alongcoast and temporal temperature variability in the subtidal, diurnal, and semidiurnal bands

The latitudinal dependence of the ST temperature variability is examined with a reconstructed $T_{1}^{(ST)}(y, t)$ from the mean and the first ST EOF. As with the M15 observations (Figure 4), the primary ST signal is northward-propagating, warm-water plumes consistent with the wind relaxation events with colder upwelling-type conditions occurring in-between (Figure 8a). The first warm-water plume relaxation event was strong (starting yearday 247) with southern array temperature change of $\Delta T=6 \, ^{\circ}C$. This plume propagated northward through the entire array at 24.5 km d$^{-1}$ (0.28 m s$^{-1}$) and maximum plume temperature slowly decayed northward from 19.2 $^{\circ}C$ to 17.1 $^{\circ}C$ (Figure 8a). Plume duration increased slightly north of Pt. Purisima, suggesting
interaction with the headland. North of Pt. Sal, the warm water (>17 °C) is present for >10 d to yearday 260. The yeardays 255-258 warm-water event south of Pt. Sal has warmer water to the north and no propagation sense. Thus, it is not considered a plume. A warm-water event occurs on yeardays 255-258 south of Pt. Sal but without a sense of propagation and with warmer water to the north, and is not considered a plume. On yearday 271, a weak plume with ΔT=2 °C propagates northward at 15.7 km d⁻¹ ending at Pt. Sal. On yearday 283, the last warm-water plume with ΔT=3 °C propagates northward at 14.8 km d⁻¹ ending just south of Pt. Sal. During plume events, the weak ST EOF vertical structure $\phi_1^{(ST)}(z/h)$ (Figure 6b) also increases the ST stratification, consistent with previous observations (e.g., Melton et al., 2009).

In the diurnal band, $A_1^{(DU)}(y,t)$ varies ±1°C, with strongest variability in the entire region north of Pt. Sal. The weakest $A_1^{(DU)}(y,t)$ variability is south Pt. Purisima (Figure 8b). Just south of Pt. Sal, the $A_1^{(DU)}(y,t)$ variability is very weak. The DU-band variability is strongest from yeardays 250-270 north of Pt. Sal, and between Pt. Sal and Pt. Purisima. However, south of Pt. Purisima, the DU variability is relatively uniform in time (Figure 8b). The $A_1^{(DU)}(y,t)$ variability is not modulated by the subtidal changes in stratification, in contrast with Aristizabal et al. (2017) who suggested that the warm plume increased stratification increased the DU temperature signal.

In the semidiurnal band, $A_1^{(SD)}(y,t)$ varies ±1.5 °C (Figure 8d) with strongest variability in the between 34.9N and 35.0N, in the region near and north of Pt. Sal. The largest SD variability occurs between yeardays 250-258 (Figure 8d) along the straight coasts north of Pt. Sal and Pt. Purisima (Figure 1b). The SD variability is less spatially coherent than the DU variability. Consistent with previous studies (Aristizabal et al., 2017), the ST-varying stratification associated with plume events modulates the SD and DU-band variability (Figure 8b,c). As with the mid-water
column temperature spectra (Figure 5), the $A_{1}^{(SD)}(y,t)$ have broad semidiurnal peaks. At each mooring, applying a harmonic analysis (i.e., T-tide, Pawlowicz et al., 2002) reveals that >93% of the $A_{1}^{(SD)}(y,t)$ is incoherent with the M2, N2, S2 barotropic tides.

3.6 Latitudinal Bulk Temperature and Stratification Statistics

The latitudinal variability and the influence of the headlands Pt. Sal and Pt. Purisima is better understood through time-averaged statistics. Although most of the array is in 9-m depth, locations near the rocky Pt. Sal headland were in deeper depths from 10-16 m (Figure 9a). The time-averaged (over experiment duration) near-surface temperature $\langle T_s \rangle$ varies from 14.7 to 15.9 °C, with a general warming trend from south to north with colder $\langle T_s \rangle$ located just south of the two points, particularly at Pt. Sal (Figure 9b). The latitudinal variation in mean stratification $\langle N^2 \rangle$ is qualitatively similar to that of $\langle T \rangle$ (Figure 9b,c) with $\langle N^2 \rangle$ generally largest to the north (near 35.1N) and decreasing towards Pt. Sal (Figure 9c). Just south of Pt. Sal, $\langle N^2 \rangle$ weakens and subsequently increases toward Pt. Purisma, where it is relatively uniform from 35.85N to 34.65N (Figure 9c). The stratification $N^2$ variability in the ST-band (gray shading in Figure 9c) associated with relaxation and upwelling events is largest north of Mussel Pt. at 25% of the mean, and weaker south of Pt. Sal at about 12% of the mean. In the ST band, $\langle (A_{1}^{(ST)})^2 \rangle^{1/2}$ is relatively uniform in latitude, increasing gently to the south closer to the warm water source and with a sharp gradient just south of Pt. Sal (Figure 9d). In the DU band, $\langle (A_{1}^{(DU)})^2 \rangle^{1/2}$ is largest north of Pt. Sal, and is roughly constant from 35.07N to Pt. Sal (Figure 9e). Just south of Pt. Sal, a sharp reduction of $\langle (A_{1}^{(DU)})^2 \rangle^{1/2}$ occurs, which gently increases southward towards Pt. Purisima. South of Pt. Purisima, $\langle (A_{1}^{(DU)})^2 \rangle^{1/2}$ is also reduced before increasing gently farther south. The spatial pattern of the
dominant SD variability \((A^{(SD)}_1)^2\) is similar to that for DU variability (Figure 9f), also highlighting the strong effects of the headlands. One difference between SD and DU variability is that \((A^{(SD)}_1)^2\) decreases more strongly northward from Pt. Sal than for \((A^{(DU)}_1)^2\) (Figure 9e, f).

### 3.7 Diurnal Band Alongcoast EOF Analysis

To further evaluate DU-temperature variability, \(A^{(DU)}_1(y, t)\) is plotted as a function of local time of day in hours versus yearday at moorings M1 and M7 (Figure 10), separated by about 20 km. At each mooring, the \(A^{(DU)}_1(y, t)\) daily maximum is generally phase locked (with some variability) with mean occurring around 1700 and 2200 for M1 and M7, respectively (Figure 10). Thus, M1 and M7 have a DU-temperature variability alongshore phasing with a 5 h lag. For reference, the hour of occurrence for the daily maximum DU-winds consistently develops at 1600 (not shown).

Here, the alongcoast coherent and propagating diurnal temperature variability, \(A^{(DU)}_1(y, t)\), is examined with a Hilbert EOF (e.g., Horel, 1984) that resolves stationary and propagating covariability (e.g., Horel, 1984; Merrifield and Guza, 1990). For the DU-band, a complex \(A^{*(DU)}_1(y, t)\) time-series is generated

\[
A^{*(DU)}_1(y, t) = A^{(DU)}_1(y, t) + i\tilde{A}^{(DU)}_1(y, t),
\]

where \(\tilde{A}^{(DU)}_1\) is the quadrature function (Hilbert transform) of \(A^{(DU)}_1(y, t)\) and \(i = \sqrt{-1}\). The \(A^{*(DU)}_1(y,t)\) variability is decomposed into horizontal complex EOFs (CEOF), \(H_n(y)\), such that

\[
A^{(DU)}_1(y, t) = \sum_{n=1}^{N} B_n(t) H_n^{(DU)}(y),
\]
where the CEOFs $H_n^{(DU)}(y)$ are the eigenvectors of the Hermitian $A_{1 \times (DU)}$ covariance matrix.

The DU first CEOF alongcoast phase is estimated as

\[
\theta_1^{(DU)}(y) = \arctan\left(\frac{\text{Im}[H_1^{(DU)}(y)]}{\text{Re}[H_1^{(DU)}(y)]}\right),
\]

where $\text{Re}$ and $\text{Im}$ represent real and imaginary components. The alongcoast DU magnitude is represented by $|H_1^{(DU)}(y)|$.

The first CEOF of alongcoast DU temperature explains 67% of the latitudinal spatial variability. Consistent with large variance fraction explained, the first CEOF magnitude $|H_1^{(DU)}(y)|$ (Figure 11a) has a structure similar to $\langle (A_1^{(DU)})^2 \rangle^{1/2}$ (Figure 9d). The first CEOF phase $\theta_1^{(DU)}(y)$ reveals a coherent southerly propagation of 1.9 ms\(^{-1}\) between 35.15N and Pt. Sal (Figure 11b) with <5% propagating phase error (Merrifield and Guza, 1990) where the $|H_1^{(DU)}(y)|$ is largest (Figure 11a). Where $|H_1^{(DU)}(y)|$ is elevated between Pt. Sal and Pt. Purisima, the first CEOF phase $\theta_1^{(DU)}(y)$ also has a sense of coherent southerly propagation of 0.9 ms\(^{-1}\), slower than north of Pt. Sal.

Diurnal winds are often invoked as locally driving baroclinic diurnal variability (Lerczak et al., 2001; Cudaback and McPhee-Shaw, 2009; Nam and Send, 2013; Kumar et al., 2016; Aristizabel et al., 2016). Here, we test whether an alongshore propagating diurnal wind phase drives a local alongshore propagating diurnal response, giving an alongcoast propagating $\theta_1^{(DU)}(y)$ as observed (Figure 11b). An alongshore CEOF decomposition was performed independently on the diurnal-band northing and easting winds (denoted with superscript “DW”). The first CEOF of easting and northing diurnal wind explains 93% and 68% of the diurnal wind variance, respectively, and the E and N magnitude $|H_1^{(DW)}(y)|$ varies by a factor of 2 over the
However, the phase \( \theta_1^{(DW)}(y) \) is relatively constant alongcoast, indicating the diurnal winds are in phase alongcoast (Figure 12b). Therefore, the observed diurnal temperature southward propagation is not directly forced by the wind.

### 3.8 Semidiurnal band depth-integrated energy and decorrelation length-scales

For depth-uniform mean stratification \( \langle N^2 \rangle \) (Figure 9c) and for kinetic and potential energy equipartition, the depth- and time-averaged semidiurnal energy \( E^{(SD)} \), accurate to second order, is a function of the first EOF amplitude \( A_1^{(SD)} \) and mean stratification \( \langle N^2 \rangle \) as,

\[
E^{(SD)} = \frac{g^2 \alpha_2 \langle (A_1^{(SD)})^2 \rangle}{4 \rho_o \langle N^2 \rangle}. \quad [J/m^2] \tag{8}
\]

The expression for \( E^{(SD)} \) is independent of \( \phi_1^{(SD)}(z/h) \) because of the normalization (2). The depth-averaged energy (as opposed to depth-integrated) is chosen to remove the effect of mooring depth variations (Figure 9a). Both standing and propagating energy \( E^{(SD)} \) (e.g., Lerczak et al. 2003; Kumar et al. 2016) are included in (Eq. 8). The latitudinal dependence of the SD-band energetics is examined with a normalized SD-band energy \( \bar{E}^{(SD)} \), defined as

\[
\bar{E}^{(SD)} = E^{(SD)}/\max(E^{(SD)}). \tag{9}
\]

The overall alongshore variation of \( \bar{E}^{(SD)} \) (Figure 13) also highlights the strong effects of the headlands in the SD band. The \( \bar{E}^{(SD)} \) is weak (about 0.2) to the north, increases strongly south of 35N toward a maximum \( \bar{E}^{(SD)} = 1 \) at Pt. Sal (Figure 13), and sharply decreases by a factor of 4 just to the south of Pt. Sal. Just north of Purisima Pt. a secondary \( \bar{E}^{(SD)} = 0.5 \) maximum occurs, which decays south of Purisima Pt. although not as dramatically as at Pt. Sal (Figure 13).
Any spatial coherence of the SD band $A_1^{(SD)}(y, t)$ (Figure 8c) is less clear than for ST or DU variability (Figure 8a,b). SD band $A_1^{(SD)}(y, t)$ alongshore spatial decorrelation scales are estimated from lagged cross-correlation analysis. Lagged $A_1^{(SD)}(y, t)$ cross-correlation functions, $r_{ij}$, are estimated over all mooring pairs for 48-hr windows, such that

$$ r_{ij}(\tau) = \frac{E[A_1^{(SD)}(y_i, t)A_1^{(SD)}(y_j, t+\tau)]}{E[A_1^{(SD)}(y_i, t)A_1^{(SD)}(y_j, t)]}, $$

(10)

where $E$ is the expectation operator, $i$ and $j$ denote unique pair of moorings, $t$ is time, and $\tau$ is the temporal lag. The maximum $r_{ij}(\tau)$ per mooring pair is referred to as $r_{max}^{(SD)}$ and is averaged in 3-km spatial bins. The $r_{max}^{(SD)}$ decays rapidly with alongshore instrument separation (Figure 14) with an e-folding decay scale of 7.5 km, indicating that $T_1^{(SD)}(y, t)$ decorrelates on relatively short alongshore spatial scales. Although the $A_1^{(SD)}(y, t)$ variance (and $E^{(SD)}$) largely varies smoothly with latitude, except near Pt. Sal (Figure 9e), SD-band temperature variability is only coherent over short distances relative to the the ST (Figure 8a) and DU (Figure 11) variability.

4. Discussion

4.1 Subtidal Temperature Variability

Previously, Melton et al. (2009) and Washburn et al. (2011) observed ST poleward warm plumes, originating out of the Santa Barbara Channel, from north of Pt. Conception up to Pt. Sal with 3 nearshore mooring locations on the south side of the topographic points. Here, the increased alongshore and vertical density of the temperature observations that also extended north of Pt. Sal highlight differences in ST response north and south of Pt. Sal. The vertical structure of the first EOF of ST temperature, which explains >95% of the variance (Figure 6b), affects the local
stratification which influences, in particular, the SD-temperature variability (Figure 8c), consistent with Aristizabal et al. (2017). Melton et al. (2009) found that the stratification was weak prior to the arrival of the warm plume, and increased after the plume. The ST latitudinal variability, due to variations in plume excursion, highlights the relationship of temporal and alongshore temperature variability across frequency bands.

Three warm plumes with varying poleward propagation speeds and ΔT were observed during wind relaxation events. The strongest warm plume (yearday 247, early September) supported a ΔT as large as 6 °C with propagation speed of 25 km d\(^{-1}\) and extended north of Pt. Sal to M1 (Figure 8a). This warm plume occurred during a relatively long, 10-day (yearday 245-255) wind relaxation event (Figure 2b). This large warm plume is consistent with Melton et al. (2009) who found more warm plumes in September and longer propagation distances associated with longer lasting wind relaxations. Weaker warm plumes that occurred on yearday 271 and 283 (Figure 8a) were limited to south of Pt. Sal and associated with shorter (2-3 day) wind relaxations (Figure 2b).

The warmer water observed north of Pt. Sal for yearday 251-258 appears disconnected from the waters south of Pt. Sal (Figure 8a). Potential mechanisms that can explain the warm water north of Pt. Sal include: First, the the strong yearday 247 plume that propagated north of Pt. Sal could have warm water trapped between Pt. Sal and Pt. San Luis headlands. Second, another warm plume originating out of the SBC could be deflected offshore near Pt. Conception and re-attach north of Pt. Sal. Washburn et al. (2011) observed a 17-km diameter submesoscale anticyclonic eddy that develops between Pt. Arugula and Pt. Purisima. This eddy could deflect the warm plume out of SBC farther offshore, where the reconnection occurs somewhere north of Pt. Sal. A third explanation is that warmer water north of Pt. Sal is associated with warmer water that develops
within upwelling shadow of Pt. San Luis that tends to occur in late summer through the early fall (Walter et al., 2018).

4.2 Diurnal Temperature Variability

The DU first EOF represented 75%-95% of the variance and its vertical structure was a mix of barotropic and a linear baroclinic structure (Figure 6c). This contrasts with the DU temperature variability farther offshore in 50-100 m depth (Pidgeon and Winant, 2005), which required two EOFs to represent only 66% of the variance. The offshore first mode was surface intensified, attributed to surface forcing and the 2nd mode had mid-depth maximum, attributed to baroclinic tidal processes (Pidgeon and Winant, 2005). Here, the DU first EOF does not have a surface maximum indicating that surface forcing is not the dominant process.

Inner-shelf DU-band temperature variability is always evident (Figure 8), consistent with SMB observations both farther offshore in 50-100 m depths (Pidgeon and Winant, 2005) as well as in 15-m depth (Cudaback and McPhee-Shaw, 2009; Aristizabal et al., 2016). The offshore DU temperature standard deviation was near 0.2 °C (Pidgeon and Winant, 2005) is of the same order as the 0.2-0.4 °C$\left(\left(A_{1}^{(DU)}\right)^{2}\right)^{1/2}$ (Figure 9e). The near-bed nearshore DU temperature variability over multiple years (Aristizabal et al. 2016) is consistent with that observed here during the Fall 2017. In contrast with Cudaback and McPhee-Shaw (2009) and Aristizabal et al. (2017), a relationship was not observed between the envelope of $A_{1}^{(DU)}(y,t)$ and the time-varying stratification, given as a proxy by $A_{1}^{(ST)}(y,t)$.

Aristizabal et al. (2016) observed that the DU winds and DU near-bed temperature were in phase in the SBC, concluding that DU temperature signal was directly DU wind forced and did not propagate alongshore. In contrast, the DU temperature has a linear phase between SLO Bay
and north of Pt. Sal, suggesting a 1.9 m/s southerly propagation of $A_{1}^{(DU)}(y,t)$ over a region of 18 km (Figure 11b). There is also sense of southerly propagation between Pt. Sal and Purisima Pt. (Figure 11b). This southward propagation is unexpected as internal waves at diurnal frequencies are evanescent ($< f$) at this latitude. The ubiquitous DU temperature variability (Figure 8b and Pidgeon and Winant, 2005; Cudaback and McPhee-Shaw, 2009; and Aristizabel et al., 2016) suggests that subtidal vorticity reducing the local Coriolis parameter (e.g., Lerczak et al. 2001) to allow propagating waves is not the mechanism here. An internal Kelvin wave could support frequencies $< f$, but would propagate northward.

However, if this were a propagating internal wave, a strong baroclinic vertical structure would be expected, as opposed to the mixed barotropic and baroclinic vertical structure (Figure 6c). The inferred phase speed (1.9 m/s) is too rapid to be advection. The inferred phase speeds may be larger than the actual propagation speeds if the propagation direction is not alongshore aligned. The modeled COAMPS winds had nearly uniform alongshore phase, indicating simultaneous occurrence along the entire array (Figure 12b). Thus the DU winds do not directly force generate this DU-band $A_{1}^{(DU)}(y,t)$ southward phase propagation, which is consistent with the lack of near-surface intensified DU first EOF (Figure 6c).

One proposed mechanism for the observed DU temperature response is associated with the diurnal lateral advection of water masses of differing temperatures, which would represent a DU standing motion. Walter et al. (2017) found that the DU seabreeze transported warm water out of the San Luis Obispo (SLO) Bay (at the northern end of our array Figure 1 b) at the surface and transported colder water in the bay at the bottom. When the DU sea breeze relaxed, the water advected back into the bay as a buoyant plume front. This mechanism is roughly consistent with the vertical structure (Figure 6c). Concomitantly, during the late summer through the early fall an
upwelling shadow generates a persistent warm body of water north of Pt. Sal (Walter et al., 2018). The core of this warm water would extend from M1 to M4 with a temperature gradient extending to M7 (see Figure 9 in Walter et al., 2018). DU winds could be advecting these water masses such that alongshore temperature gradient produces an alongshore variable $A_1^{(DU)}(y, t)$ with a time lag.

4.3 Semidiurnal Temperature Variability

The separation of semidiurnal temperature variability into vertical modes and alongshore structure elucidates how both the SD vertical structure (Figure 7) and the depth-integrated energy $\tilde{E}^{(SD)}$ (Figure 13) vary alongshore. Previous SD-band temperature analyses over similar alongshore scales have focused on a single near bed thermistor (e.g., Aristizabal et al., 2016), resulting in aliasing SD vertical structure into alongshore variability. For example, the ratio of near-bed $\phi_1^{(SD)}(z/h)$ between M14 and M1 is $> 2$, even as their $\tilde{E}^{(SD)}$ are similar. Using only near-bed thermistors would lead to the conclusion that M14 has much stronger SD variability than M1. Alongshore variability analysis in lower frequency bands (e.g. Tapia et al, 2014) does not have such aliasing problems as the vertical structure of low-frequency variability is much more depth uniform (Figure 6).

The SD depth integrated energy $\tilde{E}^{(SD)}$ varied strongly (factor 3X) alongshore in 9-16 m water depth during Sept-Oct 2017, particularly south of Pt. Sal and Pt. Purisima (Figure 13). During June-July 2015 in 50-m water depth, spanning 7 km north and 3 km south of Pt. Sal, the incident SD-band energy flux varied weakly in the alongshore by a factor of 1.25X (Kumar et al., 2019). This suggests that the bathymetry near the headlands refracted the incident SD energy onto the points, particularly Pt. Sal. Onshore of 50-m depth, SD-band energy is also likely dissipating (Colosi et al. 2018), reflecting (e.g., Lerczak et al., 2003; Kumar et al., 2016), or being transferred.
to higher frequencies (e.g., Holloway et al. 1999). The weak $E^{(SD)}$ at M10 (in the bay south of Pt Sal, Figure 1c) is likely due, in part, to strong bathymetric refraction within the bay and also the shallower bathymetric slopes (relative to M8 at Pt. Sal) giving greater propagation distance for dissipation. Strong SD alongshore temperature variability over 10 km was also observed in Todos Santos Bay, MX in depths $\approx 20$ m attributed to stratification and bathymetric variability (Filonov et al., 2014).

The SD amplitude $A_1^{(SD)}(y,t)$ had relatively, short (7.5 km) alongshore decorrelation length-scales (Figure 14). This is consistent with the incoherent SD-band near-bed temperature variability at all location along the mainland from within the Santa Barbara Channel up to Pt. Sal (Aristizabal et al., 2016). In contrast, the Aristizabal et al. (2016) SD-temperature variability coherently propagated in locations near Santa Cruz Island (Channel Islands), which was very close to a significant internal tide generation region (Buijsman et al., 2012). Additionally, SD band temperature variability at the north and south ends of a 6-km long Chilean bay was coherent with the barotropic tide, presumably due to proximity to a submarine canyon, a likely internal tide generation region (Bonicelli et al., 2014). In 50-m water depth offshore of Pt. Sal in summer 2015, about half of the SD energy flux was coherent (i.e., at SD tidal frequencies, M2, N2, S2) and half incoherent in both observations and models (Kumar et al, 2019). In contrast, 4 km farther onshore, in about 10 m water depth, the $A_1^{(SD)}(y,t)$ have broad spectra with very little of the variability (<7%) coherent with SD tidal frequencies (M2, N2, S2). Thus, the variable stratification and Doppler shifting that induce incoherence are far larger in the shallower regions (<50 m depth) than deeper on the shelf.

The $\phi_1^{(SD)}(z/h)$ vertical structure (Figure 7) also varies alongshore, suggesting at different locations linear internal waves, nonlinear cold bores, and a potentially quasi-barotropic response
due to advection. SD-band internal waves are effective at transporting larvae and nutrients across
the inner shelf (e.g., Pineda 1999; Lucas et al., 2011), yet such induced transport likely depends
on the internal wave vertical structure. Alongshore variations in SD-band temperature vertical
structure (Figure 7) alone could reflect spatial variability of transport, and potentially alongcoast
population variability. Locations with nonlinear internal waves would be expected to have larger
energy $\bar{E}^{(SD)}$ than locations with a linear response. The relationship between the $\phi_1^{(SD)}(z/h)$
vertical structure and SD depth-averaged energy $\bar{E}^{(SD)}$ is examined here, to better diagnose the
processes driving the SD temperature variability. The misfit of $\phi_1^{(SD)}(z/h)$ from a linear mode one
baroclinic response $\sin[\pi (z + h)/h]$ (compare blue and orange dashed curves, Figure 7) is
represented by mean square misfit $\epsilon^2$, defined as

$$
\epsilon^2 = \int_{-1}^{0} \left( \phi_1^{(SD)}(\frac{z}{h}) - \sin[\pi (z/h + 1)] \right)^2 d(z/h), \tag{11}
$$

where $\epsilon^2 = 0$ is a linear mode 1 response.

The misfit $\epsilon^2$ varies from 0.01 to 0.24, indicating near-exact to strong deviation from linear
baroclinic response (Figure 15). The misfit $\epsilon^2$ and normalized SD energy $\bar{E}^{(SD)}$ are linearly related
with $r^2 = 0.61$, with stronger energy related to larger misfit (Figure 15), consistent with linear
baroclinic structure for weak SD internal tide and nonlinear bottom intensified structure with
stronger SD internal tides. The SD misfit and energy are heuristically grouped by geography and
$\phi_1^{(SD)}(z/h)$ (colors in Figure 15). The generally largest $\bar{E}^{(SD)}$ and largest $\epsilon^2$ are clustered into a
group denoted B (red in Figure 15) consisting of M5 to M10, all locations relatively close to
Mussel Pt. and Pt. Sal, indicating that a regional internal wave hotspot. In contrast, the generally
smallest $\bar{E}^{(SD)}$ and $\epsilon^2$ are associated with the locations near SLO Bay (M1-M4, group A) or
moorings near and south of Pt. Purisima (M13-M15 and M16-M20, group D) consistent with
weaker and more linear SD internal waves. Intermediate $\mathcal{E}^{(SD)}$ and $\epsilon^2$ located between Pt. Sal and Pt. Purisima (M11, M12, and M16, group C) have quasi-barotropic $\phi_1^{(SD)}(z/h)$. This may also reflect nonlinear internal wave processes or SD advective processes (such as headland wakes) inducing observed temperature variability. This A-D grouping is heuristic, particularly for groups C and D south of Pt. Sal. For example M15 and M17 (Figure 7) could be in Group C and not group D. Nevertheless, the relationship between $\mathcal{E}^{(SD)}$ and $\epsilon^2$ is robust and is independent of this grouping.

Mooring M10 is located in the bay south-east of Pt Sal (Figure 1c) and is an outlier in the linear relationship between $\mathcal{E}^{(SD)}$ and $\epsilon^2$ (red open circle of group B, Figure 15). M10 has relatively large $\epsilon^2 = 0.17$, consistent with other moorings of its geographical grouping B, but weak $\mathcal{E}^{(SD)} = 0.27$, relative to other group B moorings (red, Figure 15). Mooring M10 is less than 2 km separated from the M8 just west of Pt. Sal (see Figure 1c) which has strongest $\mathcal{E}^{(SD)}=1$. M10 is the most sheltered mooring location yet has vertical structure consistent with a nonlinear cold bore (Figure 7j). Offshore of M10 is also more shallowly sloped than at nearby locations (e.g., M8, M9, and M11, Figure 1c). The SD internal waves incident to Pt. Sal are strongly nonlinear (Colosi et al., 2018). The M10 deviation in the $\mathcal{E}^{(SD)}$ and $\epsilon^2$ relationship suggests that the cross-shelf nonlinear transformation of the SD-band internal waves into M10 is fundamentally different than at other location.

5. Summary
Twenty moorings with a high vertical density of thermistors were deployed in 9-16 m water depths along 50 km of the central California inner shelf within the Santa Maria Basin to measure the temporal, vertical and alongcoast inner-shelf temperature ($T$) variability (Figure 1). The coastline has regions of relatively straight beach shorelines interrupted by coastal headlands, Pt. Sal and Pt. Purisima in particular. COAMPS modeled winds described the observed winds reasonably well, and had strong diurnal and synoptic variability. Temperature spectra had peaks with significant variability in the subtidal (ST), diurnal (DU) and semidiurnal (SD) frequency bands (Figure 5) motivating evaluation of (co-)variability in each band. In each frequency band, the vertical $T$ structure at each mooring is characterized by the first EOF $\phi_1(z/h)$ describing a high fraction of variance (Figure 6).

The ST vertical structure $\phi_1^{(ST)}(z/h)$ was quasi depth-uniform with weak linear vertical structure and minimal alongshore variability (Figure 6b). The ST variability was dominated by three warm plumes with varying poleward propagation speeds and $\Delta T$ that occurred as a result of wind relaxation events. The strongest warm plume supported a $\Delta T$ as large as 6 °C with propagation speed of 25 km d$^{-1}$, extended north of Pt. Sal, during a relatively long (10-day) wind relaxation event (Figure 2b, 8a). Weaker warm plumes (Figure 8a) were limited to the south of Pt. Sal and were associated with shorter (2-3 day) wind relaxations (Figure 2b). The ST variability is relatively uniform in latitude, increasing gently to the south closer to the warm water source from the SBC and with a sharp gradient just south of Pt. Sal (Figure 9d), the typical northward extent of the warm water plumes. The warm plumes also enhance the ST-band stratification $N^2$ (gray shading in Figure 9c) is largest north of Mussel Pt., and weakens south of Pt. Sal.

Inner-shelf DU-band temperature variability was always evident (Figure 8), consistent with previous SMB observations (Cudaback and McPhee-Shaw, 2009; Aristizabal et al., 2016). The
DU vertical structure $\phi_1^{(DU)}(z/h)$ was a mix of barotropic and linear baroclinic structure (Figure 6c), and did not have a surface maximum indicating that surface forcing is not the dominant process. This differs from previous regional observations in deeper water (Pidgeon and Winant, 2005). The time-envelope of DU variability was not modulated by the ST stratification changes, in contrast to previous regional results (e.g., Cudaback and McPhee-Shaw, 2009; Aristizabal et al., 2017). The magnitude of DU temperature variability (0.2-0.4 °C, Figure 9e) is similar to offshore observations (Pidgeon and Winant, 2005), and is largest north of Pt. Sal with sharp reductions south of Pt. Sal and Pt. Purisima. The alongcoast first CEOF of DU temperature explains 67% of the latitudinal variability. The DU first CEOF has an alongcoast linear phase north of Pt. Sal, suggesting a 1.9 m/s southerly propagation (Figure 11b). The modeled COAMPS winds (Figures 2, 3) had nearly uniform alongshore phase (Figure 12b), indicating that the DU winds do not directly generate this DU temperature southward phase propagation, in contrast to previous regional near-bed inner-shelf temperature observations (Aristizabal et al., 2016).

Inner-shelf SD-band temperature variability is modulated by the ST-band stratification (Figure 8), consistent with previous results. The SD vertical structure $\phi_1^{(SD)}(z/h)$ (Figure 6d, 7) also varies alongshore, suggesting at different locations semidiurnal linear internal waves, nonlinear cold bores, and potentially quasi-barotropic response due to advective processes (Figure 7). The SD depth-averaged energy, $\tilde{E}^{(SD)}$, varied strongly (factor 3X) alongcoast particularly near the headlands (Figure 13), with locations near to Mussel Pt. and Pt. Sal a regional internal wave hotspot. SD temperature variability was incoherent with barotropic tides and decorrelates alongshore in 7.5 km alongshore. This contrasts with SD temperature variability in 50-m water depth a few km offshore which is 50% coherent and has much longer alongshore length-scales (Kumar et al., 2019). Normalized SD depth-averaged energy was found to be linearly related
(r²=0.61) to baroclinic vertical structure (linear to non-linear forms) and heuristically related to geographic regions. These alongshore variations in SD variability and vertical structure are likely important to larval transport.

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Figure 1. a) Map of the Santa Maria Basin in context with the Santa Barbara Channel and Pt. Conception. Overview of the study region highlighted with a black box, which is shown in b). b) Map of the experiment site. c) Map of Pt. Sal. Temperature mooring positions are indicated by black dots. Location of wind observations are indicated by black asterisks. Depth contours are depicted as gray lines.
Figure 2. Hovmoller diagrams of spatially-interpolated to mooring locations (Figure 1b) COAMPS modeled a) easting winds, $u_e$, and b) northing, $u_n$, wind velocities as a function of latitude and yearday. Horizontal dashed black lines indicate the locations of Pt. Sal and Purisima Pt. The color-scale represents wind velocity.
Figure 3. COAMPS modeled (Figure 2) and observed W1-W4 (Figure 1b) comparison for $u_e$ (circles) and $u_n$ (squares) (a) modeled versus observed wind variance with station location colored (see legend) and (b) correlation coefficient versus station location. Black dashed line represents the linear 1:1 line.
Figure 4. Temperature observations versus 2017 yearday and height above the bed at M15 for the a) total, b) subtidal, c) diurnal, and d) semidiurnal frequency bands (see Figure 1b). e) Tidal elevations versus 2017 yearday.
Figure 5. Mid-water column (at $z=-4.0$ m) temperature spectra versus frequency (in cycles per hour, cph) at M1 (solid) and M7 (dashed). The subtidal (ST), diurnal (DU), and semidiurnal (SD) bands are indicated. Spectra were computed using 6-day hamming window with 50% overlap resulting in 10 DOF and 0.002 cph frequency resolution. Thick vertical black line in the right-hand side represents the 95% confidence levels.
Figure 6. The mean (solid lines) and +/- standard deviations (dashed) of temperature statistics across all 20 moorings as a function of normalized depth \( z/h \): (a) normalized time-mean temperature, and normalized first EOF \( \phi_1(z/h) \) for the (b) subtidal (c) diurnal, and (d) semidiurnal frequency band. The moorings span a depth range of 9 to 16 m (Figure 1a).
Figure 7. Semidiurnal band first EOF $\phi_1^{(SD)}(z/h)$ (blue lines) and normalized linear baroclinic mode 1 structure, $\sin[\pi (z + h)/h]$ (orange dashed lines) as a function of normalized depth $z/h$ for the 20 moorings (Figure 1b) as indicated above each panel spanning a depth range of 9 to 16 m. The normalization is given in (1) and (3).
Figure 8. Hovmoller diagrams of (a) mean plus subtidal near-surface temperature and (b) first EOF amplitude $A_1^{(DU)}$, and (c) $A_1^{(SD)}$ as a function of latitude and yearday. Horizontal dashed black lines indicate the locations of Pt. Sal and Pt. Purisima. The color-scale represents temperature in Celsius. Diamonds on the ordinate represent mooring locations.
Figure 9. Mooring temperature statistics (blue curve and dots) versus latitude: (a) mean depth (b) mean (time-averaged) temperature $\langle T \rangle$ at near-surface location $z = -1.7$ m below mean tide level. (c) mean buoyancy frequency squared $\langle N^2 \rangle$ ± ST-band standard deviation (shading), (d) subtidal first EOF amplitude standard deviation $\langle (A_1^{(ST)})^2 \rangle^{1/2}$, (e) diurnal first EOF amplitude standard deviation $\langle (A_1^{(DU)})^2 \rangle^{1/2}$, (f) semidiurnal first EOF amplitude standard deviation $\langle (A_1^{(SD)})^2 \rangle^{1/2}$.

Horizontal dashed black lines indicate the locations of Pt. Sal and Purisima Pt.
Figure 10. The DU EOF1 amplitude $A_{1}^{(DU)}(y, t)$ as a function of local time of day in hours and yearday for moorings (a) M1 and (b) M7.
Figure 11. Alongshore diurnal temperature first Hilbert EOF versus latitude: (a) magnitude, $|H_1^{(DU)}(y)|$, and (b) phase, $\theta_1^{(DU)}(y)$. This first Hilbert EOF explains 67% of the spatial variability of the diurnal first EOF amplitude $A_1^{(DU)}(y,t)$ (Eq. 2). Horizontal dashed black lines indicate the locations of Pt. Sal and Purisima Pt. The magenta dashed lines indicate the inferred alongshore (latitude) propagation speeds north of Pt. Sal ($c=1.9$ m/s) and between Pt. Sal and Pt. Purisima ($c=0.9$ m/s).
Figure 12. Alongshore first Hilbert EOF $H_{1}^{(DW)}(y)$ of the spatially-interpolated COAMPS, diurnal winds for easting ($u_e$, blue lines), and northing ($u_n$, orange lines) for (a) magnitude $|H_{1}^{(DW)}(y)|$ and (b) phasing, $\theta_{1}^{(DW)}(y)$, versus latitude. $H_{1}^{(DW)}(y)$ explains 93% of the spatial variability of the diurnal first amplitude. Horizontal dashed black lines indicate the locations of Pt. Sal and Purisima Pt.
Figure 13. Normalized SD depth-averaged energy, $\tilde{E}^{(SD)}$ (Eqs. 8, 9), as a function of latitude. The locations of Pt. Sal and Pt. Purisima are indicated with the dashed lines.
Figure 14. Maximum SD band $A_1^{(SD)}$ cross-correlation coefficient $r_{max}^{(SD)}$ (Eq. 10) versus alongshore separation. The $r_{max}^{(SD)}$ is bin-averaged over 3 km and mean (black squares) and 95% confidence interval (vertical lines) are shown.
Figure 15. Normalized semidiurnal depth-averaged energy $\tilde{E}^{(SD)}$ (Eqs. 8, 9) versus misfit from baroclinic mode one $\epsilon^2$ (Eq. 11). The squared correlation is $r^2 = 0.61$. Color represents heuristic groupings (A, B, C, D) based on geography (Figure 1b) and on $\phi_1^{(SD)}(z/h)$ vertical structure (Figure 6) as discussed in Section 4.3. Red-circled red dot represents results from M10.