



NAVAL POSTGRADUATE SCHOOL

MONTEREY, CALIFORNIA

DISSERTATION

**CROSS-SHELF HEAT FLUX AND TEMPERATURE
VARIABILITY ON THE CALIFORNIA INNER SHELF**

by

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September 2018

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REPORT DOCUMENTATION PAGE			<i>Form Approved OMB No. 0704-0188</i>	
Public reporting burden for this collection of information is estimated to average 1 hour per response, including the time for reviewing instruction, searching existing data sources, gathering and maintaining the data needed, and completing and reviewing the collection of information. Send comments regarding this burden estimate or any other aspect of this collection of information, including suggestions for reducing this burden, to Washington headquarters Services, Directorate for Information Operations and Reports, 1215 Jefferson Davis Highway, Suite 1204, Arlington, VA 22202-4302, and to the Office of Management and Budget, Paperwork Reduction Project (0704-0188) Washington, DC 20503.				
1. AGENCY USE ONLY (Leave blank)	2. REPORT DATE September 2018	3. REPORT TYPE AND DATES COVERED Dissertation		
4. TITLE AND SUBTITLE CROSS-SHELF HEAT FLUX AND TEMPERATURE VARIABILITY ON THE CALIFORNIA INNER SHELF			5. FUNDING NUMBERS N0001418WX00229	
6. AUTHOR(S) Thomas M. Freismuth Jr.				
7. PERFORMING ORGANIZATION NAME(S) AND ADDRESS(ES) Naval Postgraduate School Monterey, CA 93943-5000			8. PERFORMING ORGANIZATION REPORT NUMBER	
9. SPONSORING / MONITORING AGENCY NAME(S) AND ADDRESS(ES) N/A			10. SPONSORING / MONITORING AGENCY REPORT NUMBER	
11. SUPPLEMENTARY NOTES The views expressed in this thesis are those of the author and do not reflect the official policy or position of the Department of Defense or the U.S. Government.				
12a. DISTRIBUTION / AVAILABILITY STATEMENT Approved for public release. Distribution is unlimited.			12b. DISTRIBUTION CODE A	
13. ABSTRACT (maximum 200 words) The inner continental shelf is generally defined as the region from the edge of depth-limited wave breaking to water depths of tens of meters. It is an important region because it is the connection between the coastal ocean and the surf zone. During the summer of 2015 and 2017, two major field experiments were conducted to study water temperature and currents along the central California inner shelf. A headland-and-bay topography characterizes the field site with stretches of sandy or rocky coastlines and two headlands. Vertical profiles of co-located ocean temperature and currents were obtained in a dense cross-shore array from the 50- to the 5-m isobaths during the 2015 experiment. During the 2017 field experiment, vertical profiles of temperatures were obtained in an alongshore array on the 9-m isobath that spanned 55 km of coastline. The results show the contributions from coastal upwelling, sea breeze-driven processes, and semidiurnal internal waves to the net transport of heat onto the inner shelf. The first observations of warm internal tide bores propagating to the surf zone, and the spatial and temporal scales of the temperature variability over 55 km of coastline are discussed. The Office of Naval Research funded this research.				
14. SUBJECT TERMS inner shelf, cross-shore heat flux, temperature variability, nearshore, internal tide bores, coastal oceanography, heat budget, subtidal, diurnal, semidiurnal			15. NUMBER OF PAGES 101	
			16. PRICE CODE	
17. SECURITY CLASSIFICATION OF REPORT Unclassified	18. SECURITY CLASSIFICATION OF THIS PAGE Unclassified	19. SECURITY CLASSIFICATION OF ABSTRACT Unclassified	20. LIMITATION OF ABSTRACT UU	

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**CROSS-SHELF HEAT FLUX AND TEMPERATURE VARIABILITY ON THE
CALIFORNIA INNER SHELF**

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DOCTOR OF PHILOSOPHY IN PHYSICAL OCEANOGRAPHY

from the

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ABSTRACT

The inner continental shelf is generally defined as the region from the edge of depth-limited wave breaking to water depths of tens of meters. It is an important region because it is the connection between the coastal ocean and the surf zone. During the summer of 2015 and 2017, two major field experiments were conducted to study water temperature and currents along the central California inner shelf. A headland-and-bay topography characterizes the field site with stretches of sandy or rocky coastlines and two headlands. Vertical profiles of co-located ocean temperature and currents were obtained in a dense cross-shore array from the 50- to the 5-m isobaths during the 2015 experiment. During the 2017 field experiment, vertical profiles of temperatures were obtained in an alongshore array on the 9-m isobath that spanned 55 km of coastline. The results show the contributions from coastal upwelling, sea breeze-driven processes, and semidiurnal internal waves to the net transport of heat onto the inner shelf. The first observations of warm internal tide bores propagating to the surf zone, and the spatial and temporal scales of the temperature variability over 55 km of coastline are discussed. The Office of Naval Research funded this research.

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Acknowledgments

I would like to thank my advisor, Jamie MacMahan, for his guidance during the PhD process. Every day was a challenge, and Jamie made it fun every day. I hope I have the chance to get my hands dirty with some fieldwork in the future. Jamie is quite simply the best scientist I have worked with.

Committee members, co-authors, and faculty from both the Department of Oceanography and the Department of Meteorology were instrumental in providing invaluable mentorship throughout the process, and I am grateful for their help and guidance. I am grateful for the many scientific discussions with Tim Stanton. His perspective and input made my work better. I would like to extend a special thanks to Keith Wyckoff, whose technical expertise, vast field knowledge, and incredible attitude made every trip to the field a great experience.

I am grateful for the mentorship and, more importantly, the friendship of CDR Mathias Roth and LCDR Casey Gon. I have benefitted greatly from discussions with both of them, both professional and personal, and hope our paths cross again. They have made me a better naval officer.

To my wife, Genevieve, and children, Julian and Ione, thank you for allowing me to be absorbed in my research over the past 5 years. You sacrifice so much, and I am always at a loss to show how much it means to me. My parents and sister have always been there for me, and continue to be the shining example of hard work, dedication, and true Penguins fans. I am truly lucky to have my family and friends in my life. This work is really dedicated to them.

Finally, I would like to thank the Office of Naval Research for funding the Inner Shelf Departmental Research Initiative, and allowing me to be a part of such an amazing experiment.

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CHAPTER 1:

Introduction

The inner shelf is defined as the region between the seaward edge of depth-limited wave breaking and where the surface and bottom boundary layers overlap (Lentz and Fewings 2012). This region connects the outer continental shelf to the coast; therefore transport across the inner shelf has important ecological impacts. It is was often considered too shallow for the oceanographers, and deep for the coastal scientists (Cudaback et al. 2005). Ekman transport (Austin and Lentz 2002), cross-shore wind-driven transport (Horwitz and Lentz 2014, 2016), and nonlinear internal waves (Lucas et al. 2011; Sinnett et al. 2018) have been found to drive onshore transport onto the inner shelf. Surf zone processes, such as rip currents, drive offshore transport onto the inner shelf (Suanda et al. 2016). Previous inner shelf studies have typically focused on currents and circulation (e.g., Lentz and Fewings 2012, and references therein) or seawater temperature (Sinnett et al. 2018) at single timescales and at single locations or cross-shore transects to describe processes on the inner shelf.

The Office of Naval Research sponsored a multi-institution Inner Shelf Dynamics Departmental Research Initiative (Inner Shelf DRI) 5-year field and numerical modeling experiment to investigate the exchange of seawater properties across the inner shelf over a range of spatial and temporal scales. The field site was along 54 km of the central California coast with complex bathymetry, such as canyons and headlands. The field portion of the experiment included both in situ and remotely sensed observations of seawater temperature. The field experiment was conducted over two periods: (1) a pilot exercise during the summer of 2015 and (2) the main experiment during the fall of 2017. It was expected that surf zone processes would dominate the cross-shore exchange owing to the weak cross-shore barotropic currents on the inner shelf; however, the relative contributions from Stokes drift, internal waves, mixing, and wind-driven transport were not known.

In this dissertation, a unique set of seawater temperature and current observations are presented from Inner Shelf DRI field experiments performed during the summer of 2015 and the fall of 2017. Seawater temperature is an effective tracer for studying transport owing to the natural variability in seawater temperature over both the horizontal and vertical.

In this current study, temperature variability, cross-shore heat flux, and the mechanisms driving the variability and flux are investigated at various temporal and spatial scales on the central California coast to investigate the relative contributions to cross-shore exchange from internal waves and wind-driven processes.

The body of this dissertation is composed of three journal papers submitted for peer review. Owing to the complexity of the observations and analyses, detailed background information and conclusions are provided within each chapter. The following is a brief outline and overview of the chapters presented in this dissertation.

In Chapter 2, heat fluxes on the mid- and inner shelf near Pt. Sal, California are investigated at subtidal, diurnal, and semidiurnal timescales. Inner shelf transport mechanisms have been identified in field (Horwitz and Lentz 2014, 2016; Lucas et al. 2011; Suanda et al. 2011) and modeling (Austin and Lentz 2002) studies. At subtidal timescales, Ekman transport drives cross-shore transport on the mid-shelf, which can extend onto the inner shelf under stratified conditions, and leads to cooling of inner shelf water. During times when upwelling favorable winds cease, pressure gradient flows drive heating on the inner shelf. At diurnal timescales, sea breeze and baroclinic tidal processes can drive transport. The baroclinic tide is the dominant mechanism for cross-shore transport at semidiurnal timescales. Here, cross-shore heat fluxes at subtidal, diurnal, and semidiurnal timescales are estimated from measures of temperatures and currents over the vertical from a dense cross-shore array from the 5 to the 50-m isobaths. A novel approach of using wavelet co-scalograms to estimate cross-shore heat fluxes is introduced. The contribution to the net heat storage owing to cross-shore heat fluxes and surface heat flux are estimated to identify the dominant forcing mechanisms in each frequency band.

In Chapter 3, observations of the semidiurnal internal tide are presented. Nonlinear internal waves, such as internal tide bores, are important transport mechanisms. Near-bottom, cold boluses have been observed to transport cold water from offshore to the coast (Sinnott et al. 2018). The observations presented here are the first observations of warm, internal tide bores propagating to the edge of the surf zone. Warm internal tide bores induce mixing on the inner shelf. Alongshore differences in mixing owing to differences in bathymetry are estimated with temperature overturns.

In Chapter 4, temperature variability is investigated in 9 to 16-m water depths along

54 km of the inner shelf between Pismo and Pt. Arguello, California. Patterns in the subtidal, diurnal, and semidiurnal temperature fluctuations are described through principal component analyses over both the horizontal for depth-averaged temperature and over the vertical for profiles of temperature. Decorrelation length scales are estimated with lagged cross-correlation functions on scales ranging from 150 m to 54 km. These results are compared with observations of diurnal and semidiurnal band temperature fluctuations in the Santa Barbara Channel, located approximately 20 km south of the field site presented here.

In Chapter 5, a summary of the cross-shore heat flux and temperature variability is provided. The heating and cooling of the inner shelf owing to concomitant processes at various time- and spatial scales, as well as the interaction amongst these scales are investigated as important drivers of nearshore hydrographic variability. The results presented here are expected to improve current understanding of the representativeness of remotely sensed inner shelf seawater temperature.

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CHAPTER 2:

Spatial and Temporal Contributions of the Cross-shore Heat Flux for the Midshelf and Inner shelf near Pt. Sal, California

This chapter has been submitted to the Journal of Physical Oceanography with minor formatting changes. As the main author of this work, I made the major contributions to the research and writing. Co-authors include Jamie H. MacMahan, John A. Colosi, Matt K. Gough, Sutara H. Suanda, and Nirnimesh Kumar.

2.1 Introduction

Cross-shore exchange often dominates the distribution and retention of nutrients, biota, pollution, and heat over the mid- and inner shelf (Brink 2016). The mechanisms driving this exchange occur owing to geophysical processes at subtidal ($f < 1/33$ cph), diurnal ($1/33 < f < 1/16$ cph), and semidiurnal ($1/16 < f < 1/10$ cph) timescales. At subtidal and diurnal timescales, winds and waves drive nearshore, cross-shore transport (Fewings et al. 2008; Hendrickson and MacMahan 2009; Lentz and Fewings 2012). At semidiurnal timescales, the baroclinic internal tide drives cross-shore transport. Because temperature is an effective tracer of cross-shore exchange on the mid- and inner shelf, heat budgets are a practical technique to quantify the cross-shore exchange.

Subtidal (hereafter, ST) heat budgets on the mid- and inner shelf have predominantly focused on the heating and cooling that result from wind- and wave driven circulations. For upwelling conditions along the northern California coast, a ST heat budget (over 105 days) indicated that the largest terms in the mean heat budget were net surface heat flux into the ocean and offshore heat flux that was a result of offshore flow of warm water in the upper 30 m of the water column (Lentz 1987). Furthermore, offshore fluctuations in the cross-shore heat flux were consistent with wind-driven upwelling. Send et al. (1987) observed nearshore warming during times of upwelling relaxation when upwelling favorable winds

decay or reverse. They summarized the heating in three ways: (1) the heating was not a simple reversal of the cooling from the upwelling response where near-surface warm water returned to the coast; (2) the observed, transient cross-shore heat flux did not account for the observed warming; and (3) the warming was consistent with the poleward, alongshore advection of relatively warm water. A mean, poleward, pressure gradient force sets-up between 35°N to 38°N along the California coast as a response to wind stress forcing (Hickey and Pola 1983). Superimposed on this mean alongshore pressure gradient are local pressure gradients generated by the interaction of flow with topographic heterogeneities such as headlands (Gan and Allen 2002). Consistent with observations on the northern California coast, observations (Fewings et al. 2015; Melton et al. 2009; Washburn et al. 2011) and models (Suanda et al. 2016) from near Pt. Conception on the central California coast show a poleward-propagating coastal buoyant plume that increases nearshore water temperature by as much as 5°C during upwelling wind relaxation. At these locations and at ST-frequencies, the cross-shore, alongshore, and surface heat flux all contribute to the mid- and inner shelf heat budgets.

Processes at diurnal (hereafter, DU) frequencies, such as sea-breeze driven responses (Hendrickson and MacMahan 2009; Lerczak et al. 2001; Nam and Send 2013), tidal mechanisms such as diurnal period internal waves (DPIWs, Beckenbach and Terrill 2008), and thermally driven exchanges (Herdman et al. 2015; Molina et al. 2014; Monismith et al. 2006), contribute to a DU heat budget. In northern Monterey Bay, California, Suanda et al. (2011) found that the alongshore component of the sea breeze resulted in a diurnal upwelling response that cooled the inner shelf waters, followed by subsequent warming on the inner shelf as the sea breeze ceased. A similar response was observed in San Luis Bay, California (Walter et al. 2017). Lerczak et al. (2001) and Nam and Send (2013) found that the sea breeze generated a resonant response whereby DPIWs are generated and freely propagate offshore. Poleward of the critical latitude of 30°N, DPIWs will only freely propagate away from the generation site if there is a sufficient background vorticity or the correct sign or shear (Kunze 1985). An effective inertial frequency results from the background vorticity, and allows for a resonant response poleward of the critical latitude (Beckenbach and Terrill 2008; Lerczak et al. 2001). Exchange owing to differential heating (thermally-driven) have been found to be an important mechanism on the inner shelf of tropical reef systems where the wind- and wave driven circulations tend to be less important (Molina et al. 2014).

At semidiurnal (hereafter, SD) frequencies, the internal tide is the dominant process that contributes to the heat budget through the onshore transport of either near-surface warm water or near-bottom cold water (Boehm et al. 2002; Kumar et al. 2016; Lucas et al. 2011). The semidiurnal internal tide can be generated locally when the barotropic tide interacts with a stratified shelf break or continental slope (Huthnance 1989; Noble et al. 2009). This interaction leads to a baroclinic structure in the currents (Lamb 2014; Woodson 2018) that can transport larvae (Pineda 1999), heat (Boehm et al. 2002; Lucas et al. 2011; Nam and Send 2013), and nutrients (Lucas et al. 2011; Omand et al. 2012) across the inner shelf as the internal tide shoals. Sinnett and Feddersen (2014) and Sinnett et al. (2018) observed temperature fluctuations within the surf zone that they interpreted as the arrival of cold, internal tidal bores that contributed a negative (offshore) cross-shore heat flux to the surf zone heat budget. Colosi et al. (2018a) and Freismuth et al. (submitted) observed warm internal tidal bores from the 50–5-m isobaths at Pt. Sal, California.

The distinction between warm bores of depression or cold bores of elevation propagating across the shelf is a result of either the background stratification, background currents, or a combination of both (Lamb 2014). Internal tide bores generated at the shelf break are generally waves of depression that perturb the near surface pycnocline to deeper depths and draw warmer, less dense water into the interior of the water column (Lamb 2014). As the pycnocline depth increases to greater than 50% of the water depth, a wave of depression reverses polarity to become a wave of elevation (Helfrich 1992; Shroyer et al. 2009). The waves of elevation draw near-bottom, dense water into the interior of the water column. Cold internal bores can also be generated by subthermocline waters being pumped out of deep canyons, that then propagated across the shelf (Walter and Phelan 2016). Owing to the interaction of bathymetry, stratification, and background currents with the internal tide as it shoals across the continental shelf (Nash et al. 2012), the induced cross-shore exchange on the inner shelf is inherently nonstationary.

In the present, the depth-integrated heat budget is estimated from cross-shore and surface heat fluxes for a cross-shore line that extend from the 50-m isobath to coast and from the 20-m isobath to the coast near Pt. Sal, California for ST-, DU-, and SD-bands. The alongshore heat flux cannot be estimated from this data, but is addressed as a residual to the heat balance. Mean budgets and fluctuating budgets show the dominant contributions to the heat storage in each band. Cross-shore heat fluxes in the DU- and SD-bands are

calculated using a wavelet co-scalogram technique to account for the nonstationarity of the signal. Finally, the tidal band (DU- and SD-band) cross-shore heat fluxes and surface heat flux variability are related to longer timescale (60 hours) nearshore heat storage.

2.2 Field Experiment

2.2.1 Field Site

Temperatures, ocean currents, and meteorological data were obtained near the Pt. Sal headland on the central California coast from 16-Jun-2015 to 22-Jul-2015 at part of the Pt. Sal Inner Shelf Experiment (PSIEX, Figure 2.1). Pt. Sal is a rocky headland located approximately 55 km north of Pt. Conception, CA. North of the rocky headland, the sandy coastline is oriented 14.5° east of north (Figure 2.1a). Mussel Pt. is a small, rocky promontory 3 km north of Pt. Sal. Mussel Pt. (Figure 2.1a) extends 350 m in the alongshore direction, and protrudes 350 m offshore. Centered approximately 3 km offshore of Mussel Pt., there is a nearly circular submerged rocky outcrop approximately 2 km in diameter. The outcrop shoals to a depth of 15 m from an offshore depth of 30 m.

2.2.2 Observations and Methods

Ocean currents were obtained with 7 bottom-mounted, upward-looking Acoustic Doppler Current Profilers (ADCPs) deployed in a cross-shore array and colocated with thermistor string (T-string) moorings in water depths $h = 7$ (X02), 11 (X04), 15 (X06), 20 (X08), 30 (X10), 40 (X11), and 50 (X12) m (Figure 2.1 and Table 2.1). Throughout this manuscript, moorings with co-located ADCP and T-string will be denoted by their name and the water depth in parentheses. The ADCPs in $h < 20$ m were sampled with 2 MHz Nortek Aquadopps that sampled at 1 Hz with 0.5-m bins. The ADCPs at X06(15m) and X08(20m) were 1 MHz Nortek Aquadopps that sampled at 1 Hz with 1.0 m bins. The remaining ADCPs were 300 kHz RDI systems that had 5-s sample rate with 2-m bins.

Two different methods are used to quality control the near-surface ADCP bins. For the deep moorings ($h \geq 30$ m), 2 ADCP bins (4 m) below the tidal surface are removed from the velocity profile. Surface gravity wave amplitudes can be a sizable fraction of the shallow-water depths ($h \leq 20$ m). Significant wave heights were therefore calculated from co-located pressure measurements using linear wave theory (Dean and Dalrymple

Table 2.1. PSIEX moorings

Moorings	Water Depth (m)	Mean Sensor Depths (m)	Sample Rate T/ADCP (s)	ADCP Bin Size (m)	Location (Lat,Lon)
X01	5	0.5, 1.5, 2.5, 3.5, 4.5	1.0/1.0	—	34.9249,-120.6658
X02	7	0.5, 1.5, 2.5, 3.5, 4.5, 5.5, 6.5	1.0/1.0	0.5	34.9238,-120.6665
X03	9	0.5, 1.5, 2.5, 3.5, 4.5, 5.5, 6.5, 8.0	1.0/1.0	—	34.9240,-120.6671
X04	11	1.5, 2.5, 3.5, 4.5, 5.5, 6.5, 8.0, 9.5, 11	1.0/1.0	—	34.9244,-120.6695
X05	13	1.5, 2.5, 3.5, 4.5, 5.5, 7.0, 9.0, 10.5, 12.0	1.0/1.0	—	34.9243,-120.6702
X06	15	0.5, 1.5, 2.5, 3.5, 5.0, 6.5, 8.0, 10.0, 14.0	1.0/1.0	1.0	34.9246,-120.6737
X07	17	0.5, 3.5, 5.5, 7.5, 9.5, 11.5, 13.5, 16.0	1.0/1.0	—	34.9250,-120.6750
X08	20	0.5, 3.0, 6.0, 9.0, 12.0, 15.0, 17.0, 19.0	1.0/1.0	1.0	34.9253,-120.6771
X09	16	1.5, 3.0, 4.5, 6.0, 8.0, 10.0, 12.5, 15.0	1.0/1.0	—	34.9265,-120.6875
X10	30	6.7, 11.6, 16.6, 21.6, 26.5	30.0/5.0	2.0	34.9311,-120.6965
X11	40	5.1, 10.2, 20.1, 30.0, 35.0	30.0/5.0	2.0	34.9326,-120.7115
X12	50	6.2, 16.3, 29.7, 36.1, 46.1	30.0/5.0	2.0	34.9368,-120.7339
N50	50	—	—/5.0	2.0	34.9673,-120.7318
LR50	50	—	—/5.0	2.0	34.9069,-120.7272
S50	50	—	—/5.0	2.0	34.8795,-120.7264
N30	30	—	—/5.0	2.0	34.9580,-120.6878
LR30	30	—	—/5.0	2.0	34.9073,-120.6940
S30	30	—	—/5.0	2.0	34.8795,-120.6930
8A	8	—	—/1.0	0.5	34.9084,-120.6714
8B	8	—	—/1.0	0.5	34.9343,-120.6673

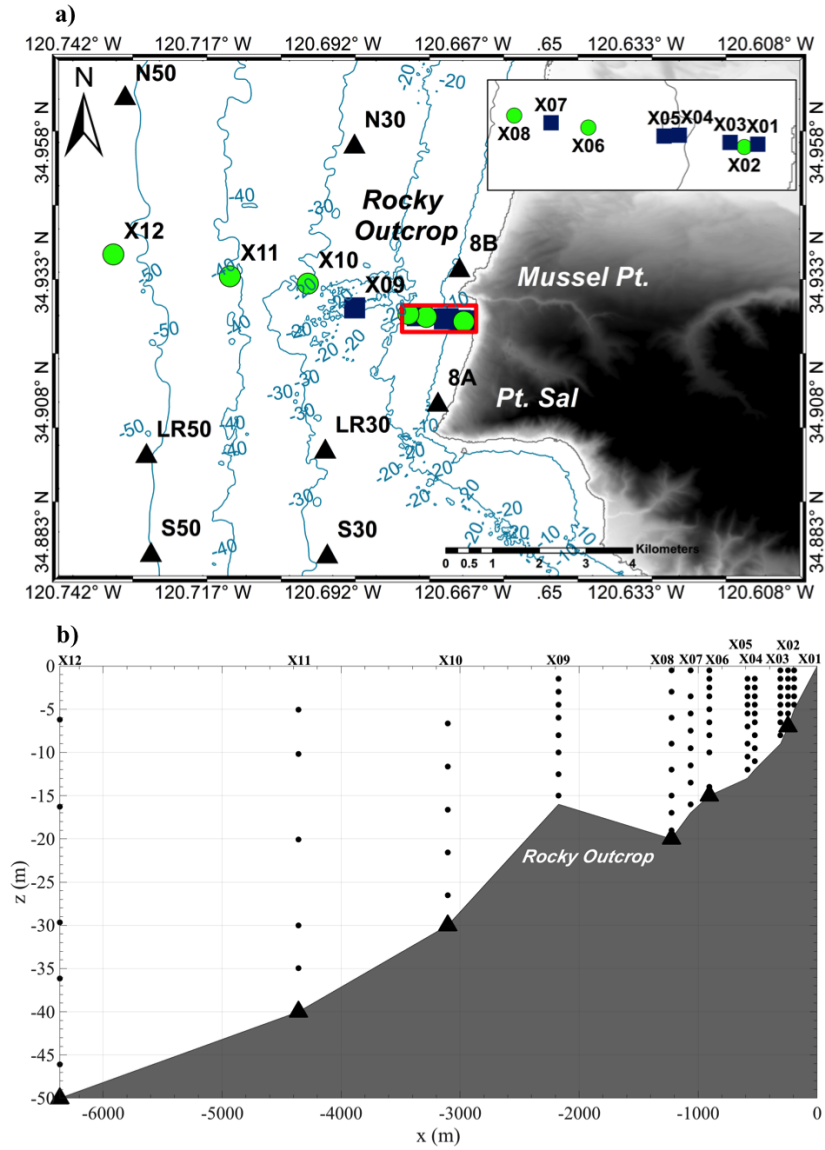


Figure 2.1. a) Map of experiment site where green dots indicate co-located T-strings and ADCPs, blue squares represent T-strings, black triangles indicate ADCPs, and the black dot indicates the on-site meteorological station. The red box is the location of the inset of the zoomed region. c) Instrument layout in the cross-shore array where black dots indicate thermistors and black triangles indicate ADCPs. Mooring names are labeled along the top of (b).

1991), and ADCP bins located one bin below wave trough and above are removed from the

shallow-water velocity profiles.

T-string moorings were deployed in a cross-shore array of 12 moorings in water depths from 5 m to 50 m, and are referred to as the X array (Figure 2.1 and Table 2.1). Temperatures were sampled at 1 Hz with RBR SoloT thermistors at moorings X1 - X09. For moorings X10(30m), X11(40m), and X12(50m), data were collected with SBE9 thermistors that sampled every 30 seconds. Thermistors were at various depths on each mooring (Figure 2.1b and Table 2.1). Thermistor data were hourly averaged, temporally interpolated to a common time with the velocity profiles, and spatially interpolated to the same depth-normalized scale as the velocity profiles to estimate cross-shore heat flux terms.

The velocity profiles were hourly averaged and depth-normalized to remove tidal modulation allowing for depth integration and comparison between time records longer than a few hours (Hendrickson and MacMahan 2009; Suanda et al. 2011). The normalized scale was set to $z/h = -1.0$ at the sea bed to $z/h = 0$ at the sea surface with 0.05 vertical increments, where z is the elevation. To extend the profile throughout the full water column, the top, good velocity bin value is repeated to the sea surface, and a no-slip bottom boundary condition was imposed at the sea bed, such that the bottom ADCP bin was linearly interpolated to zero velocity at the sea bed (Suanda et al. 2011). To analyze the background circulation, the velocity profiles were low passed filtered (subtidal, $f < 1/33$ cph) and depth averaged. Background vorticity ζ , is estimated from the depth-averaged, subtidal currents in two different manners. First, following Lerczak et al. (2001),

$$\zeta \approx \frac{\bar{v}_2^{ST} - \bar{v}_1^{ST}}{\Delta x}. \quad (2.1)$$

Here, $\bar{v}_2^{ST}$ and $\bar{v}_1^{ST}$ are the depth-averaged, subtidal alongshore velocities at X10(30m) and X12(50m), respectively. Second, Saucier (1955) and Molinari and Jr. (1975) show that horizontal divergence is equivalent to the fractional time rate of change of the horizontal area, A , of a parcel, such that

$$\frac{\partial \bar{u}^{ST}}{\partial x} + \frac{\partial \bar{v}^{ST}}{\partial y} = \frac{1}{A} \frac{dA}{dt}, \quad (2.2)$$

where $\bar{u}^{ST}$ and $\bar{v}^{ST}$ are the subtidal, cross- and alongshore depth-averaged, subtidal currents,

respectively. Upon a 90° clockwise rotation of the velocity vectors where $\bar{u} \rightarrow -\bar{v}'$ and $\bar{v} \rightarrow \bar{u}'$. Substituting these rotations into Eq. 2.2, the fractional time rate of change of a new horizontal area, A' , is equal to ζ , such that

$$\frac{\partial \bar{v}^{ST}}{\partial x} - \frac{\partial \bar{u}^{ST}}{\partial y} = \frac{1}{A'} \frac{dA'}{dt}. \quad (2.3)$$

Davies-Jones (1993) applied Bellamy's method to calculate the differential kinematic properties (DKP) of divergence and vorticity from measures of wind speed and direction from fixed weather stations. Following the treatment of Davies-Jones (1993), the N30, N50, LR30, and LR50 moorings (Figure 2.1a) define the corners of a polygon. The area of this polygon is given by

$$A = \frac{\Lambda(x, y)}{2}, \quad (2.4)$$

where, for a four-sided polygon,

$$\Lambda(l, g) \equiv (l_1 g_2 - l_2 g_1 + l_2 g_3 - l_3 g_2 + l_3 g_4 - l_4 g_3 + l_4 g_1 - l_1 g_4), \quad (2.5)$$

and l and g are any two variables. Hence, the fractional time rate of change of (2.4) is equivalent to the horizontal divergence as in Eq. 2.2, and is given by

$$\frac{\Lambda(\bar{u}, y) + \Lambda(x, \bar{v})}{\Lambda(x, y)} = \frac{1}{A} \frac{dA}{dt}. \quad (2.6)$$

With the rotations described above, ζ can be estimated. The second technique is referred to as the DKP method in this study. The DKP method was also performed using N30, N50, S30, and S50, and resulted in similar estimates of ζ .

For the cross-shore heat flux estimates, a local coordinate system was defined at each ADCP mooring based on the principal axis direction of the hourly-averaged velocity. The principal axes were typically aligned with the local bathymetry and coastline. A regular Cartesian coordinate system was defined such that the onshore direction was positive x , and the poleward, alongshore direction was positive y . Currents were decomposed into the baroclinic component by subtracting the depth-averaged current from the observed current such that $u'(z, t) = u(z, t) - \overline{u(t)}$, where u' is the baroclinic current, $u(z, t)$ is the total current, $\overline{u(t)}$ is the depth-averaged or barotropic current, and t is time.

Measures of hourly wind were obtained at the National Data Buoy Center (NDBC) station 46011 located approximately 30-km offshore of Pt. Sal. Here, wind direction, θ is reported as “direction toward” where $\theta=0^\circ$ are winds to the East. The hourly measures of wind are interpolated to a common time with the temperature and currents. Surface heat flux, Q_o , is estimated as a sum of the incident shortwave solar radiation, sensible, and latent heating, and longwave radiation, such that

$$Q_o = Q_{sw} + Q_{sen} + Q_{Lv} + Q_{lw}.$$

Data for the surface heat flux estimate were obtained approximately 30 km north of Pt. Sal at Port San Luis and Morro Bay, California (Walter et al. (2017), data available at www.slosea.org/about/archive.php). Bulk flux algorithms were used to estimate Q_{sen} and Q_{Lv} (Fairall et al. 2003), Q_{lw} is estimated according to Dickey et al. (1994) in a similar manner as Connolly and Lentz (2014).

2.2.3 Heat Budget and Cross-Shore Heat Flux

The heat equation for a two-dimensional cross-shore slice that extends from the coast at $x = 0$ to an offshore extent at $x = -L$, is bounded by the sea surface and sea floor, and assumes zero cross-shore flux at $x = 0$ has been derived in previous studies (e.g., Austin 1999; Dever and Lentz 1994; Suanda et al. 2011), and can be written as

$$\begin{aligned} \rho c_p \int_{-L}^0 dx \int_{-h}^0 \frac{\partial T}{\partial t} dz &= \rho c_p \int_{-h}^0 uT \Big|_{x=-L} dz \\ &\quad - \rho c_p \int_{-L}^0 dx \int_{-h}^0 \left(T \frac{\partial v}{\partial y} + v \frac{\partial T}{\partial y} \right) dz + \int_{-L}^0 Q_o dx. \end{aligned} \quad (2.7)$$

Here, ρ is the density of sea water, c_p is the specific heat of sea water at constant pressure, T is the water temperature, u and v are the horizontal velocities in the x (cross-shore) and y (alongshore) directions, and Q_o is the net air-sea heat flux. The total heat storage in the two-dimensional slice (HS , left-hand side of Eq. 2.7) is controlled by the cross-shore heat flux (CHF, first term on the right-hand side of Eq 2.7), the alongshore heat flux (second term), and surface heat flux (SHF, third term). Dever and Lentz (1994) decomposed u , v , and T into spatial average and perturbation quantities, u' , v' , and T' . The spatial average is over depth for the cross-shore components, for example $T'(z, t) = T(z, t) - \bar{T}(t)$, where

$\bar{T}(t)$ is the depth-averaged temperature. Unfortunately, the alongshore heat flux cannot be estimated in this study because the alongshore gradients in T and v are not known inshore of the 30-m isobath, and is therefore neglected in this analysis. Hence, the heat balance used in this study is the contribution of CHF and SHF to HS , such that

$$\rho c_p \int_{-L}^0 dx \int_{-h}^0 \frac{\partial T}{\partial t} dz = \rho c_p \int_{-h}^0 u'T' \Big|_{x=-L} dz + Q_o L + F_{res}. \quad (2.8)$$

where $F_{res} = HS - CHF - SHF$ is the residual of the balance (similar to the approach of Sinnett and Feddersen 2014). The implications of neglecting the alongshore heat flux are addressed in the *Discussion* section. HS is estimated with trapezoidal integration where each temperature observation is weighted by the area defined as half the distance to an adjacent instruments or the sea surface, sea bed, or offshore boundary at $x = -L$ (Austin 1999; Dever and Lentz 1994; Suanda et al. 2011). A mean heat budget over the duration of the experiment is performed using the depth-integrated, unfiltered HS , CHF , and SHF . To estimate the fractional contribution of CHF and SHF to HS in the ST-, DU-, and SD-bands, each term is band (DU and SD) or low-pass (ST) filtered. A linear regression is performed on the bin meaned HS and the bin meaned sum of the CHF and SHF , such that $HS = \alpha(CHF + SHF) + b$. The y -intercept represents the experimental mean flux or heat storage, which is included in the subtidal analysis.

To analyze the spatial and temporal evolution of CHF in the ST, DU, and SD bands, two different techniques are used: a wavelet co-scalogram approach for the DU-and SD-bands, and low-pass filter for the ST-band. The continuous wavelet transform is an effective technique to analyze non-stationary time series to identify a signal's time and frequency, f , distribution (Torrence and Compo 1998; Walter et al. 2017). In previous studies, momentum flux was calculated through a wavelet co-scalogram that identifies the co-varying signals at a given frequency and time (Hudgins et al. 1993; Laxague et al. submitted). Here, this wavelet co-scalogram technique is extended to calculate the depth-integrated, CHF , such that

$$CHF(t, f) = \rho c_p \int_{z=-h(t)}^0 \text{Re}(W_u^* W_{T'}) h(t) d(z/h), \quad (2.9)$$

where ρ is the density of sea water, c_p is the specific heat capacity of water, W is the wavelet transform, and $*$ is the complex conjugate. The Morlet wavelet is used as the mother wavelet in each wavelet transform calculation (Torrence and Compo 1998). The depth-integrated

cross-shore heat flux at each mooring is scale-averaged for DU and SD such that,

$$CHF^{BP}(t) = \frac{\delta j \delta t}{C_\delta} \sum_{j=j_1}^{j_2} \frac{CHF(t, s_j)}{s_j}, \quad (2.10)$$

where the *BP* superscript refers to either DU or SD, δj is the scale resolution, δt is the sampling time interval, $C_\delta = 0.776$ is a reconstruction constant, and s_j is the wavelet scale (Torrence and Compo 1998). The wavelet scale is analogous to the Fourier frequency. The subtidal, ST ($f < 33^{-1}$ cph), depth-integrated heat flux is calculated as

$$CHF^{ST}(t) = \rho c_p \int_{z=-h(t)}^0 (u'T')^{ST} h(t) d(z/h), \quad (2.11)$$

where $(u'T')^{ST}$ is the low-pass filtered ($f < 33^{-1}$ cph) *CHF* that includes the experimental mean. This technique is used for ST owing to the errors at the beginning and end of the wavelet co-scalogram for longer periods representative of ST frequencies (Torrence and Compo 1998). Here, it is assumed that total *CHF* is sum total of the CHF^{BP} and CHF^{ST} . This assumption is true if and only if CHF^{BP} and CHF^{ST} is averaged over an integer number of waves, and there is no physical process that couples two different frequencies (cf. Aristizabal et al. 2017). The decorrelation timescale is determined from the autocorrelation function of the hourly-averaged, depth-integrated, unfiltered $u'T'$. The decorrelation time at X12(50m) and X08(20m) is ~ 60 hours. Performing a moving average on CHF^{BP} and CHF^{ST} using a 60-hr window allows for direct comparison of CHF^{BP} in each band and CHF^{ST} , and an estimate of how DU-band and SD-band contribute to the total flux over long (> 60 hr) timescales.

2.3 Results

The background conditions at Pt. Sal are characterized by periods of upwelling favorable winds that persist for two to four days, and periods of relaxations when the wind stress, τ_s , tends to 0 Pa (Figure 2.2a). ST-band τ_s is averaged over five-day intervals, $\langle \tau_s \rangle$, between yeardays 167 and 202 to summarize the low-frequency wind forcing (I-VII in Figure 2.2a and Table 2.2). Maximum $\langle \tau_s \rangle$ of 0.10 Pa occurs during period I and has a mean orientation toward $\langle \theta^{ST} \rangle = -56.0 \pm 3.0^\circ$ relative to east. For reference, onshore winds for the region between Mussel Pt. and Pt. Sal are oriented $\sim \theta = -15.0^\circ$. Between periods II

and IV, $\langle \tau_s \rangle$ decreases from 0.07 Pa to 0.01 Pa, while the direction rotates to more of an onshore direction. From periods V to VI, $\langle \tau_s \rangle$ increases to 0.07 Pa, and becomes upwelling favorable again. Period VI is similar to period II, however the variability in mean wind orientation, σ_θ is an order of magnitude less during period VI. Winds relax during period VII. Following Fewings et al. (2008), weak winds are defined as $\tau_s < 0.03$ Pa. Significant wave height, H_s , during PSIEEX varied between 0.7 and 1.8 m with mean $H_s = 1.1$ m, and increased during periods of increased τ_s (Figure 2.2b). Net air-sea heat flux, Q_o varied between approximately -250 and 1100 W m⁻², and has a consistent pattern during the experiment with a peak at midday (Figure 2.2c). Here, positive Q_o indicates a net flux into the ocean. Incident solar radiation dominates the Q_o signal.

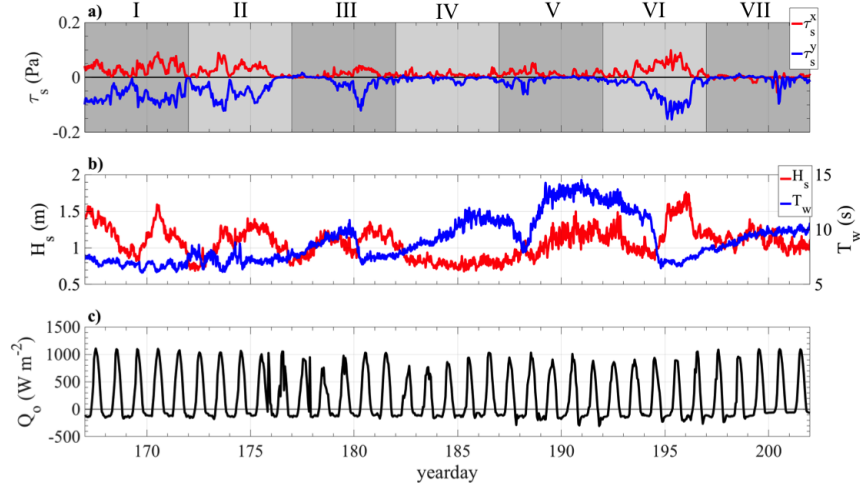


Figure 2.2. a) Cross- (τ_s^x , red line) and alongshore wind-stress (τ_s^y , blue line) at NDBC46011. Positive values indicate northward and eastward directions. The gray-shaded areas represent averaged periods I-VII that are summarized in Table 2.2. b) Significant wave height (H_s , red line) and wave period (T_w , blue line) during PSIEEX. c) Surface heat flux, Q_o over the course of the experiment. Positive values indicate net flux into the sea surface.

The inner shelf is identified when the combined thickness of the surface (δ_s) and bottom (δ_b) boundary layers are greater than the water depth (h). Weatherly and Martin (1978) model the thickness of the surface boundary layer as

$$\delta_s = C_1 \sqrt{\frac{\tau_s}{\rho_o}} \frac{1}{f_i (1 + \frac{N^2}{f^2})^{1/4}}, \quad (2.12)$$

Table 2.2. Wind stress and direction averaged over 5-days and standard deviations.

Period	Yeardays	$\langle \tau_s^{ST} \rangle$	σ_τ	$\langle \theta^{ST} \rangle$	σ_θ
I	167-172	0.10	0.03	-55.7	3.0
II	172-177	0.07	0.04	-53.3	23.6
III	177-182	0.03	0.03	-48.0	53.4
IV	182-187	0.01	0.01	-35.9	13.2
V	187-192	0.02	0.01	-34.0	16.1
VI	192-197	0.07	0.06	-52.4	4.7
VII	197-202	0.005	0.004	-93.0	78.2

where C_1 is a dimensionless fitting parameter, τ_s is the wind stress, ρ_o is the density of seawater, $f_i = 21^{-1}$ cph is the inertial frequency, and N is the buoyancy frequency. The value of C_1 has been estimated from 1.0 (Lentz 1992) to 2.3 (Tilburg 2003). Here, C_1 is assumed to be 2.3, consistent with Fewings et al. (2008); Hendrickson and MacMahan (2009); Tilburg (2003). Fewings et al. (2015) estimated the bottom boundary layer thickness by replacing τ_s with bottom stress, $\tau_b = \rho_o r \mathbf{u}_{da}$, where $r = 5 \times 10^{-4} \text{ m s}^{-1}$ is a linear resistance coefficient, and $\mathbf{u}_{da}$ is the depth-averaged velocity. The boundary layer thicknesses are subtidal filtered to establish the mean inner shelf depth. If $h < \delta_s + \delta_b$, then the mooring is considered to be on the inner shelf, and the circulation is governed by inner shelf dynamics (Fewings et al. 2008, 2015). Owing to the dependence on τ_s , τ_b , and N^2 , the width of the inner shelf varies by ~ 6 km with a mean width, $\langle L_{IS} \rangle \sim 1.9$ km over the experiment. Mooring X10(30m) is in the inner shelf less than 60% of the experiment duration and X11(40m) and X12(50m) are in the inner shelf less than 25% of the experiment duration. These moorings are considered to be in the midshelf. Moorings X02(7m), X06(15m), and X08(20m) are in the inner shelf greater than 90% of the time and are considered to be in the inner shelf.

The mean, depth-averaged circulation on the midshelf is consistent with previous observations near Pt. Conception on the central California coast (Fewings et al. 2015). However, the mean inner shelf circulation differs from previous observations on the inner shelf near Pt. Conception, California, where Cudaback et al. (2005) and Fewings et al. (2015) found that the mean inner shelf flow was directed equatorward. The circulation pattern inshore of the 20-m isobath at Pt. Sal reveals a consistent pattern (Figure 2.3), where north of Mussel

Pt., currents at mooring 8B are directed down-coast and vary between 0.05 and 0.1 m s^{-1} . Currents veer shoreward at X08(20m) and X06(15m) immediately south of Mussel Pt., and X02(7m) shows a cyclonic circulation in the lee of Mussel Pt. This mean inner shelf flow is likely indicative of the influence of the Pt. Sal headland on the inner shelf circulation. The depth-averaged currents between the 30-m and 50-m isobaths highlight the influence of the synoptic wind forcing. During period I, when the upwelling favorable winds are developing, the currents at the 30 and 50-m isobaths north of Pt. Sal are generally weak ($\sim 0.01 \text{ m s}^{-1}$) and directed onshore (Figure 2.3a). The currents at the LR and S-moorings south of Pt. Sal are stronger ($0.05\text{-}0.1 \text{ m s}^{-1}$), and indicate a possible anticyclonic circulation. During period II, the upwelling favorable winds are developed and the currents north of Pt. Sal are directed down-coast consistent with a coastal upwelling response (Figure 2.3b). South of Pt. Sal, the currents are half those during period I. As the winds relax during Period III, the currents are directed poleward with the strongest currents ($\sim 0.12 \text{ m s}^{-1}$) along the 50-m isobath and decreasing to $\sim 0.05 \text{ m s}^{-1}$ along the 30-m isobath (Figure 2.3c). This poleward flow during upwelling relaxation is consistent with previous observations (Cudaback et al. 2005; Fewings et al. 2015; Washburn et al. 2011) and modeling studies (Suanda et al. 2016) of the region. The poleward flow continues until upwelling favorable winds begin to reverse the flow (Figure 2.3e) and down-coast currents fully develop during period VI (Figure 2.3f) with upwelling favorable wind forcing. Near the end of the experiment, the upwelling winds relax again, and the flow is directed poleward (Figure 2.3g). These periods of mean circulation are the background on which the processes that drive cross-shore heat flux at ST, DU, and SD timescales are superposed.

Temperatures and currents are similar across the X-array, and observations from X08(20m) are shown as an example of the temperature and current trends during the experiment (Figure 2.4a). Times of relatively colder water are present during periods II, and VI. These times are characterized by upwelling favorable winds (Figure 2.2a). The period of warm water between year days 177–181 is consistent with poleward advection of warm water from the Santa Barbara Channel during upwelling relaxation (Melton et al. 2009; Suanda et al. 2016; Washburn et al. 2011). Temperature spectra were calculated at $z/h = -0.2$ with a 144-hr spectral window with 50% overlap resulting in 32 degrees of freedom (Figure 2.4c). Significant (at the 95% significance level) peaks at the $f_{DU} = 1/24 \text{ cph}$ and $f_{M2} = 1/12.42 \text{ cph}$. At these higher frequencies, temperatures vary by as much as 3.5°C and occur

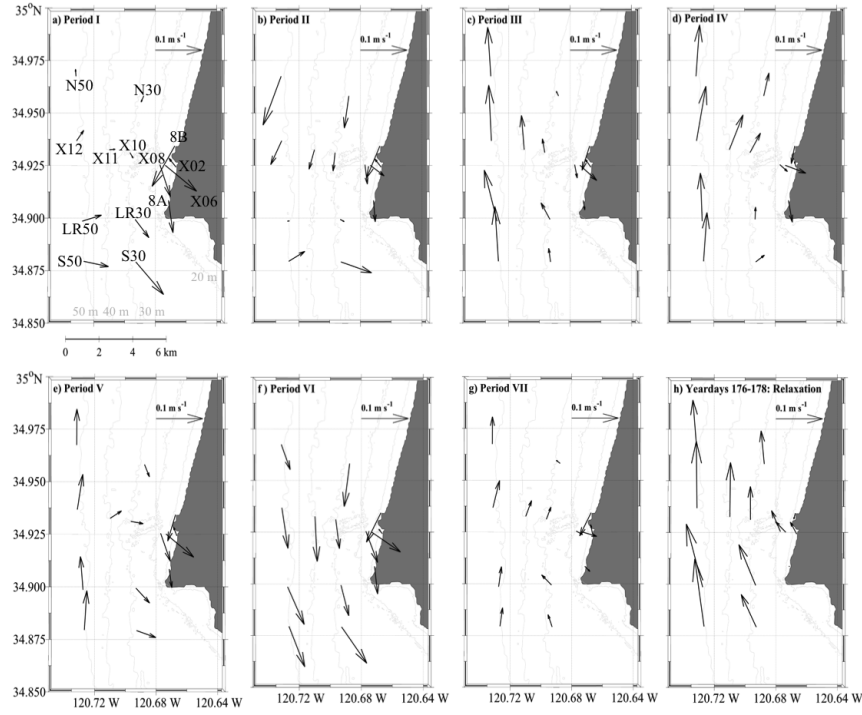


Figure 2.3. a-g) Depth-averaged currents during periods I-VII (Table 2.2). h) Depth-averaged currents during the relaxation of upwelling favorable winds (year days 176–178 at the beginning of period III, Figure 2.2a).

throughout the entire water column at X08(20m) (Figure 2.4a).

The cross-shore baroclinic currents exhibit a mode-1 structure with velocities as high as 0.12 m s^{-1} at X08(20m). The currents are faster at X12(50m) and decrease toward the coast at X02(7m). The vertical structure of the ST, DU, and SD-band baroclinic currents (u', v') is characterized at each mooring in the X-array with a complex empirical orthogonal function (CEOF) analysis (Kumar et al. 2016; Kundu and Allen 1976). In the ST-band, a first mode baroclinic velocity explains 50% of the variance at X12(50m), and increases onshore to 89% at X02(7m) (table 2.2). A first mode baroclinic velocity explains between 76–86% of the variance in the DU-band, and 87–90% of the variance in the SD-band, consistent with a study of diurnal and semidiurnal internal motions in water depths between 26–8 m at San Pedro Bay, California (Kumar et al. 2016). In the SD-band, the percent of variance explained by a first mode decreases from X12(50m) to X02(7m).

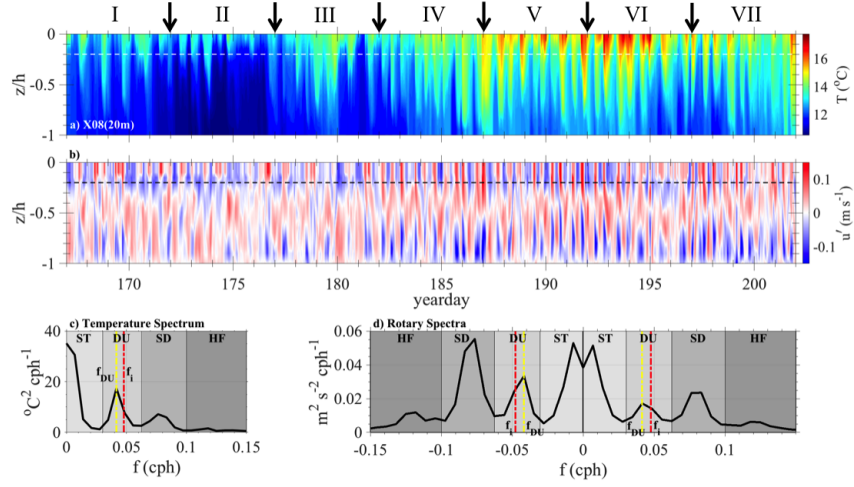


Figure 2.4. a) Example of hourly-averaged temperature, T over the vertical plotted against time at X08(20m). b) Example of baroclinic, cross-shore currents, u' , at X08(20m). The trends in the temperature and current are similar across the X-array from $h = 5$ -50 m. Arrows above the plots indicate the periods I-VII (Table 2.2). c) Spectrum of temperature at $z/h = -0.2$ (dashed, white line in [a]). The gray-shaded regions represent the ST-, DU-, SD- and HF-bands. The $f_{DU} = 1/24$ cph frequency is marked with a vertical yellow, dashed line, and $f_i = 1/20.9$ cph is marked with a vertical red line. d) Rotary spectra for the $z/h = -0.2$ (dashed, black line in [b]). Negative frequencies represent clockwise rotation and positive frequencies represent counterclockwise rotations (Thomson and Emery 2014). The gray-shaded regions and vertical lines are similar to those in (c).

Rotary spectra of u' and v' separate the currents into clockwise (CW) and counterclockwise (CCW) motions (Thomson and Emery 2014). Spectra were calculated at $z/h = -0.2$ with a 144-hr spectral window with 50% overlap resulting in 32 degrees of freedom. The DU-band CW rotary variance (negative frequencies in Figure 2.4d) is approximately 2.3 times greater than CCW variance at X08(20m). This trend is consistent at each of the moorings in the X-array with the exception of X02(7m) where the ratio of the CW to CCW variance is approximately equal to 1. The factor of 2 in the ratio of the CW to CCW rotary variance is similar to results from a previous study that looked at diurnal period internal motions forced by the sea breeze (Lerczak et al. 2001). In the SD-band, the rotary spectra peak at the M2 tidal frequency (Figure 2.4d).

Cross-shore heat flux, CHF , varies both spatially and temporally (Figure 2.5). The exper-

Table 2.3. Percent variance of first mode baroclinic velocity CEOF for the total record and for subtidal, diurnal, and semidiurnal frequencies.

Mooring	Subtidal	Diurnal	Semidiurnal
X12(50m)	50	76	90
X11(40m)	60	86	93
X10(30m)	70	86	90
X08(20m)	67	76	86
X06(15m)	78	82	88
X02(7m)	89	81	87

imental mean CHF^{ST} ranges from -178.14×10^4 to -2.95×10^4 W m⁻¹ from X12(50m) to X02(7m), and is anomalously positive at X08(20m) (Table 2.4). Negative CHF^{ST} occurs on the midshelf during periods I, II, and VI when winds are upwelling favorable (Figure 2.6a). During periods of weak winds, positive CHF^{ST} occurs over both the mid- and inner shelf (black lines on Figure 2.5). In the DU-band, CHF^{DU} is intermittent. Negative CHF^{DU} occurs on the midshelf during periods I and VI, and a time of positive CHF^{DU} is found near yearday 176 during period IV (Figure 2.5b). On the inner shelf, CHF^{DU} is generally positive with pulses occurring at yeardays 171, 180, and 187 (periods I, III, and VI). Experimental mean CHF^{DU} is negative at X12(50m) and X11(40m), is positive at X10(30m) and X08(20m), and maximum at the 20-m isobath (Table 2.3). Experimental mean CHF^{SD} decreases from a maximum of 237.90×10^4 W m⁻¹ at X12(50m) to -0.91×10^4 W m⁻¹ at X02(7m) (Table 2.4). Pulses of positive CHF^{SD} occur from yeardays 167–170, 182–184, 186–194, and 197–202 during periods of weak or no CHF^{ST} or CHF^{DU} (Figure 2.6c).

2.4 Discussion

2.4.1 Synoptic Wind-Driven CHF^{ST}

On the midshelf, τ_s^y is more efficient than τ_s^x at driving cross-shore transport (Lentz and Fewings 2012). Whereas, on the inner shelf, τ_s^x is more efficient than τ_s^y at driving cross-shore transport (Fewings et al. 2008; Tilburg 2003). The mid- and inner shelf are characterized by times of positive and negative CHF^{ST} throughout PSIEEX (Figure 2.5). At

Table 2.4. Experimental mean values and standard deviations of cross-shore heat flux at in the subtidal, diurnal, and semidiurnal bands.

		X12(50m)	X11(40m)	X10(30m)	X08(20m)	X06(15m)	X02(7m)
Subtidal	Mean	-17.81	-24.28	-8.37	3.24	-1.29	-0.29
	Std. Dev	46.04	35.87	24.33	6.72	4.53	0.92
Diurnal	Mean	-2.91	-0.95	0.91	1.70	0.06	-0.07
	Std. Dev	11.60	5.38	2.96	3.07	1.62	0.24
Semidiurnal	Mean	23.79	22.73	6.12	5.71	2.02	-0.09
	Std. Dev	36.31	26.09	8.08	19.33	3.26	0.34

Units are in $\times 10^5 \text{ W m}^{-1}$.

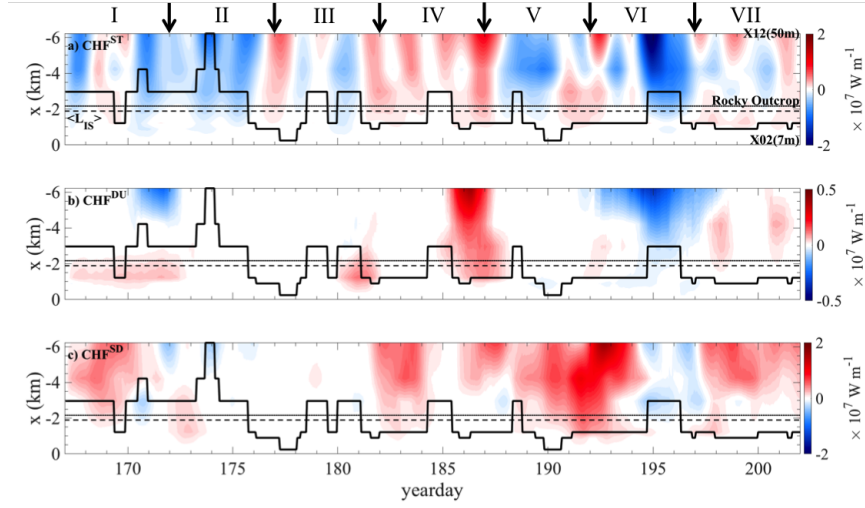


Figure 2.5. Cross-shore heat flux in the a) ST-band, b) DU-band, and c) SD-band plotted against time from the coast to $h = 50$ m. Arrows above the plots indicate the periods I-VII (Table 2.2). The solid, black line indicates the width of the inner shelf. The horizontal, dashed, black line indicates the mean inner shelf width, $\langle L_{sz} \rangle$. The cross-shore location of the rocky outcrop is marked by the horizontal, dotted, black line.

subtidal timescales, the cross-shore heat flux is wind-driven. While the strongest negative CHF^{ST} occurred during times of the largest τ_s , the direction did not vary significantly (at the 95% significance level) between periods of positive and negative CHF^{ST} (Figure 2.6). Wind-stress is conditionally averaged during periods of negative or positive CHF^{ST} ($-\langle CHF^{ST} \rangle$ and $+\langle CHF^{ST} \rangle$, respectively, and $\langle \cdot \rangle$ represents the conditional averaging) to estimate the difference in forcing during each flux regime at each mooring in the X-array. For times when $-\langle CHF^{ST} \rangle, \langle \tau_s \rangle > 0.04$ Pa (blue line in Figure 2.7a). From X12(50m) to X02(7m), the magnitude of $-\langle CHF^{ST} \rangle$ tends to decrease with a gradient of 806 W m^{-2} ($r^2=0.97$) from a maximum of $-4.4 \times 10^6 \text{ W m}^{-1}$ (blue line, Figure 2.7b). For times when $+\langle CHF^{ST} \rangle, \langle \tau_s \rangle < 0.04$ Pa (red line in Figure 2.7a). From X12(50m) to X02(7m), the magnitude of $+\langle CHF^{ST} \rangle$ tends to decrease with a gradient of 379 W m^{-2} ($r^2=0.98$) from a maximum of $2.4 \times 10^6 \text{ W m}^{-1}$ (red line, Figure 2.7b). This asymmetry in the negative and positive CHF^{ST} indicates unique processes driving CHF^{ST} , rather than a simple reversal of one process (Send et al. 1987).

Times of negative CHF^{ST} (Figure 2.5a) at Pt. Sal are consistent with wind-driven coastal

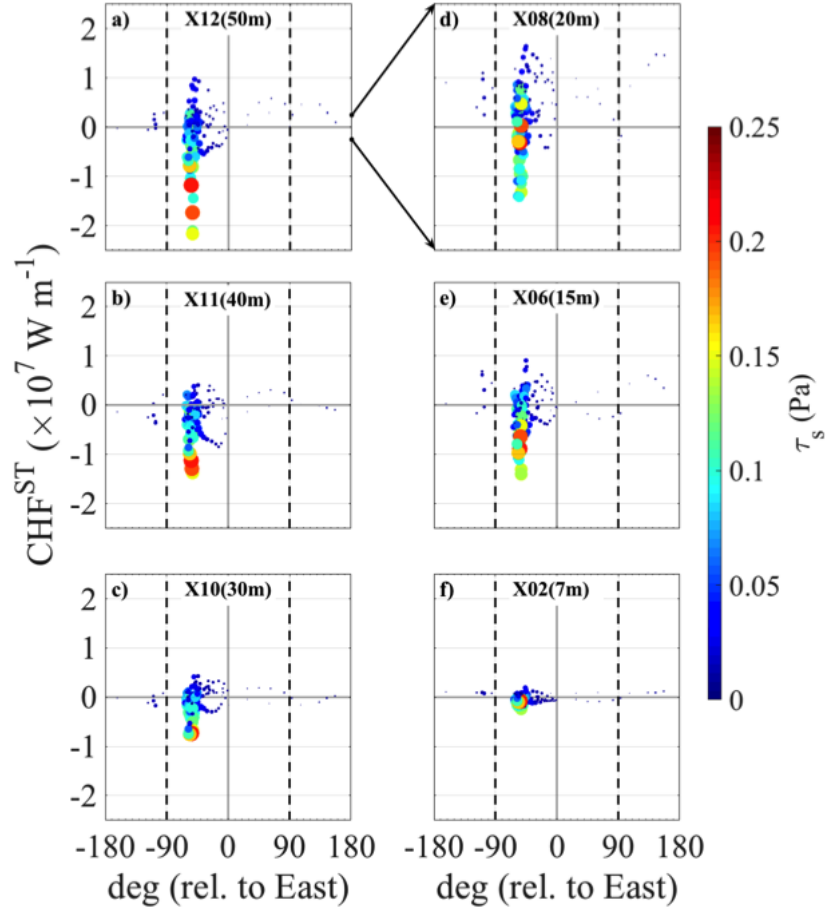


Figure 2.6. Cross-shore heat flux in the ST-band (CHF^{ST}) plotted as a function of wind-direction relative to East from a) X12(50m) to f) X02(7m). The magnitude of the total wind stress (τ_s) is indicated by both the color and size of the dots. Larger, warmer-colored dots indicate stronger wind stress. Zero CHF^{ST} is marked by a horizontal, black line. Directions between -90° and $+90^\circ$ (vertical, dashed black lines) are indicative of a onshore component of the wind stress, and directions between 0° and -90° are indicative of an upwelling favorable component of the wind stress. The y-axis scale for (a-c) is an order of magnitude larger (-2.5×10^6 – $2.5 \times 10^6 \text{ W m}^{-1}$) than the y-axis for (d-f). The arrows from (a) indicate the relative range of CHF^{ST} in (d).

upwelling (Figure 2.2a), and are similar to the findings of Lentz (1987). On the midshelf, the offshore transport of near-surface, warm water and the compensating onshore transport of mid- to near-bottom, cold water (Figure 2.8a) both act to create a negative CHF^{ST} during periods I, II, III, V, and VI (Figure 2.8c). Austin and Lentz (2002) found during times of

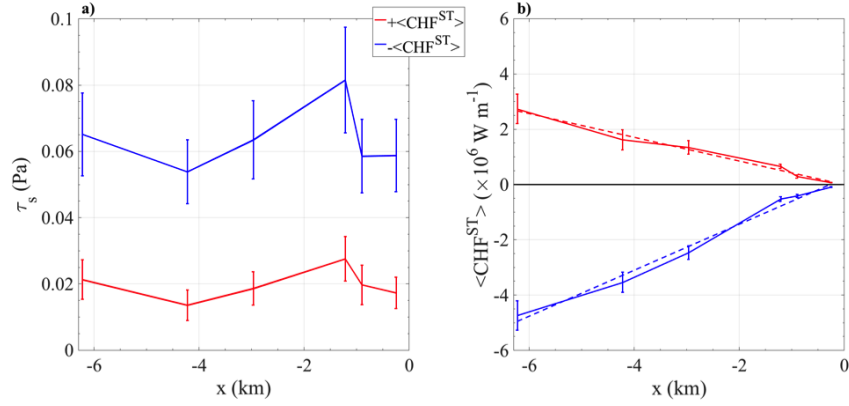


Figure 2.7. a) Wind stress averaged over times of positive (red line) and negative (blue line) ST-band, cross-shore heat flux ($+\langle CHF^{ST} \rangle$ and $-\langle CHF^{ST} \rangle$, respectively) plotted as a function of cross-shore location. b) Conditionally averaged ST-band, cross-shore heat flux plotted against cross-shore location during times of positive (red line, $+\langle CHF^{ST} \rangle$) and negative (blue line, $-\langle CHF^{ST} \rangle$) fluxes. The dashed lines represent fits from a linear regression analysis, and the zero flux line is marked with a horizontal, solid black line. Vertical lines represent the 95% confidence intervals at each point for both plots.

continuous stratification on the inner shelf, that a weak upwelling circulation extended into the inner shelf. During PSIEX, the inner shelf was typically stratified during upwelling conditions (Figure 2.8b), and CHF^{ST} was negative (Figure 2.5) on the inner shelf during times of upwelling favorable winds (Figure 2.2a). Upwelling circulation is the primary mechanism that drives negative CHF^{ST} on both mid- and inner shelf at Pt. Sal.

Positive CHF^{ST} is observed during times of wind relaxation (Figure 2.7). Upwelling winds create a negative alongshore pressure gradient between Port San Luis, CA, and Santa Monica, CA (Fewings et al. 2015; Melton et al. 2009; Suanda et al. 2016; Washburn et al. 2011). Furthermore, equatorward currents interacting with coastline heterogeneities, such as the Pt. Sal headland, create local poleward pressure gradients south of coastal irregularities along the California coast (Gan and Allen 2002). When upwelling favorable winds relax, the negative alongshore pressure gradients drive poleward flow (Figure 2.3h). The relaxation that occurs between yeardays 176–178 is unique for this data set in that the poleward flow advects warm water from the Santa Barbara Channel in the form of a

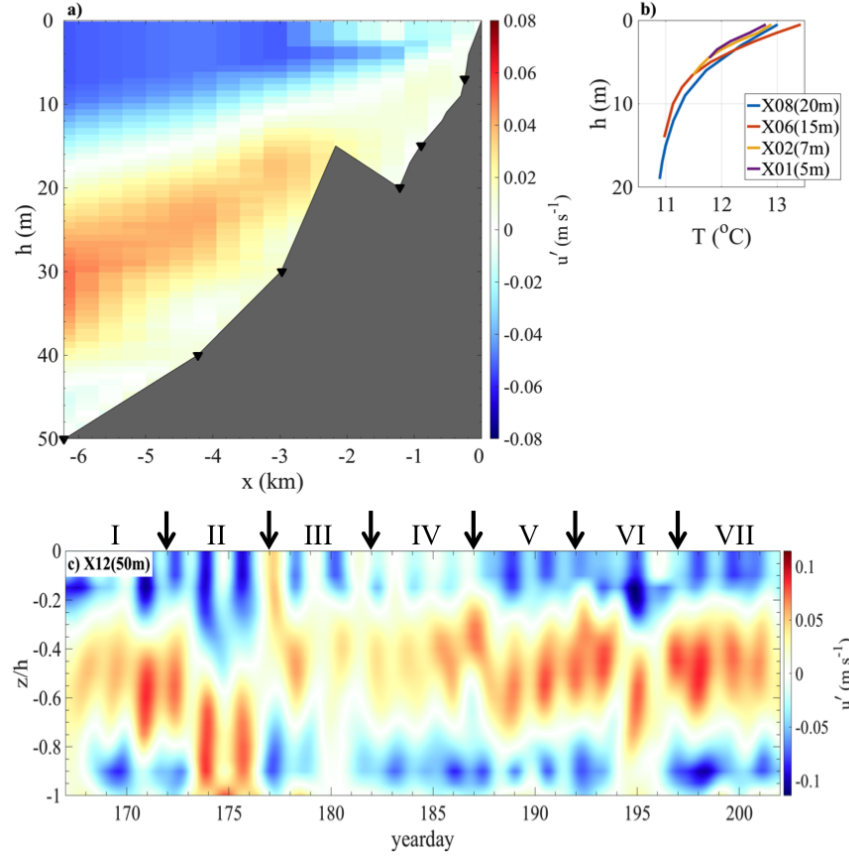


Figure 2.8. a) Cross-shore currents (shading) during upwelling favorable winds during period II. Locations of ADCPs are marked with black triangles. b) Time-averaged temperature profiles over the vertical at X08(20m), X06(15m), X02(7m), and X01(5m) during period II. c) Baroclinic, cross-shore currents at X12(50m) in the ST-band. Arrows above the plot indicate periods I-VII (Table 2.2).

coastal buoyant plume that extends to Pt. Sal (Fewings et al. 2015; Melton et al. 2009; Suanda et al. 2016; Washburn et al. 2011). Washburn et al. (2011) observed cross-shore circulation with near-surface onshore currents and near-bottom currents that persisted while temperature increased. This two-layer circulation contributes to a positive CHF^{ST} between year days 176–178. A poleward flow persists on the midshelf through period IV, consistent with a seasonal mean poleward alongshore pressure gradient on the US West Coast (Hickey and Pola 1983). The alongshore pressure gradient creates a mid-water column onshore geostrophic current as large as 0.02 m s^{-1} (periods III and IV in Figure 2.8c) on the midshelf

(Fewings et al. 2015; Lentz 2008). There is evidence of anticyclonic circulation on the inner shelf between Pt. Sal and Mussel Pt. (Figure 2.3d,e). The positive CHF^{ST} could be a result of an eddy or headland wake generated by the poleward flow interacting with Pt. Sal (Black et al. 2005; Signell and Geyer 1991). The mechanisms contributing to positive CHF^{ST} include the circulation associated with a coastal buoyant plume, geostrophic response to an alongshore pressure gradient, and anticyclonic circulation resulting from flow around coastal promontories.

2.4.2 Sea Breeze Driven CHF^{DU}

Wind-driven CHF^{DU} is manifested as diurnal Ekman transport (Suanda et al. 2011; Woodson et al. 2007) on the midshelf, down-wind forced currents on the inner shelf (Kaplan et al. 2003), or as a resonantly forced diurnal period internal waves (DPIWs) in the nearshore (Lerczak et al. 2001; Nam and Send 2013). Periods of increased diurnal wind stress (red line in Figure 2.9a) are consistent with periods of increased DU-band, baroclinic, depth-averaged kinetic energy, KE_{DU} (blue line in Figure 2.9a) on both the mid- and inner shelf. On the inner shelf, the observed impulses of CHF^{DU} are positive (Figure 2.5b). This positive CHF^{DU} is consistent with previous research that found that the diurnal sea breeze drives a near-surface, onshore circulation on the inner shelf (Hendrickson and MacMahan 2009) similar to the wind-driven cross-shore transport at subtidal time scales (Fewings et al. 2008). Because these periods of increased diurnal wind occur during upwelling favorable winds, τ_s^x is acting across a negative cross-shore density gradient which increases the efficiency of the wind-driven transport (Horwitz and Lentz 2014). Walter et al. (2017) studied the diurnal temperature patterns in Port San Luis, California, 30 km north of Pt. Sal, during the same time period. They found that the diurnal temperature variability was a result of the sea breeze advecting the upwelling front into the bay. On the midshelf, CHF^{DU} is negative during periods II and VI and corresponds with the periods of upwelling favorable winds and negative CHF^{ST} (Figure 2.5a,b). Woodson et al. (2007) found that an alongshore, diurnal sea breeze drove a diurnal Ekman transport that tended to cool the northern Monterey Bay inner shelf. In a similar manner, the diurnal, alongshore wind during these periods are enhancements to the coastal upwelling circulation that drives a negative CHF^{ST} .

Lerczak et al. (2001) found that the diurnal sea breeze can induce a diurnal period internal wave poleward of the critical latitude (30°N) given a sufficient negative background vorticity.

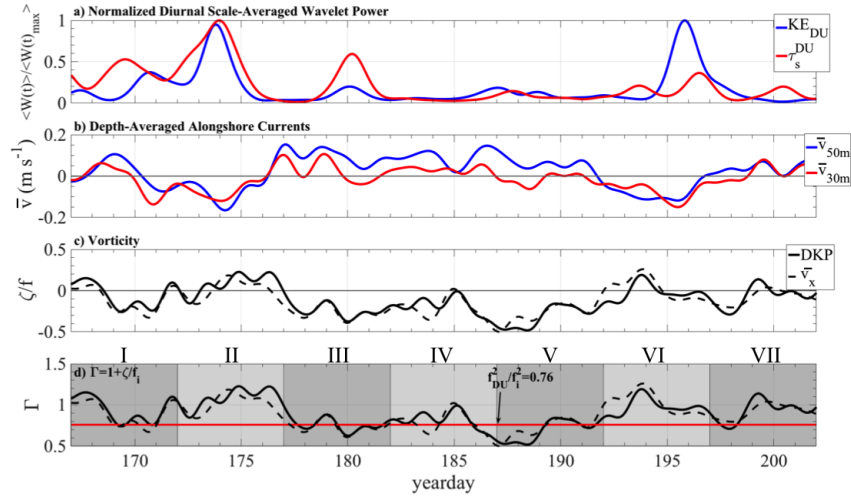


Figure 2.9. a) Scale-averaged wavelet power in the DU-band for the depth-averaged, baroclinic kinetic energy (blue line, KE_{DU}) and wind stress (red line, τ_s^{DU}) plotted as a function of time. The scale-averaged wavelet power is normalized by the maximum power for each variable. b) Depth-averaged, alongshore currents at the 50- and 30-m isobaths (blue line, $\bar{v}_{50m}$ and red line, $\bar{v}_{30m}$, respectively) plotted as a function of time. c) Vorticity (ζ) scaled by the inertial frequency (f_i) estimated from the depth-averaged, alongshore currents (dashed, black line) in a similar manner as Lerczak et al. (2001) and from the DKP calculation between the N and LR moorings (solid line) described in the text. d) The dynamical variable $\Gamma = 1 + \zeta/f_i$ plotted as a function of time. When $\Gamma \leq f_{DU}^2/f_i^2 = 0.76$ (horizontal, red line), diurnal period internal waves can freely propagate. The dashed line represents Γ calculated by the estimate of ζ from the depth-averaged currents, and the solid line represents Γ calculated from the estimate of ζ from the DKP calculation. The gray-shaded regions depict periods I-VII (Table 2.2).

The background vorticity creates an effective inertial frequency, $f_{eff} = f_i + \zeta$, where ζ is the subtidal vorticity. They defined the quantity Γ , such that

$$\Gamma = f_{eff}/f_i = 1 + \zeta/f_i. \quad (2.13)$$

The estimates for ζ using the depth-averaged alongshore currents at X12(50m) and X11(30m) (Eq. 2.1 and Figure 2.9b) or using the DKP method yield consistent results (Figure 2.9). When $\Gamma \leq f_{DU}^2/f_i^2$, the sea breeze can generate diurnal period internal waves poleward of the critical latitude. At Pt. Sal, the threshold for resonance is

$f_{DU}^2/f_i^2 = 0.76$. The system is near resonance at times of wind relaxation during periods III and IV–V (Figure 2.9d), when the background circulation is directed poleward, there is horizontal shear that leads to anticyclonic circulation (Figures 2.3c and e). Concomitant with the increase in KE_{DU} are times of positive CHF_{DU} , that are likely a result of nearshore DPIWs. At yearday 180.5, there is a positive CHF^{DU} that evolves offshore, but dissipates by yearday 182 (Figure 2.5b) when $\Gamma > 0.76$. At yearday 186, a CHF^{DU} propagates onshore from X12(50m) (Figure 2.5b).

The mechanisms that drive CHF^{DU} are the onshore component of the wind which drives a positive CHF^{DU} on the inner shelf, while the alongshore component of the wind acts to enhance the upwelling circulation on the midshelf, resulting in a negative CHF^{DU} . These observations are also consistent with wind-driven DPIWs propagating both on- and offshore with a positive CHF^{DU} on the mid- and inner shelf. Times of favorable background ζ occur during upwelling wind relaxations and allow DPIWs to propagate freely. Resonantly forced diurnal period internal waves drive a positive CHF^{DU} on both the mid- and inner shelf. The background circulation tends to dictate the DU-band wind-driven response on both the midshelf and inner shelf.

2.4.3 Semidiurnal Internal Tide Driven CHF^{SD}

Semidiurnal internal tidal bores (ITBs) drive CHF^{SD} (e.g., Boehm et al. 2002; Kumar et al. 2016; Lucas et al. 2011). Colosi et al. (2018a) performed a statistical study of internal waves during PSIEX with temperature observations from X08(20m), X10(30m), X11(40m), and X12(50m). They found that warm bores associated with ITBs propagated onshore with a group speed of $\sim 0.18 \text{ m s}^{-1}$ at X12(50m) and 0.08 m s^{-1} at X08(20m), and the frequency of bore arrival was modulated by the background stratification. The warm bores dissipated over an e-folding distance of approximately 2.4 km, had an energy dissipation rate of approximately 144 W m^{-1} , and were not observed during upwelling owing to the reduction in stratification. During PSIEX, CHF^{SD} is typically positive and extends on the inner shelf (Figure 2.3c). Freismuth et al. (submitted) showed that near-surface ITBs propagated to the edge of the surf zone. The decrease in CHF^{SD} from X12(50m) to X06(7m) (Figure 2.5c) is consistent with the energy flux propagation loss reported by Colosi et al. (2018a). Lucas et al. (2011) estimated an areal-averaged temperature change owing to the experimental mean CHF^{SD} from the 20-m isobath to the coast to be $0.28^\circ\text{C day}^{-1}$. At Pt. Sal, the net

increase in temperature, for a similar area, owing to CHF^{SD} is $1.0^{\circ}\text{C day}^{-1}$.

Although CHF^{SD} tends to be positive during PSIEX, there are times of negative CHF^{SD} (yeardays 170.5, 188, and 194–197.5 in Figure 2.5c). The transition from positive CHF^{SD} at X12(50m) to negative CHF^{SD} at X10(30m) suggests the tide bores are undergoing a polarity reversal as they shoal between the 50-m and 30-m isobaths (Helfrich 1992; Shroyer et al. 2009). After yearday 185, the ST-band thermocline extends deeper into the water column, and likely exceeds the critical depth of $\sim 0.5h$ (Figure 2.10). Between yeardays 195 and 197.5, the polarity reversal likely occurs in $h > 50$ m. The negative CHF^{SD} does not extend past the rocky outcrop (dotted line in Figure 2.5). This could be result of increased dissipation as the bores shoal across the outcrop (Freismuth et al. submitted).

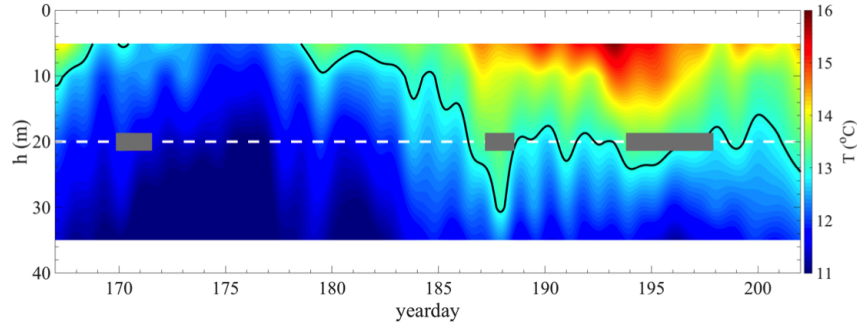


Figure 2.10. Subtidal temperature at X11(40m) plotted over the vertical. The solid, black contour represent the average isotherm over the experiment ($T = 12.9^{\circ}\text{C}$). The horizontal, dashed, white line marks the nominal critical depth at 50% of h . The gray-shaded boxes indicate times when CHF^{SD} at X10(30m) is negative.

2.4.4 Heat Budget

Two mean heat budgets are estimated: one from the 50-m isobath to the coast and one from the 20-m isobath to the coast. The former includes both the mid- and inner shelf, while the latter includes only the inner shelf. Because the alongshore heat flux (AHF) terms cannot be estimated inshore of the 20-m isobath, the experimental mean heat balance is estimated as

$$\langle HS \rangle = \langle CHF \rangle + \langle SHF \rangle + \langle F_{res} \rangle. \quad (2.14)$$

Here, $\langle \cdot \rangle$ represents the temporal mean over the duration of the experiment. For the control volume that extends from the 50-m isobath to the coast, the largest terms of the heat budget

are $\langle CHF \rangle = -17.9 \times 10^5 \text{ W m}^{-1}$ and $\langle SHF \rangle = 13.8 \times 10^5 \text{ W m}^{-1}$ (Table 2.5). This balance is similar to that observed during upwelling conditions on the northern California coast during the CODE2 experiment (Lentz 1987). Large fluctuations in $\langle HS \rangle$ are dominated by fluctuations in $\langle CHF \rangle$ (Table 2.5). For the inner shelf control volume, that extends from the 20-m isobath to the coast, $\langle CHF \rangle$, $\langle SHF \rangle$, and $\langle F_{res} \rangle$ are all of the same magnitude, and $\langle F_{res} \rangle = -5.7 \times 10^5 \text{ W m}^{-1}$ is the largest magnitude mean flux (Table 2.5). Here, it is speculated that F_{res} is representative of AHF . Lentz (1987) described poleward heat fluxes on the inner shelf that were related to wind relaxations that tended to heat the nearshore. At Pt. Sal, negative $\langle F_{res} \rangle$ is consistent with a poleward advection of a negative temperature gradient. This is consistent with the coastal buoyant plume described by Melton et al. (2009), Washburn et al. (2011), and Suanda et al. (2016) discussed above, as well as a response to the mean poleward pressure gradient force along the United States West Coast (Hickey and Pola 1983) given the generally weak wind ($\langle \tau_s \rangle \sim 0.04 \text{ Pa}$) forcing during PSIEX. Like the heat budget extending to the 50-m isobath, the largest fluctuations occur with the $\langle CHF \rangle$ and likely drive the observed fluctuations in $\langle HS \rangle$ (Table 2.5).

The slope of the linear regression analysis between the fluctuating, bin meaned HS and bin meaned $CHF + SHF$ describes the fractional contribution of $CHF + SHF$ to HS . For the control volume extending to the 50-m isobath, $CHF + SHF$ describes 94% ($r^2 = 0.82$) of HS in the ST-band that includes the experimental mean as well (Figure 2.11a). This implies that only 6% of the HS is a result of an alongshore heat flux. In the DU-band, 60% ($r^2 = 0.99$) of HS is described by $CHF + SHF$, and in the SD-band, 73% ($r^2 = 0.99$). These fits suggest that the alongshore heat flux contributes to 40% and 27% of HS in the DU- and SD-bands, respectively (Figure 2.11b,c). To isolate the contribution from SHF , it is removed from the analysis. The slopes do not change for the ST- and SD-bands, however, it is estimated that CHF^{DU} accounts for 49% of the flux, and suggests that SHF accounts for 11% of the total flux in the DU-band.

A similar analysis is performed for the inner shelf control volume that extends from the 20-m isobath to the coast. In the ST-band, the fraction of HS described by $CHF + SHF$ decreases from 94% at X12(50m) to 63% ($r^2 = 0.60$) at X08(20m) (Figure 2.11d). The positive y -intercept is indicative of the relatively large $\langle F_{res} \rangle$ discussed above, and suggests that the alongshore heat flux is acting to cool the inner shelf while CHF and SHF are acting to warm the inner shelf. In the DU-band, CHF and SHF account for 57% and 20%

Table 2.5. Experimental mean values and standard deviations of heat budget terms

		$\langle HS \rangle$	$\langle CHF \rangle$	$\langle SHF \rangle$	$\langle F_{res} \rangle$
X12(50m)	Mean	5.5	-17.9	13.8	-9.5
	Std. Dev	246.0	186.0	25.4	38
X08(20m)	Mean	0.3	3.3	2.7	-5.7
	Std. Dev	32.6	23.0	5.0	-6.4

Units are in $\times 10^5 \text{ W m}^{-1}$

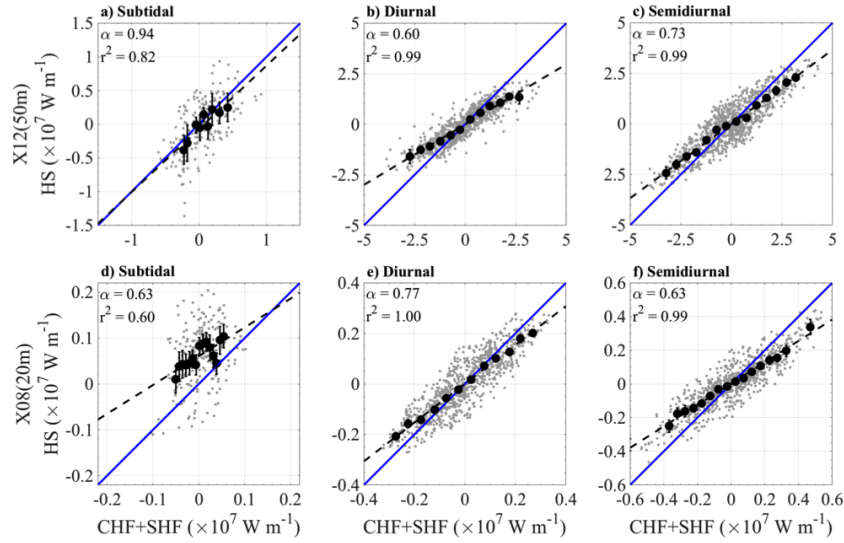


Figure 2.11. Heat budget for control volumes extending (top row) from the 50-m isobath to the coast and (bottom row) from the 20-m isobath to the coast in the subtidal (a,d), diurnal (b,e), and semidiurnal (c,f) bands. Black dots represent the binned-mean HS and the gray dots are the unbinned data. Black vertical bars represent the 95% confidence interval, and are smaller than the black dots at times. The dashed, black line is the linear fit to the data and the solid blue line is the 1-to-1 line for reference.

of the flux, respectively; suggesting the alongshore heat flux only accounts for 23% of the total flux (Figure 2.11e). In the SD-band, 63% ($r^2 = 0.99$) of the total flux is described by CHF (Figure 2.11f).

Based on the findings that CHF and SHF account for $> 50\%$ of the total flux in the

ST-, DU-, and SD-bands on both the mid- and inner shelf, it is instructive to compare the cumulative input of heat into the mid- and inner shelf at this site. Taking the 60-hr moving average of each of the CHF^{BP} and SHF terms allows for a direct comparison and summation of these fluxes. For the control volume extending to the 50-m isobath, CHF^{ST} and CHF^{SD} dominate the fluctuations while the SHF is a persistent input of flux (Figure 2.12a). Integrating CHF^{BP} and SHF provides an estimate of the cumulative input of heat into the volume per unit meter of coastline. Whereas the offshore flux associated with the upwelling circulation on the midshelf (CHF^{ST}) tends to remove heat from the volume over the course of the experiment, the persistent positive SHF and relatively large onshore CHF^{SD} from shoaling internal tide bores input heat into the volume. The cumulative effect results in a net removal of heat from the volume between yeardays 171 and 181, and a net input of heat from yearday 181 to 201 (black line in Figure 2.12b). On the inner shelf, the effect of CHF^{ST} is reduced; however, CHF^{SD} , CHF^{DU} , and SHF continually input heat into the inner shelf (Figure 2.12d). Subtracting the inner shelf heat from the heat in the volume extending to the 50-m isobath results in the heat on the midshelf (Figure 2.13). The tidal band (DU and SD-band) processes tend to add heat to the inner shelf while upwelling removes heat from the midshelf. During wind relaxations, the tidal band processes add heat to both the mid- and inner shelf. These findings suggest the tidal band processes on the Pt. Sal inner shelf drive gradients in the relative heating and cooling between the mid- and inner shelf that could have broad ecological impacts ranging from the transport of larvae to creating prime environments for nearshore, harmful algal blooms.

2.5 Summary

The relative contribution of the subtidal, diurnal, and semidiurnal depth-integrated, cross-shore heat flux and surface heat flux are observed on the mid- and inner shelf near the rocky headland of Pt. Sal on the central California coast with a cross-shore array of co-located measures of temperature and velocity profiles are presented in a cross-shore array spanning the 50-m to the 7-m isobath over the 35-day field experiment PSIEX. The depth-integrated, cross-shore heat flux, CHF , is driven by the wind at subtidal and diurnal timescales. Consistent with previous studies, the cross-shore transport on the midshelf is driven by the alongshore component of the wind stress, while the onshore component of the wind stress drives the cross-shore transport on the inner shelf. The cross-shore width

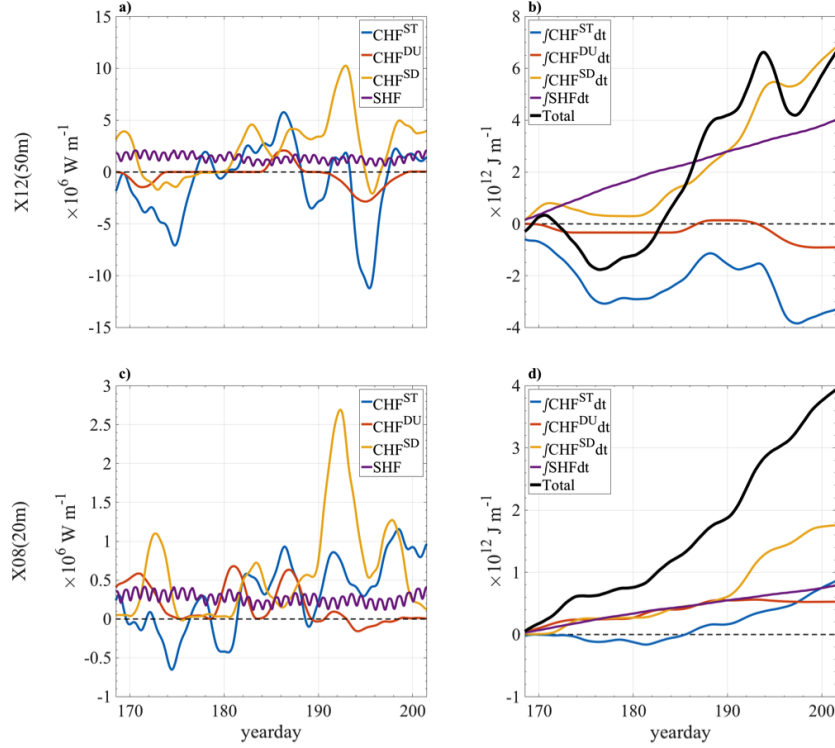


Figure 2.12. The time-averaged CHF^{BP} and SHF using a 60-hour moving average at (a) X12(50m) and (b) X08(20m). c and d) Time-integrated terms from (a) and (b), respectively. The solid black line indicates the sum total of each of the terms.

of the inner shelf ranged up to 6 km (7 to 50-m isobath), owing primarily to the variability in the subtidal, alongshore component of the wind stress.

During periods of upwelling favorable winds and continuous stratification, CHF^{ST} is negative and extends to the 7-m isobath. Wind relaxations are accompanied by positive CHF^{ST} as a likely result of alongshore pressure gradient driven flows. The diurnal sea-breeze drives a positive CHF^{DU} on the inner shelf through the onshore transport of a warm surface layer, yet drives a negative CHF^{DU} on the midshelf through a diurnal Ekman transport mechanism. Furthermore, sea breeze driven DPIWs can create a positive CHF^{DU} on the mid- and inner shelf. At semidiurnal timescales, CHF^{SD} is driven by the onshore propagation of warm, near-surface internal tidal bores. The tidal band processes drive heat fluxes at longer timescales (>60 hours) that create gradients in heating and cooling between the mid- and inner shelf. These findings have important implications for the cross-shore

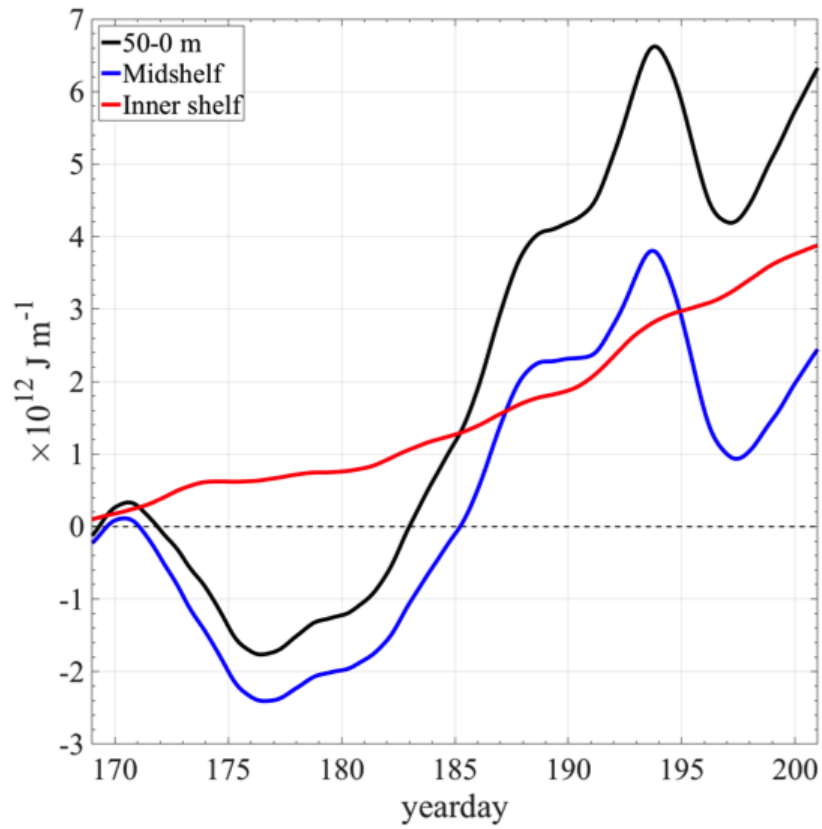


Figure 2.13. Time-integrated thermal energy in the control volume that extends from the 50-m isobath to the coast (black line), in the midshelf (blue line), and in the inner shelf (red line).

transport of heat, biota, or pollution in a coastal environment where the width of the inner shelf and dominant transport mechanisms evolve on timescales from days to seasons.

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CHAPTER 3:

Cross-shore Observations of Warm, Internal Tide Bores within the Nearshore

This chapter has been submitted to the Journal of Physical Oceanography with minor formatting changes. As the main author of this work, I made the major contributions to the research and writing. Co-authors include Jamie H. MacMahan, John A. Colosi, and Matt K. Gough.

3.1 Introduction

Nonlinear internal waves (NLIWs) are an important mechanism for transporting nutrients, biota, and temperature across the coastal ocean (Woodson 2018). NLIWs are either waves of elevation, referred to as boluses, or waves of depression, referred to here as internal tide bores (ITBs), where their generation is dependent on the background shear, stratification, and bathymetry (Lamb 2014). Recent work has highlighted the onshore transport of cold, nutrient rich water owing to shoaling boluses in the nearshore (e.g., Nam and Send 2011; Sinnett et al. 2018; Smith et al. 2016; Walter et al. 2012; Woodson 2018). Sinnett et al. (2018) used a dense, cross-shore array to observe boluses that extended directly into the surf zone decreasing water temperatures by as much as 1.7°C. Shoaling boluses also modify the nearshore by creating turbulence and mixing that acts to erode stratification and can re-suspend particulate matter on the sea floor (Woodson 2018).

Lamb (1994) developed the first generation and propagation model of ITBs that originate at the continental shelf break. ITBs are as important as the commonly-observed boluses, as they also transport near-surface larvae (Pineda 1999) and water (Lamb 2002, 2003) onshore. Although there have been numerous observations of ITBs in deep water ($h > 100$ m) (Helfrich and Melville 2006), ITBs are less commonly observed in the nearshore to the edge of the surf zone. Past nearshore observations of ITBs have occurred with minimal moorings and in intermediate h , such as in $h \sim 100$ m (Colosi et al. 2001; Holloway 1987), in $h \sim 40$ m (Pritchard and Weller 2005), and as shallow as $h = 11$ m (Pineda 1999). Lucas et al. (2011) observed the cross-shore transport of nitrates, energy, heat, and salinity owing

to ITBs in $h \sim 20$ m. However, unlike cold boluses (e.g., Sinnett et al. 2018), the fate and impact of ITBs have not been observed spatially to the edge of the surf zone. Colosi et al. (2018b) observed warm ITBs at Pt. Sal, California propagating onshore from the 50–20 m isobaths. Here, the connection is made from the 20–5 m isobath, the edge of the surf zone, with dense array(s) of thermistor strings (T-strings) that are often co-located with Acoustic Doppler Current Profilers (ADCPs). The first detailed observations of temperature and currents of shoaling ITBs outside of the surf zone are presented. The fate of ITBs is discussed as a source of nearshore mixing and net heating of the surf zone.

3.2 Methods

Vertical profiles of temperature and velocity profiles were obtained using cross-shore and alongshore arrays of T-strings and ADCPs deployed off Pt. Sal, California from 10 June to 22 July, 2015 (Figure 3.1). The field site is marked by three features: (1) Pt. Sal headland, (2) smaller headland, Mussel Pt, and (3) a submerged rocky outcrop that shoals to a water depth $h=16$ m, ~ 2 km from the coastline. The focus here is on observations for yeardays 185–195 during a period of high ITB activity. The cross-shore X-array spanned approximately 6 km from the 50–5 m isobaths and was comprised of 12 T-strings and 6 ADCPs. The alongshore Yi-array was comprised of 10 T-strings, each comprised of 12 thermistors spaced at 0.5 m over the vertical. Sample rates of the thermistors and ADCPs ranged from 0.25–2 Hz, but data were averaged over 2.5-min intervals. ADCPs at X02(7), X06(15), and X08(20) measured velocities over the vertical with 0.5–1.0-m bins. Note that in the notation, the number in the parentheses represents the mooring water depth. Velocities obtained one bin below surface wave troughs and above are removed due to significant wave height biases, which are calculated from co-located pressure measurements using linear wave theory (Dean and Dalrymple 1991). ADCPs at X10(30), X11(40), and X12(50) obtained velocities at 2-m bins. For these ADCPs, 4 bins (8 m) below the tidal surface are removed from the velocity profile owing to side-lobe contamination. The baroclinic cross-shore current is calculated as $u_{bc}(z, t) = u(z, t) - \overline{u(t)}$, where $u(z, t)$ is the measured current as function of time, t and elevation z , and $\overline{u(t)}$ is the depth-averaged current.

The bulk Richardson number, $Ri = \frac{N^2}{S^2}$, is calculated to determine the potential for mixing at Yi06. It is assumed the buoyancy frequency is only a function of temperature such that $N^2 = \alpha g \frac{\Delta T}{\Delta z}$ for this site (Colosi et al. 2018b), where α is the coefficient of thermal expansion

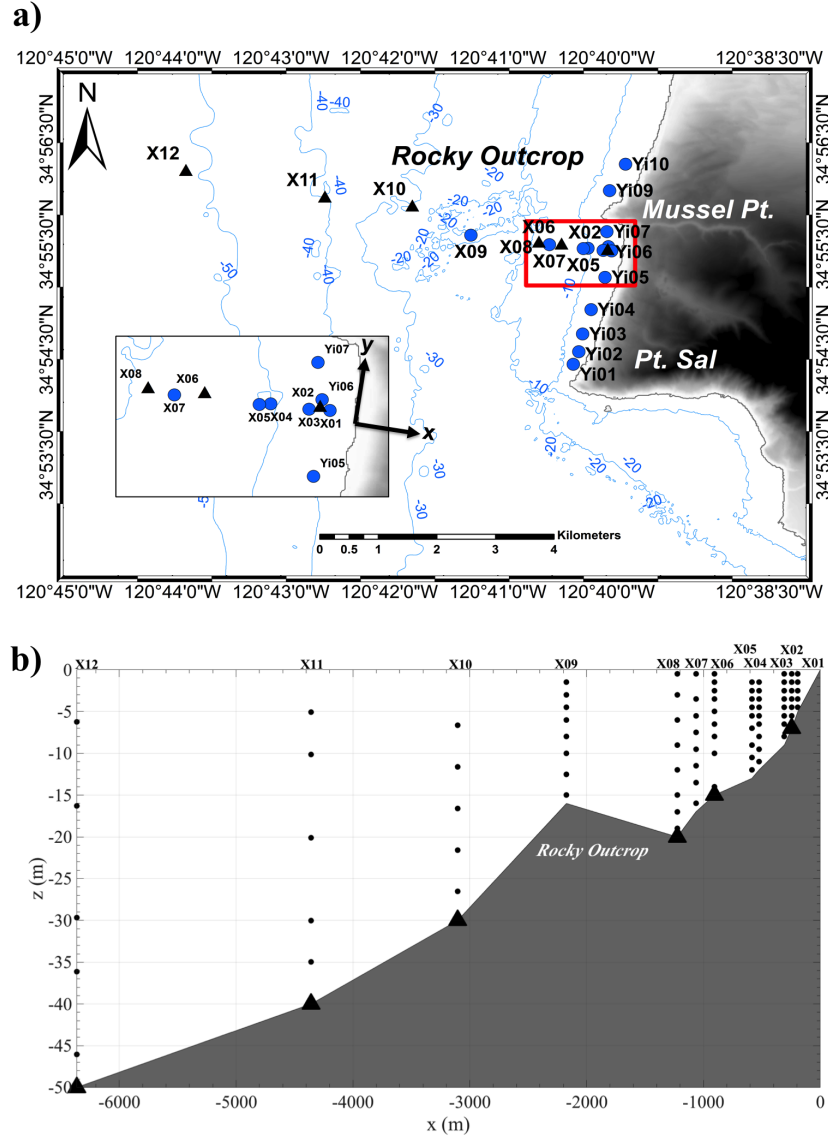


Figure 3.1. (a) Map of the experiment site. The blue dots mark the positions of T-strings, and black triangles mark the positions of co-located T-strings and ADCPs. The red box is the region for the map inset. (b) Cross-shore profile of the X-array. Black dots are thermistors and black triangles are ADCPs. In total, there were 86 thermistors in the X-array and 108 thermistors in the Yi-array

and g is the acceleration due to gravity. The vertical shear is $S^2 = \left(\frac{\Delta u}{\Delta z}\right)^2 + \left(\frac{\Delta v}{\Delta z}\right)^2$, where u and v are the cross- and alongshore velocities, respectively. To calculate Ri , values of N^2

are interpolated over the vertical where S^2 is evaluated. Because Yi06 is the only mooring with a co-located ADCP in the Yi-array, Ri cannot be calculated along the array. However, temperature overturns are calculated as a proxy for mixing. Overturns are calculated by differencing the temperature between an upper, T_z , and adjacent lower thermistor, T_{z+1} , separated by 0.5 m. If $T_z - T_{z+1} < -0.004^\circ\text{C}$, twice the instrument accuracy, it is considered an overturn. Overturns are calculated with 2.5-min averaged data to remove effects from surface gravity waves (sea/swell and infragravity).

3.3 Results

Temperature variability over the vertical at X06(15) (Figure 3.2a) reveals two distinct patterns at semidiurnal and faster timescales: 1) cold boluses (Figures 3.2a,b) and 2) warm ITBs (Figure 3.2a,d). Although the temperature and current patterns look similar for cold boluses and ITBs (Figure 3.2a), their dynamics are different. Thus, their distinction is important.

Cold bolus observations are characterized by two periods: (1) a period of near-bottom run-up of cold water and (2) a period of near-surface warm propagating onshore as a front associated with run-down (Pineda 1994; Walter et al. 2012). During the run-up, lower water column temperature decreases by $\Delta T \sim 0.75^\circ\text{C}$ in ~ 4 min (Figure 3.2b). This temperature change is consistent with a near-bottom onshore $u_{bc} \sim 10 \text{ cm s}^{-1}$ (Figure 3.2c). The run-up lasts for ~ 4 hours during which time high frequency (>1 cph), small amplitude internal waves occur throughout the water column (e.g., yearday 188.1 in Figure 3.2b). During the run-down, near-surface temperature increases with the arrival of a warm front (Figure 3.2b) owing to the compensating near-surface onshore flow (Figure 3.2c). The temperature approximately returns to the background state after the run-down period.

To characterize a cold bolus event, the cross-shore evolution is presented. The cold bolus exhibits a near-bottom surge of cold water that displaces warm surface water offshore (Figure 3.3a-c) over a 1.70-hour period. When the bottom water flows offshore as a gravity current it is more wave-like than current-like as the system returns to an equilibrium condition (Figure 3.3d) where the near-surface water returns as a warm-front (Figure 3.3e). After the cold bolus event, the system returns to the background state (Figure 3.3f). These observations are consistent with the observations of Pineda (1994, 1999); Walter et al. (2012). Note this

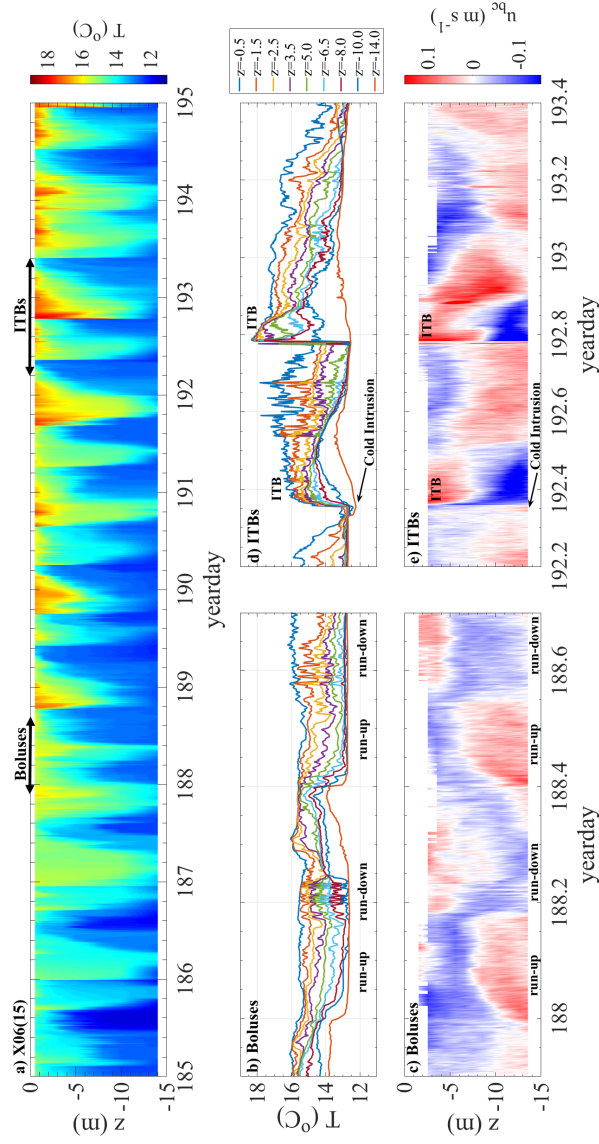


Figure 3.2. a) Temperature plotted over the vertical and time at X06(15). Plots of temperature against time for a periods of (b) boluses, and (d) ITBs. (c and e) u_{bc} plotted as a function of time over the vertical for similar time periods as (b) and (d). The bolus and ITB time periods shown in (b,c) and (d,e) are denoted in (a) by black arrows.

warm-front observation differs from ITBs discussed next.

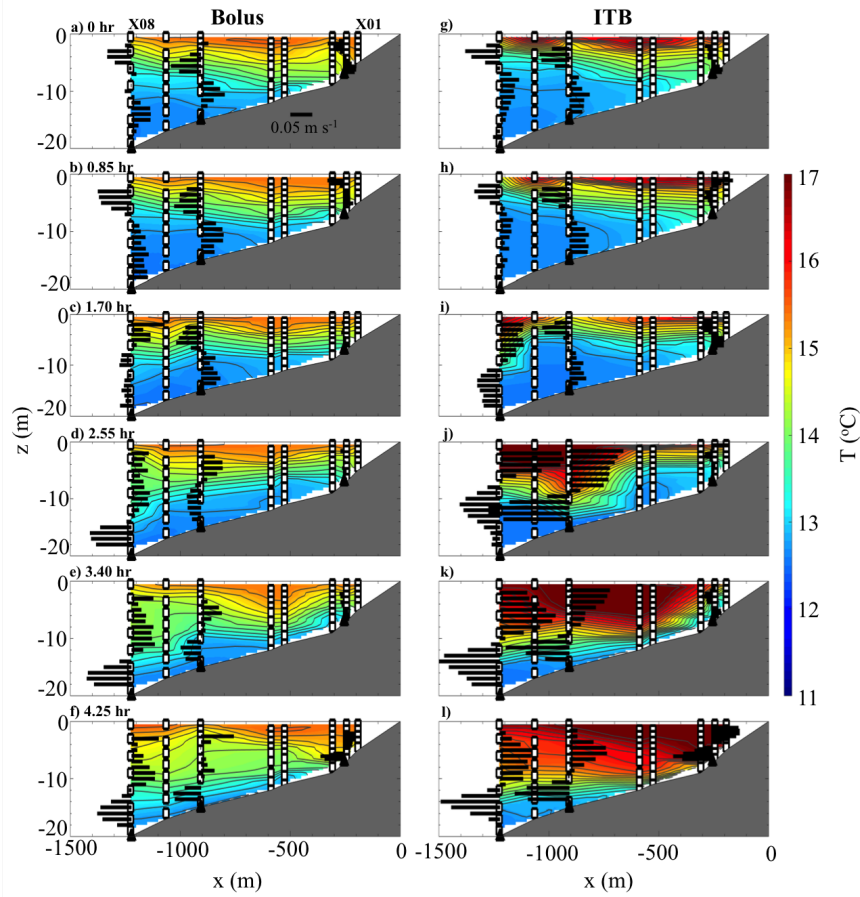


Figure 3.3. Cross-shore evolution of shoaling a (a-f) cold bolus and (g-l) ITB from X08(20) to X01(5). Each panel is a 0.85-hour time step. The shading and contours represent temperature, and each contour is 0.25°C . The black horizontal bars represent u_{bc} . White dots depict thermistor locations.

ITB arrivals are identified by sharp increases in T (Figure 3.2a,d). Between yeardays 185-195, 15 ITBs are identified. The ITBs arrive every 12.5 ± 3.14 hrs, ΔT associated with the fronts is $1.88 \pm 0.77^{\circ}\text{C}$ (Figure 3.2d), and propagation speeds vary from $8.6 \pm 1.3 \text{ cm s}^{-1}$ at X08(20) (Colosi et al. 2018b) to $6 \pm 2 \text{ cm s}^{-1}$ at X02(7). The warm-fronts of the ITBs are associated with a 2-layer u_{bc} with onshore near-surface currents as large as 13 cm s^{-1} and compensating, offshore bottom currents (Figure 3.2e). Empirical orthogonal analysis indicates that 66% of the variance is explained by a mode-1 internal wave, similar to findings by Lucas et al. (2011). A $\sim 5 \text{ cm s}^{-1}$ near-bottom onshore current precedes

the ITB arrival (Figure 3.2e), and is the source of a cold water intrusion (Figure 3.2d,e). The near-surface ITB propagates onshore, and forces the near-bottom cold water intrusion ahead of the bore upslope (Figure 3.3j,k) to the surf zone (Figure 3.3l). The ITB is similar to that of the cold bolus, but with notable exceptions. Preceding the ITB, the near-bottom cold water intrusion surges onshore (Figure 3.3g, h). The ITB arrival is evident through the sharp warm-front and near-surface convergent flow and near-bottom divergent flow accompanying the ITB front (Figure 3.3i). The warm ITB continues to propagate onshore and forces a cold intrusion onshore ahead of the ITB (Figure 3.3j). Unlike the bolus case, where the nearshore returned to the background state, the ITB results in a net heating of the nearshore (Figure 3.3l).

Two shoaling ITBs are presented as detailed examples (Figure 3.4). The sharp bore front and onshore propagation to the 5-m isobath is evident in the vertical structure of T (black arrow in Figure 3.4a-g). Surface heat flux likely contributes to the period of warm water for yeardays 192.5-192.6 at X02(7) and X01(5) (Figure 3.4f,g). The ITB current structure (Figure 3.4h-m) over the vertical is consistent with the temperature structure (Figure 3.4a-f). Downwelling-like circulation is present when the ITB arrives with surface, onshore u_{bc} as large as 30 cm s^{-1} at X20(50) and 2.0 cm s^{-1} at X02(7) (right of the black arrow in Figure 3.4h-m). Upwelling-like circulation is observed prior to the arrival of the ITB (left of the black arrow in Figure 3.4h-m) with offshore u_{bc} near the surface, and a near-bottom, onshore u_{bc} as large as 10 cm s^{-1} at X20(50) and 1.0 cm s^{-1} at X02(7). The cold water intrusion is consistent with this upwelling-like circulation (e.g., yeardays 192.6–192.75 in Figure 3.4e,l).

Associated with the ITB temperature front (vertical dashed lines in Figure 3.5a-e) is an increase in N^2 along the ITB front, and is maximum near the bottom. Concomitant with the nearly two-fold increase in N^2 is an increase in S^2 that is also maximum near the bottom (Figure 3.5c). Owing to the ADCP quality control procedure and the vertical gradients to calculate S^2 , Ri can only be obtained for the lower half of the water column (Figure 3.5c,d). The resulting Ri reveals dynamic instabilities ($Ri < 0.25$, red regions in Figure 3.5d) tend to occur during the cold water intrusions and after the passage of the bore front, however the lower water column is dynamically stable at timescales of 2.5-min vertical scales of 0.2m during the ITB. Although overturns are calculated over the water column, they tend to occur in the lower half of the water column when $Ri < 0.25$ (yellow dots in Figure 3.5d).

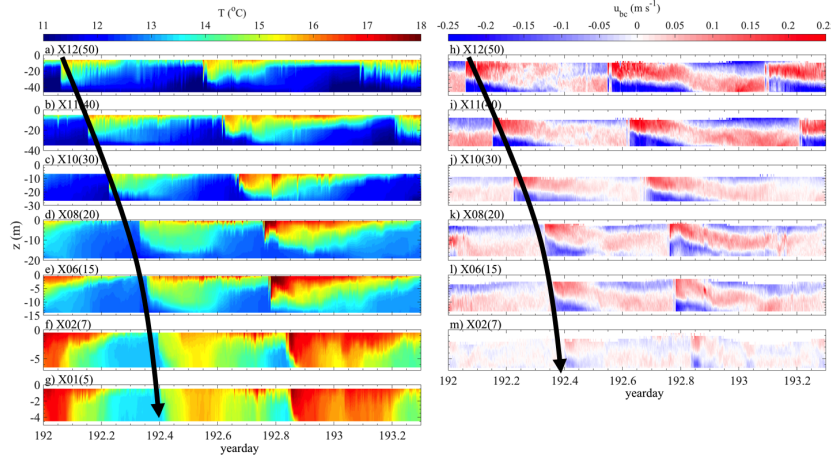


Figure 3.4. Temperature plotted as a function of time and elevation for two example bores propagating from (a) X12(50) to (g) X01(5) from year day 192 to 193.3. (h-m) Baroclinic, cross-shore currents, u_{bc} , for the same bores in (a-f). The black arrows illustrate the onshore propagation of the ITB.

This suggest the overturns are a good proxy for mixing at the other moorings in the Yi-array where there are no co-located ADCPs.

3.4 Discussion

Colosi et al. (2018b) found the ITBs off Pt. Sal propagated normal to the isobaths between 50-20 m with a mean group speed of $c = 8.6 \pm 1.3 \text{ cm s}^{-1}$ and a mean duration $t_d = 10 \pm 3.2$ hrs at the 20-m isobath. The ITB energy flux decayed exponentially with an e-folding scale of $2.4 \pm 0.1 \text{ km}$. Extending this analysis shoreward, the distance between X01(5) and X12(50) is ~ 2.57 e-foldings. This distance suggests that the ITB energy flux is reduced by $\sim 92\%$ at the edge of the surf zone relative to the 50m isobath. The internal Iribarren number is $\xi = s / \sqrt{(\zeta_o L)}$, where s is the topographic slope, ζ_o is the mean vertical isopycnal displacement, and L is the ITB wavelength (Boegman et al. 2005). Using $\zeta_o = 7.2 \pm 2.4 \text{ m}$ and $L = 1.8 \pm 0.6 \text{ km}$ (Colosi et al. 2018b), and $s = 0.016$ at X08(20), $\xi = 0.25 \pm 0.05$. The small value of ξ at this site suggests one mechanism for ITB dissipation is through a soliton fission process rather than undergoing a breaking process (Aghsaee et al. 2010). Colosi et al. (2018b) estimated the energy dissipation between the 50-20-m isobaths, and suggests the dissipated ITB energy could be going to turbulence.

Walter et al. (2012) found that the dissipation in the interior of the water column owing to the warm front of a bolus relaxation was of the same magnitude as the turbulent dissipation generated near the sea bed, ϵ . They estimated the near-bottom dissipation as

$$\epsilon \sim \frac{(C_d)^{3/2} u_b^3}{\kappa z_m} \quad (3.1)$$

assuming a law of wall scaling for the dissipation, where C_d is the drag coefficient, $\kappa=0.41$ is the Von Karman constant, and $u_b = \sqrt{u^2 + v^2}$ is the near-bottom speed at the first ADCP bin height z_m above the sea bed. For the ITBs, ϵ at $z_m = 1.08$ m increases by an order of magnitude with the ITB front (Figure 3.5e) and is an order of magnitude larger than the values reported by Walter et al. (2012) for the bolus relaxation. If the same relationship holds as found by Walter et al. (2012) such that the dissipation owing to ITBs, $\epsilon_{ITB} \sim \epsilon$, then ITBs are an important mechanism for vertical mixing at the edge of the surf zone. The likelihood for turbulence or mixing owing to ITBs is highlighted by Ri at Yi06 (Figure 3.5d). Overturns are consistent with times of the cold water intrusions when $Ri < 0.25$, and are not observed during the ITB front. This is likely because the spatial (0.5 m) and temporal (2.5 min) scales of the overturns do not represent the scales for ϵ_{ITB} . The number of overturns tending to zero is also indicative of the relatively large ϵ_{ITB} that acts to vertically mix the water column. Because the number of overturns is inversely related to ϵ_{ITB} , the distribution in the number of overturns in the Yi-array can be used to infer alongshore heterogeneities in the ϵ_{ITB} .

Less overturns occur at Yi01-Yi04 (Figure 3.5f). This implies relatively higher ϵ_{ITB} and more vertical mixing owing to ITBs. The number of overturns peaks at Yi05, indicating a relative minimum in ϵ_{ITB} , and less vertical mixing owing to ITBs. The number of overturns significantly decreases north of Yi05 (Figure 3.5f). Colosi et al. (2018b) found ITBs propagated toward $12 \pm 10^\circ$ north of east. The direction implies that Yi05-Yi07 are typically in the lee of the outcrop (grey oval in Figure 3.5g), where large ITB dissipation occurs before reaching the surf zone. Yi01-Yi04 are rarely in the lee of the outcrop (Figure 3.5g), and ITBs are not affected by the outcrop. The interaction between the ITBs and outcrop can be ecologically significant owing to the ITBs as a source of warm, potentially phytoplankton rich, surface water for the rocky outcrop (Witman et al. 1993).

Although the NLIW generation mechanisms and locations cannot be determined from this

dataset, it is speculated that the period of mixed boluses and ITBs suggests two possible generation locations with different background shear or stratification. The cold boluses can be generated from NLIWs breaking on topography (Lamb 2014) or by undergoing a polarity reversal (Shroyer et al. 2009) as a wave of depression passes through a turning point (Helfrich and Melville 2006). The ITBs are likely generated by barotropic currents interacting with topography, and propagate onto the shelf (Lamb 1994, 2002, 2003) without passing through a turning point (Kao et al. 1985). A modeling analysis of the possible generation locations would help to unwrap the complex NLIW field at Pt. Sal.

3.5 Conclusion

Observations using a dense cross-shore array measured, for the first time, the evolution of ITBs to the edge of the surf zone which stand in contrast to the more commonly observed bottom cold-water boluses (e.g., Nam and Send 2011; Sinnett et al. 2018; Smith et al. 2016; Walter et al. 2012; Woodson 2018). Comparisons of cold boluses with internal tide bores reveal that the cold bolus is similar to internal swash as described in Sinnett et al. (2018), where the run-up and run-down create a warm front that propagates onshore (Pineda 1994; Walter et al. 2012). Internal tide bores, however, transport near-surface warm water onshore that increases the nearshore temperature by 1.2°C at this site. The warm front associated with the internal tide bores is sharper than the front associated with the bolus run-down. While both the bolus and internal tide bores are mode-1 internal waves, the structure of u_{bc} over the vertical delineates the internal tide bore from the bolus with abrupt temporal changes occurring in u_{bc} for the internal tide bores. Alongshore heterogeneity in ϵ_{ITB} is inferred using temperature overturns, which indicate vertical mixing. The interaction between internal tide bores and a rocky outcrop results in less mixing in the lee of the topography. Internal tide bores are an important onshore transport mechanism that can transport not only surface material to the coast, but also near-bottom material during a cold water intrusion, which could impact nearshore ecology.

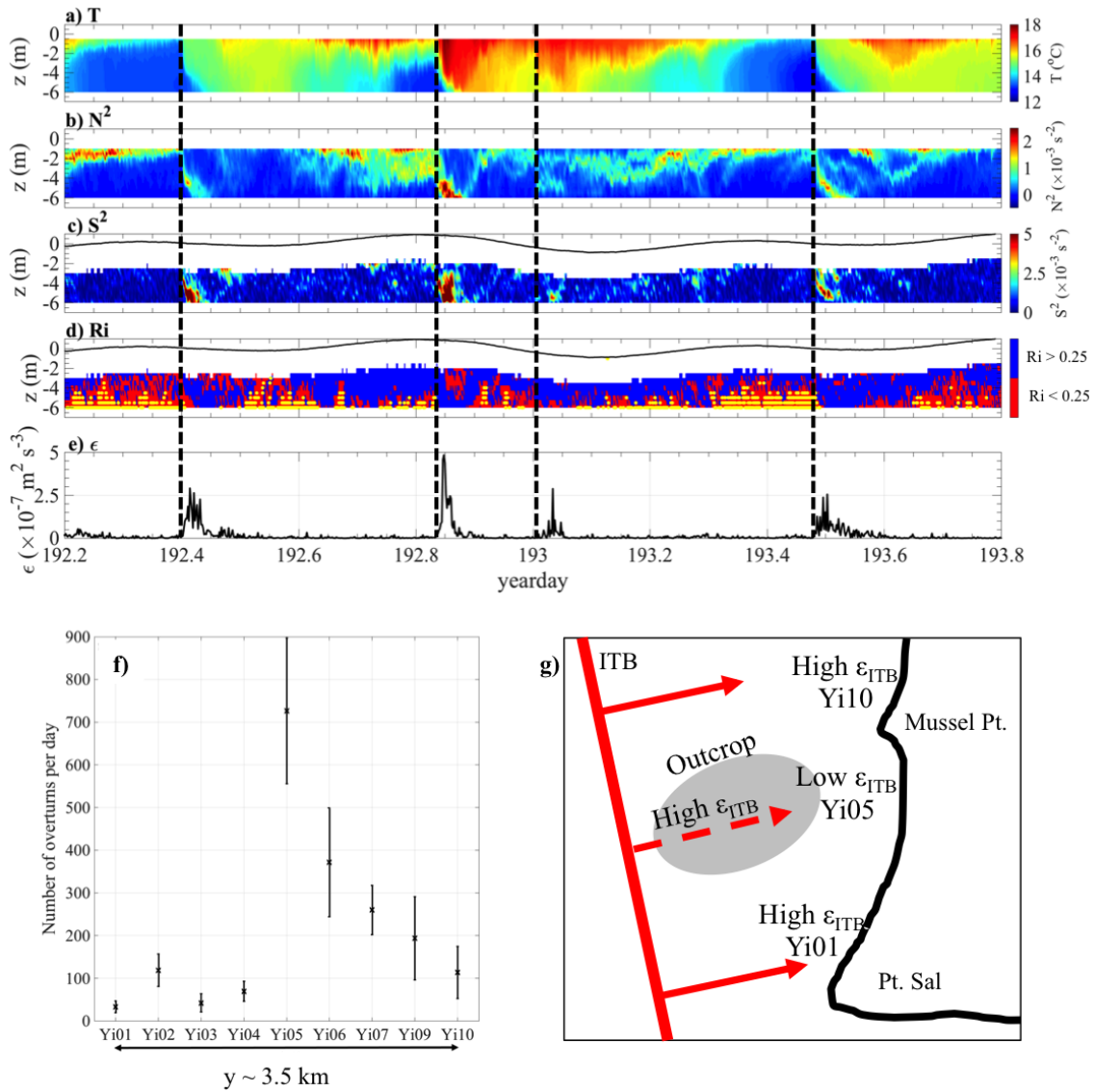


Figure 3.5. Plots of (a) temperature, (b) N^2 , (c) S^2 , (d) Ri , and ϵ at mooring Yi06. The solid black line in (c) and (d) indicate the tidal surface. In (d), the red areas indicate times when $Ri < 0.25$, blue areas indicate when $Ri > 0.25$, the yellow dots on (d) denote temperature overturns. The vertical lines denote ITB fronts. (f) Number of overturns per day averaged over 10 days at each mooring in the Yi-array. The black lines indicate the 95% confidence interval. (g) Cartoon depicting a shoaling ITB (red line, arrows indicate direction of propagation). The gray oval represents the rocky outcrop that affects ITB dissipation, ϵ_{ITB} . The locations of Yi01, Yi05, and Yi10 are shown for reference.

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CHAPTER 4:

Statistical Analysis of Alongshore Temperature Variability on the Central California Coast

This chapter has been submitted to the Journal of Physical Oceanography with minor formatting changes. As the main author of this work, I made the major contributions to the research and writing. Co-authors include Jamie H. MacMahan, Matt K. Gough, and Falk Feddersen.

4.1 Introduction

Nearshore water temperature variability has important implications in the distribution, retention, and settlement of marine biota (O'Connor et al. 2007) on various temporal (Boehm et al. 2004) and spatial scales (Tapia et al. 2014). Satellite derived measures of ocean temperature are valuable because of the large regions [O(1000 km)] sampled in a single observation (e.g., Casey and Adamec 2002). There are practical limitations, however. Satellite derived temperatures are representative of sea surface temperature (SST), and do not provide subsurface temperatures. Studies of SST variability are typically limited to seasonal timescales because of revisit times that can range between O(1-10 days) coupled with periods of coastal clouds and fog that can limit temporal resolution (Thomas et al. 2002). Although the spatial resolution [O(0.1-10 km)] is adequate to resolve nearshore features, the observations can be contaminated by land temperatures (Blanchette et al. 2008). Owing to these limitations, nearshore temperature variability is typically studied with in situ measurements. In situ measurements allow for rapid [O(1s)] and continuous [O(months)] observations of nearshore water temperature over the vertical. These sample rates and durations allow for studies of nearshore water temperature variability at subtidal (ST, $f < 33^{-1}$ cph), diurnal (DU, $33^{-1} < f < 16^{-1}$ cph), and semidiurnal (SD, $16^{-1} < f < 10^{-1}$ cph) timescales. Various spatial scales can be studied owing to relative spacing between moorings. In general, studies of nearshore temperature variability can be divided into two categories: (1) heat budgets and (2) statistical analyses of temperature variability. The aim of a heat budget is to account for all of the heat flux into a control

volume through cross-shore, alongshore, and surface heat fluxes. Inner shelf heat budgets at ST- (Austin 1999; Freismuth et al. submitted; Melanie R. Fewings and Lentz 2011), DU- (Freismuth et al. submitted; Herdman et al. 2015; Suanda et al. 2011; Walter et al. 2017), and SD-timescales (Freismuth et al. submitted) have identified the dominant mechanisms (e.g., upwelling, internal waves, and sea breeze driven transport) contributing to heating and cooling. 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 alongshore flow divergences (Dever and Lentz 1994) resulting from coastal heterogeneities, such as coastal topography or kelp forests (Suanda et al. 2011). These last two conditions can make heat budget studies impractical over long stretches of heterogeneous coastlines.

Statistical analyses of in situ temperature fluctuations have been performed at ST-timescales and longer along an 800 km stretch of coastline (Tapia et al. 2014), at DU- and SD-timescales in a small [$O(10\text{ km})$] bay (Bonicelli et al. 2014a), and for a coastal bight [$O(50\text{ km})$] at both DU- and SD-timescales (Aristizabal et al. 2016, 2017). Tapia et al. (2014) found that along the central Chilean coast regional temperature variability was related to the alongshore wind stress, and headlands modified the local variability. Furthermore, the distribution of benthic communities and algal biomass along the coast was correlated with the spatial patterns in the temperature variability. Bonicelli et al. (2014b) found that differences in the amplitude of diurnal wind forcing in a small bay led to different barnacle settlement rates across the bay. Aristizabal et al. (2016) performed a detailed study of DU- and SD-band temperature variability in the Santa Barbara Channel (SBC, Figure 4.1b). They found that sea breeze driven DU-band temperature variability was coherent within the SBC, however it was not coherent with the variability north of Pt. Conception. SD-band variability was only coherent around the Channel Islands.

Here, the focus is on temperature variability along 55-km of the central California inner shelf between Pt. San Luis and Pt. Arguello (Figure 4.1a) in the Santa Maria Basin (SMB, Figure 4.1b) at ST, DU-, and SD-timescales. The coastline here is characterized by a headland-and-bay topography with three headlands: Pt. Sal, Purisima Pt., and Pt. Arguello. North of the headlands, the coastline is generally oriented $\sim 15^\circ$ east of North for approximately 10 - 20 km, and are typically sandy (red bars in Figure 4.1). South of the headlands, the coastline is oriented typically $\sim 30^\circ$ west of north for approximately 6 - 8 km, and are typically rocky (blue bars in Figure 4.1). Processes generating temperature

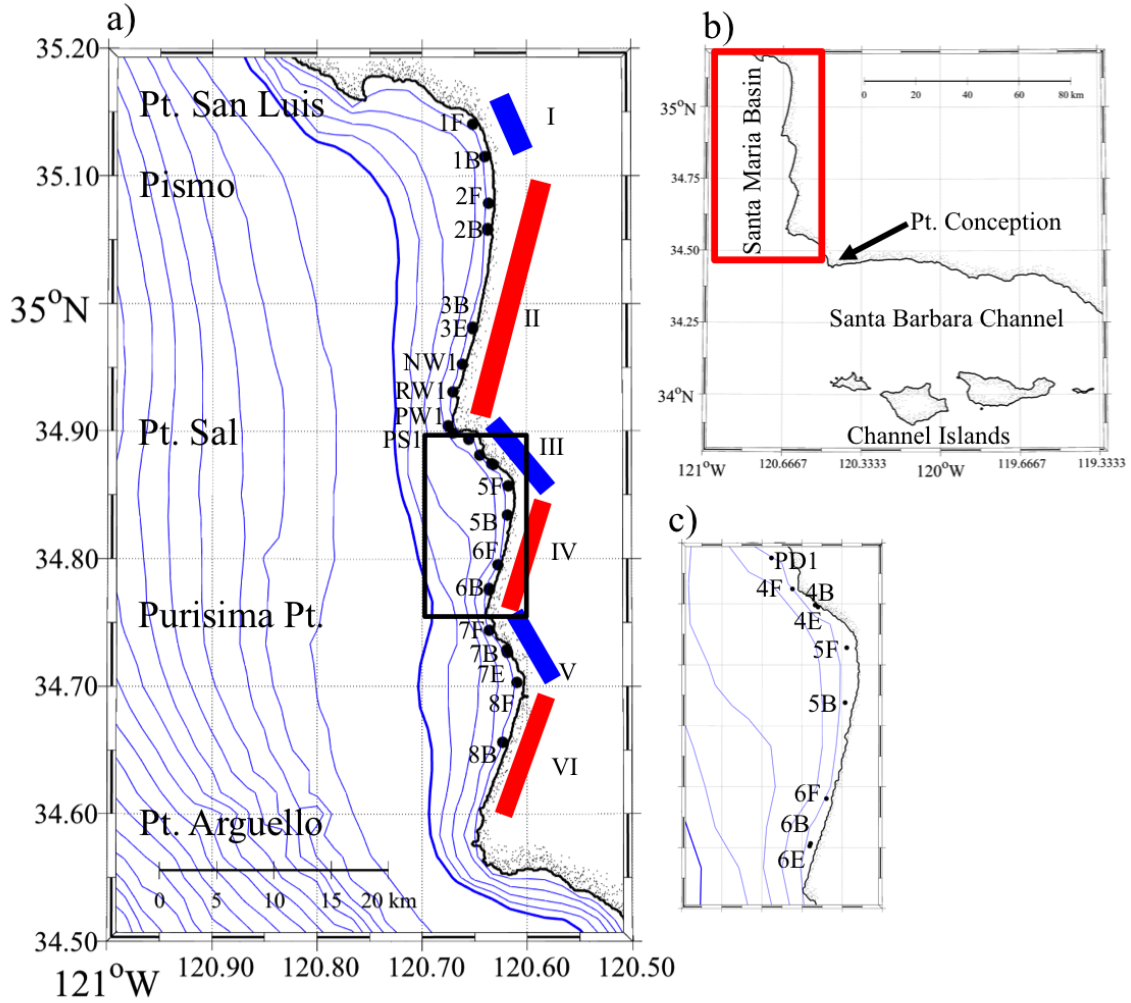


Figure 4.1. a) Map of the experiment site. Mooring positions are indicated by black dots. Red and blue bars identify the coastlines I-VI. Depth contours (blue lines) are shown for 10 to 50 m in 10 m increments, and 100 m increments in depths greater than 50 m. The 50 m contour is denoted by the thick, blue contour. The black box denotes the area of (c) inset. Contours in inset (c) are the same as in (a). b) Overview of region, red box indicates the region in (a).

variability in the region (addressed further in the Discussion section) including the SBC (Figure 4.1b) are coastal upwelling and relaxations at ST-timescales (Melton et al. 2009; Washburn et al. 2011); sea breeze driven transport (Aristizabal et al. 2016), sea breeze driven diurnal period internal waves (Cudaback et al. 2005) at DU-timescales; and the

internal tide at SD-timescales (Aristizabal et al. 2016). Aristizabal et al. (2017) observed DU- and SD- temperature fluctuations driven by the vertical advection of a background stratification by the barotropic tide.

The goal of this current study is not to reprove the processes driving temperature variability on the central California inner shelf, but rather to statistically describe the spatial patterns and representativeness of temperature variability. Temperature fluctuations are explained over the vertical and over the horizontal for depth-averaged temperature, T_{da} at ST-, DU-, and SD-timescales. The spatially coherent patterns at each timescale are described with a principal component analysis over the vertical and over the horizontal for T_{da} . The temporal evolution of variability at DU- and SD-timescales is examined with scale-averaged wavelet power. Decorrelation length scales are estimated for T_{da} at DU- and SD-timescales.

4.2 Methods

Measures of temperature T were obtained over the vertical at 25 temperature string (T-string) moorings deployed in an alongshore array on the inner shelf from Pismo to Pt. Arguello, California (Figure 4.1a) between September 1 and October 14, 2017 (yeardays 244-287). The distance between moorings ranged from 150 m to 54 km. A total of 20 T-strings were deployed along the 9-m isobath. For these T-strings, measures were obtained from 1.5 m above the sea bed to mean low-water with a vertical resolution of 0.5 m using RBR SoloT thermistors with a 1-s sample rate. An additional thermistor was attached to 2-m, rigid line to measure temperature at the free surface through all tidal cycles. Five T-strings located near Pt. Sal (NW1, RW1, PW1, PS1, and PD1, Figure 4.1a,c), were deployed in 10-16 m water depths, obtained vertical profiles from 1.5 m above the sea bed to the free surface with a vertical resolution of 1.5 m, and sampled at 30 s. These five moorings consisted of RBR SoloT thermistors. All temperature data are hourly-averaged.

Variance (σ^2) is estimated from the hourly-averaged T [$\sigma^2(T)$] at each thermistor as well for hourly- and depth-averaged temperature, T_{da} , [$\sigma^2(T_{da})$] at each T-string over the duration of the experiment. Estimates of depth-averaged subtidal [frequencies $f < 33^{-1}$ cph, $\sigma_{ST}^2(T_{da})$], diurnal [$16^{-1} < f < 33^{-1}$, $\sigma_{DU}^2(T_{da})$], and semidiurnal [$10^{-1} < f < 16^{-1}$, $\sigma_{SD}^2(T_{da})$] are obtained at each T-string, and averaged over the array (array averages are denoted by $\langle \cdot \rangle$). The temporal evolution of $\sigma_{BP}^2(T_{da})$, where BP represents the band-passed DU- or

SD-signal, is examined with scale-averaged wavelet power, such that

$$W_{BP}(t) = \frac{\delta j \delta t}{C_\delta} \sum_{j=j_1}^{j_2} \frac{|W(t, s_j)|^2}{s_j}, \quad (4.1)$$

where $|W(t, s_j)|^2$ is the wavelet power spectrum, δj is the scale resolution, δt is the sampling time interval, $C_\delta = 0.776$ is a reconstruction constant for the Morlet wavelet, and s_j is the wavelet scale (Torrence and Compo 1998).

Spatially coherent patterns of T are described with a principal component analysis (PCA), where T and T_{da} are decomposed into spatial eigenfunctions $[F_n(x_i)]$ referred to as EOFs and time-varying amplitudes $[\alpha_n(t)]$ referred to as principal components (PCs) (Preisendorfer 1988). The variability is separated into orthogonal modes, such that

$$\hat{T}(x_i, t) = \sum_{n=0}^N \alpha_n(t) F_n(x_i) \quad (4.2)$$

where $\hat{T}(x_i, t)$ is the temperature anomaly obtained by removing the temporal mean from T or T_{da} , x_i represents either position in the horizontal (x, y) or vertical (z), t is time, and N is the total number of measurement positions either in the horizontal or over the vertical and the total number of modes. Four principal component analyses are performed: (1) for every thermistor in the array; (2) over the vertical for each mooring; (3) over the horizontal for depth-averaged temperature, $\hat{T}_{da}$ at each T-string; and (4) over the horizontal for $\hat{T}_{da}^{ST}$, $\hat{T}_{da}^{DU}$, and $\hat{T}_{da}^{SD}$. For clarity, vertical modes and horizontal modes are denoted such that $EOF0(z)$ is a mode 0 over the vertical for the total signal and $EOF0^{ST}(x)$ is a mode 0 over the horizontal of $\hat{T}_{da}^{ST}$, for example.

To account for phase differences in T_{da} that result from coastal topography and differences in cross-shore propagation along the array, lagged cross-correlation functions, r are estimated for T_{da}^{ST} , T_{da}^{DU} , and T_{da}^{SD} over all combinations of moorings for 48-hr windows, such that

$$r(\tau) = \frac{\mathbb{E} [T_{da}^{BP,i}(t) T_{da}^{BP,j}(t + \tau)]}{\sigma_i \sigma_j} \quad (4.3)$$

where $\mathbb{E}$ is the expectation operator, i and j denote unique pair of moorings, t is time, τ is

lag, and σ is the standard deviation of T_{da} . The maximum spatial lag coefficients r_{max}^{DU} and r_{max}^{SD} are estimated in for the DU and SD bands, respectively.

Measures of hourly winds were obtained at the National Data Buoy Center (NDBC) station 46011 located approximately 30 km offshore of Pt. Sal. Wind stress (τ_s) was calculated according the Large and Pond (1981). Upwelling wind relaxations are identified from the low-pass ($f < 33^{-1}$ cph) filtered τ_s . Wind velocities at a height of 10 m at moorings 1B, 2B, 3E, 4B, 5B, 6B, 7B, and 8B are estimated from a 2-km horizontal resolution Coupled Ocean/Atmosphere Mesoscale Prediction System (COAMPS) forecast model (Hodur 1997). Model wind velocities are rotated into a local cross-shore direction, and τ_s^x . To explore the spatial and temporal variation of the COAMPS, diurnal τ_s^x , the diurnal scale-averaged wavelet power (Torrence and Compo 1998) of τ_s^x is estimated at the eight moorings described above.

4.3 Results

Low-pass ($f < 33^{-1}$ cph) filtered winds at NDBC 46011 are predominantly upwelling favorable that persist between 3-5 days, and undergo four relaxations (grey boxes Figure 4.2a) including reversals on yeardays 247 and 255. A period of persistent diurnal winds occurs between yeardays 265-270. The observed wind patterns for 2017 are consistent with previous wind observations (Fewings et al. 2015; Melton et al. 2009; Ohashi and Wang 2004). Diurnal scale-averaged wavelet power of the COAMPS cross-shore winds stress, $W_{DU}(\tau_s^x)$, tends to be greater north of Pt. Sal, and exhibits intermittent 2- to 5-day bursts of DU-band energy (Figure 4.2b).

Patterns in $T(z, t)$ (Figure 4.3) for ST-, DU-, and SD-bands for mooring 6B (Figure 4.1a,c) are presented as an example of the alongshore array. The ST pattern is characterized by two distinct periods (Figure 4.3b). The period between yeardays 246-263 exhibits warmer water over the vertical with near-surface $T^{ST}=16-18.5^\circ\text{C}$. Upwelling relaxations (black boxes in Figure 4.3) precede the times of warm water that exhibit $\Delta T^{ST} \sim 4^\circ$. The second period between yeardays 265-287 exhibits cooler water with T^{ST} ranging from $13.5-15.5^\circ\text{C}$. There are intermittent times of warm water occurring at ~ 5 -day intervals beginning on yearday 272. Fluctuations in T^{DU} occur throughout the water column, and are as large as 2.0°C (Figure 4.3c). The time of largest fluctuations are consistent with times of increased τ_s^C .

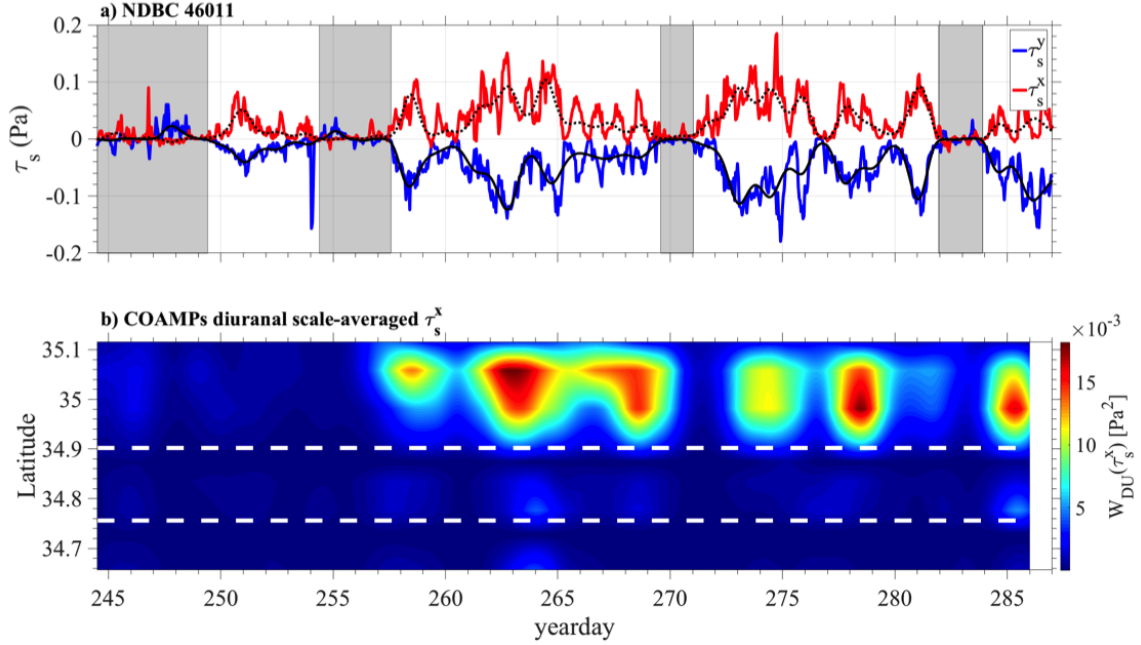


Figure 4.2. (a) Wind stress, τ_s , at NDBC station 46011. Low-frequency ($f < 33^{-1}$ cph) winds are indicated by a dotted and a solid black line for E-W and N-S τ_s , respectively. Grey boxes indicates times of upwelling relaxation. (b) Diurnal scale-averaged wavelet power for cross-shore wind stress along the experiment site from COAMPS. White dashed lines indicate the locations of Pt. Sal and Purisima Pt.

Fluctuations in T^{SD} are as large as 2.4°C , and occur over the vertical. Whereas the DU-band intermittency occurs throughout the experiment, the times of largest T^{SD} fluctuations occur between year days 255-260 following an upwelling relaxation event (Figure 4.3d).

The temporal and spatial evolution of T_{da} (Figure 4.4) reveals similar trends as T (Figure 4.3). The first warming event (year day 247) propagates poleward at a rate of 24.5 km day^{-1} assuming straight-line propagation (black arrows in Figure 4.4b). The warmest water, $T = 19.2^\circ\text{C}$, is at the southern extent of the array, and decreases to $T = 17.1^\circ\text{C}$ at the northern extent. There is a period (year days 255-258) of warm water north of Pt. Sal (top black, dashed line in Figure 4.4b) that does not appear to be a result of a propagation of warm along the 9-m isobath; however, there is evidence for poleward propagation at 22.8 km day^{-1} from Pt. Sal to the north. The warm events at year day 271 and 283 propagate poleward at 15.7 km day^{-1} and 14.8 km day^{-1} , respectively. These events do not appear to propagate

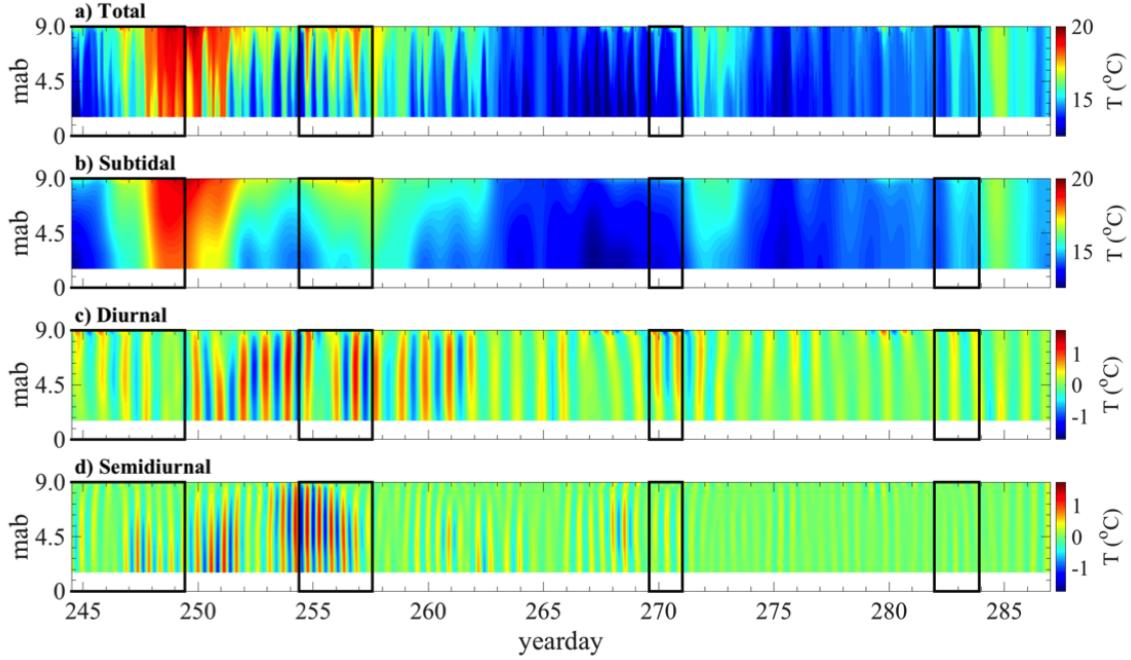


Figure 4.3. Temperature plotted over the vertical at mooring 3E for (a) the total, (b) subtidal, (c) diurnal, and (d) semidiurnal frequencies. Black boxes indicate times of upwelling relaxations.

poleward of Pt. Sal, unlike the first two warming events. These poleward-propagating warm events are consistent with the observations of Melton et al. (2009); Washburn et al. (2011) and modeling results of Suanda et al. (2016). Fluctuations in T_{da}^{DU} tend to be coherent along the array, however the largest fluctuation occur north of Pt. Sal (Figure 4.4c). Fluctuations T_{da}^{SD} are less spatially coherent along the array compared with the ST- and DU- bands (Figure 4.4d). The largest fluctuations occur between yeardays 250-258 (Figure 4.4d) along coasts II and IV (Figure 4.1) The SD-band variability decreases during times of upwelling favorable winds (Figure 4.2a), consistent with the findings of Pineda and Lopez (2002); Walter et al. (2014).

Fluctuations in T_{da}^{DU} and T_{da}^{SD} power are examined with scale-averaged wavelet power ($W_{DU}(T_{da})$ and $W_{SD}(T_{da})$, respectively). The times of largest $W_{DU}(T_{da}) \sim 4^{\circ}\text{C}^2$ and $W_{SD}(T_{da}) \sim 7^{\circ}\text{C}^2$ occur after wind relaxations and are north of Pt. Sal (Figure 4.5a,b). Times of increased $W_{DU}(T_{da})$ (Figure 4.5a) are consistent with times of increased $W_{DU}(\tau_s^x)$ (Figure 4.2b). Although $W_{DU}(\tau_s^x)$ is relatively weak prior to yearday 257, there are periods

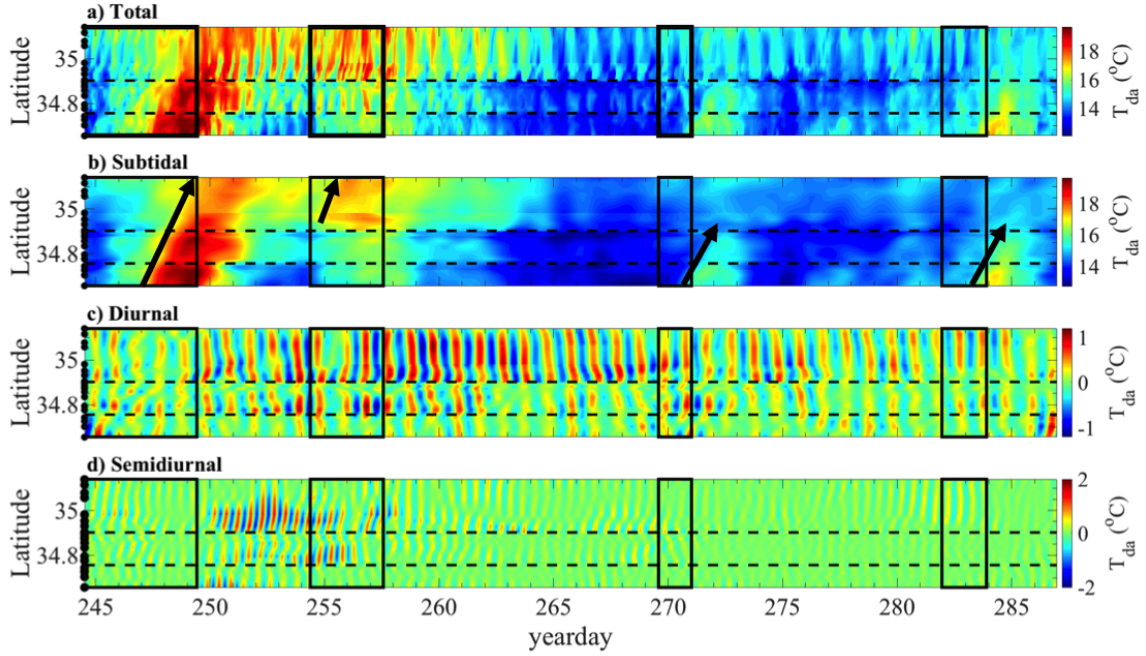


Figure 4.4. Hovmoeller diagrams of depth-averaged temperature plotted for (a) the total, (b) subtidal, (c) diurnal, and (d) semidiurnal frequencies. Black boxes indicate times of upwelling relaxations. Horizontal dashed black lines indicate the locations of Pt. Sal and Purisima Pt.

of increased $W_{DU}(T_{da})$ north of Pt. Sal and Purisima Pt. (Figure 4.5a). The largest $W_{SD}(T_{da})$ is observed north of Pt. Sal between yeardays 250-254 (Figure 4.5b). Between yeardays 250-257, when $W_{SD}(T_{da})$ is relatively stronger, $W_{DU}(T_{da})$ is conversely weaker. After yearday 257, $W_{SD}(T_{da})$ is relatively weak along the array. Both $W_{DU}(T_{da})$ and $W_{SD}(T_{da})$ are weak south of Purisima Pt. throughout the experiment (Figure 4.5a,b). The intermittency and times of largest $W_{DU}(T_{da})$ and $W_{SD}(T_{da})$ suggest a connection between the ST-band and the DU- and SD-bands.

Temperature fluctuations are further characterized by estimating σ^2 along the array. The variance averaged over all thermistors is $\langle \sigma^2(T) \rangle = 1.98 \pm 0.31^\circ\text{C}^2$. Array averages are denoted by $\langle \cdot \rangle$ and reported with the corresponding standard deviation. The first EOF mode (*EOF0*) accounts for 76% of $\sigma^2(T)$, and the second mode (*EOF1*) accounts 11%. PCA over the vertical at each mooring indicates that $\sim 93\%$ of $\sigma^2(T)$ is explained by *EOF0*(z), and *EOF1*(z) explains $\sim 5\%$ of the variance. In the ST-, DU-, and SD-bands, *EOF0*(z) accounts for approximately 96%, 78%, and 83% of the variance, respectively. The *EOF0*(z)

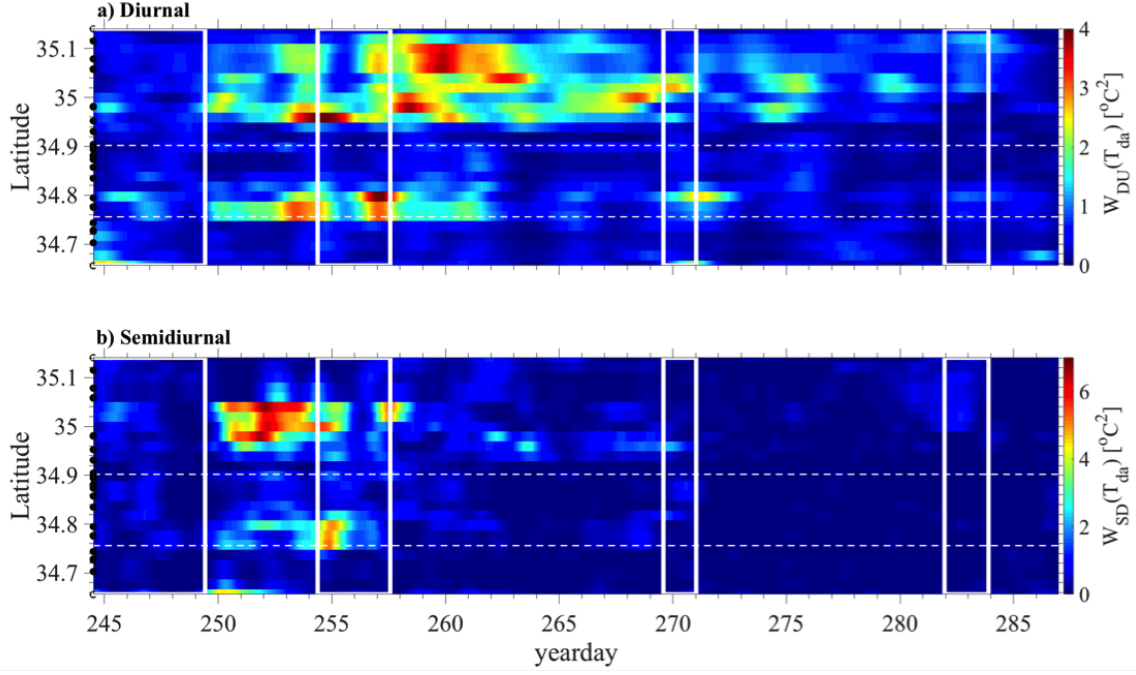


Figure 4.5. Hovmoeller diagrams of a) Diurnal and b) semidiurnal scale-averaged power. White boxes indicate times of upwelling relaxation, and horizontal white dashed lines indicate the locations of Pt. Sal and Purisima Pt.

structure is similar to a depth-uniform temperature structure, therefore T_{da} is used to discuss this mode throughout this manuscript owing to the high percentage of variance explained in each band.

Over the array, $\langle \sigma^2(T_{da}) \rangle = 1.86 \pm 0.28^\circ\text{C}^2$, $\langle \sigma_{ST}^2(T_{da}) \rangle = 1.65 \pm 0.35^\circ\text{C}^2$, $\langle \sigma_{DU}^2(T_{da}) \rangle = 0.11 \pm 0.06^\circ\text{C}^2$, and $\langle \sigma_{SD}^2(T_{da}) \rangle = 0.07 \pm 0.05^\circ\text{C}^2$. The ST-band accounts for $\sim 83\%$ of the total observed variance, whereas the DU-band accounts for $\sim 6\%$ and the SD-band accounts for $\sim 4\%$.

The spatial patterns in the temperature variability over the horizontal in the ST-, DU- and SD-bands are described with PCA. The two leading modes in the ST-band explain 87% and 10% of $\sigma_{ST}^2(T_{da})$ (Figure 4.6a). The first mode [$EOF0^{ST}(x)$] is described by a monopole structure where the variability is in phase along the array. The associated PC (Figure 4.6b) exhibits fluctuations on timescales of 3-5 days, and is positive between yeardays 246-261 and negative between yeardays 261-283. The times of positive expansion

coefficients correspond to the times of upwelling relaxations, and the time of negative expansion coefficients correspond to the times of upwelling. In the DU-band, $EOF0^{DU}(x)$ explains 67% of the variance and the $EOF1^{DU}(x)$ explains 12% of $\sigma_{DU}^2(T_{da})$ (Figure 4.6c). The variability is in phase along the array, however $EOF0^{DU}(x)$ amplitudes are reduced along coastlines I, III, V, and VI (Figure 4.1a). The corresponding PCs indicate that the largest $EOF0^{DU}(x)$ amplitudes occur between yeardays 257-262 (Figure 4.6d), consistent with $W_{DU}(T_{da})$. In the SD-band, the two leading modes combined explain 52% of the variance (Figure 4.6e), indicative of the SD-band being less coherent along the array. The leading mode ($EOF0^{SD}(x)$) accounts for 32%, exhibits a similar structure as $EOF0^{DU}(x)$ with reduced amplitudes along coastlines I, III, V, and VI (Figure 4.1a). The second EOF mode ($EOF1^{SD}(x)$) explains 20% of $\sigma_{ST}^2(T_{da})$, and is described by a dipolar structure where the variability north of Pt. Sal is out of phase with the variability between Pt. Sal and Purisima Pt. The amplitudes are essentially zero south of Purisima Pt. The PCs exhibit the observed intermittency with the largest amplitudes following the wind relaxation near yearday 250 (Figure 4.6f).

Cross-correlation coefficients, r are estimated for T_{da}^{ST} , T_{da}^{DU} , and T_{da}^{SD} between each pair of moorings in the array. As an example, cross-correlations between mooring 1F and each mooring result in maximum r in the ST-band that range between $0.6 < r_{max}^{ST} < 1.0$ with corresponding lags $0 < \tau_{DU} < 53$ hr, $0.3 < r_{max}^{DU} < 0.9$ with corresponding lags $-4.2 < \tau_{DU} < 1.3$ hr in the DU-band, $0 < r_{max}^{SD} < 0.7$ with lags $-2.8 < \tau_{SD} < 4.2$ hr for the SD-band (Figure 4.7). Negative τ corresponds with signals at 1F leading. The correlations at spatial scales, where T_{da}^{ST} in the south leads T_{da}^{ST} in the north by ~ 50 hr, correspond to the alongshore, poleward propagation of warm plumes from the SBC (Figure 4.7a). The correlation pattern in the DU-band (Figure 4.8b) is suggestive of the shape of the coastline and cross-shore process, owing to the fact that the moorings farthest west lag moorings farthest east. Although the correlations are weak ($r_{max}^S < 0.5$) farther from 1F, the patterns suggest that the SD-band signal leads at the headlands, which are farther west. This is suggestive of onshore propagation in the SD-band. Spatial scales of the array range from 150 m to 54 km. The decorrelation length scales (L_d^{DU}, L_d^{SD}) are estimated from bin-averaged (3 km) r_{max}^{DU} and r_{max}^{SD} calculated for each mooring pair. The bin-averaged r_{DU}^{max} suggest T_{da}^{DU} slowly decorrelates, yet remains correlated ($r_{DU}^{max} \geq 0.5$) at scales of 54 km (Figure 4.8a). There is a significant (at the 95% confidence level) increase in r_{max}^{DU}

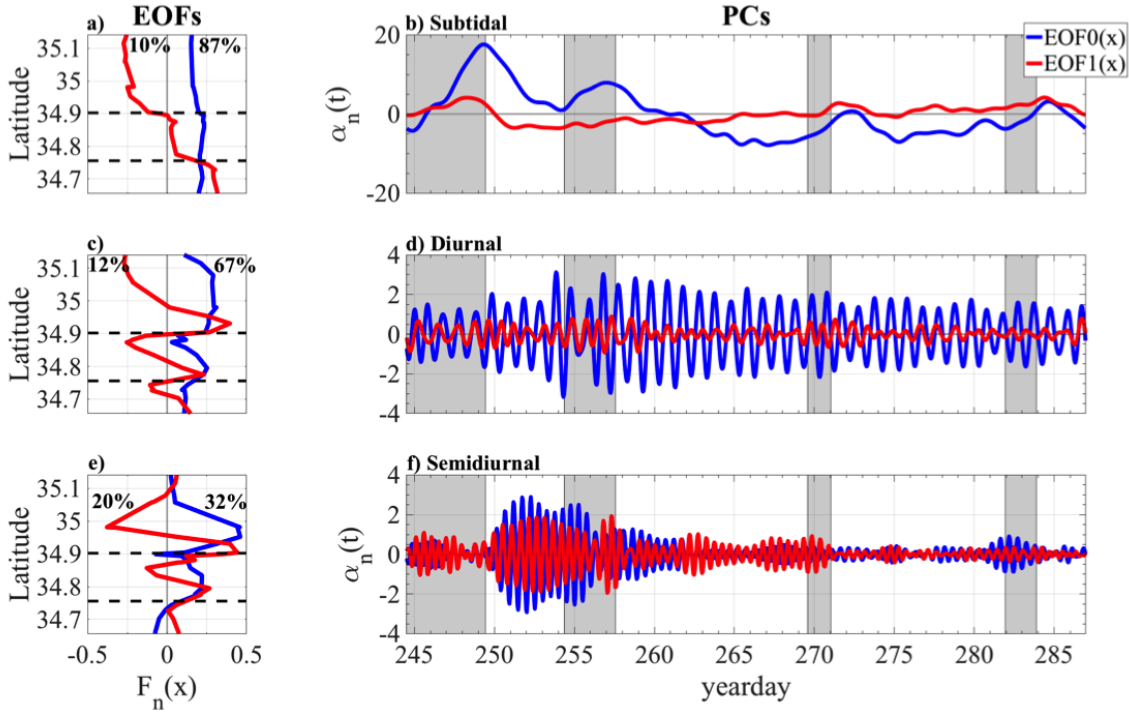


Figure 4.6. The two-leading EOF modes plotted as a function of alongshore location and corresponding PCs at (a,b) subtidal, (c,d) diurnal, and (e,f) semidiurnal frequencies. The percent of variance explained by a mode is indicated near the top of (a), (c), and (e). Black dashed lines in (a), (c), and (e) indicate the location of Pt. Sal and Purisima Pt, and grey boxes indicate times of upwelling relaxations in (b), (d), and (f).

between 30 - 40 km. Bin-averaged r_{max}^{SD} indicates that T_{da}^{SD} decorrelates on spatial scales of approximately 10 km (Figure 4.8b).

4.4 Discussion

At all timescales, PCA over the vertical suggests that temperature variability is depth uniform on the SMB inner shelf, where $EOF0(z)$ accounts for $> 80\%$ of the variance in each band. During times of upwelling, the inner shelf is the region inshore of the upwelling front, and is typically well mixed at ST-timescales (Lentz and Fewings 2012). During relaxations, pressure gradients drive poleward propagating plumes of warm water from the SBC (Melton et al. 2009). The warm plumes are attached to the coast, and tend to occupy the entire water column on the inner shelf (Suanda et al. 2016; Washburn et al. 2011). At DU-timescales,

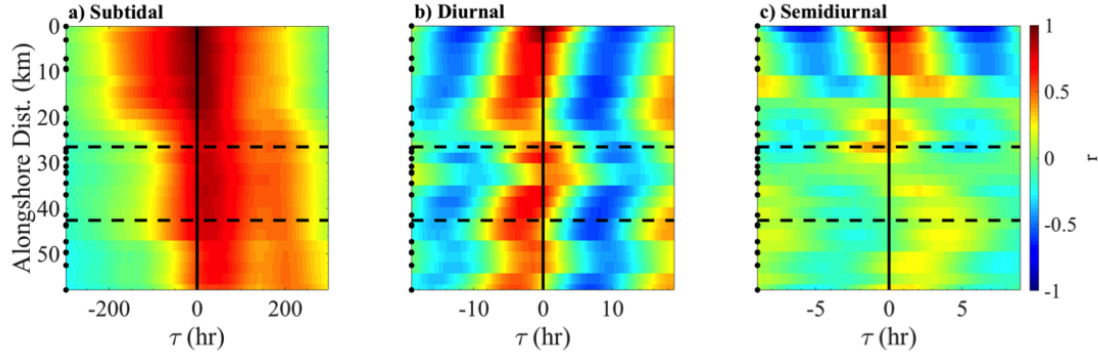


Figure 4.7. Cross-correlation functions $r(\tau)$ plotted as a function of alongshore distance from 1F and lags, τ for a) ST-band, b) DU-band and c) SD-band. Negative lags correspond to the 1F signal leading. Vertical, solid line marks $\tau = 0$ hr, and the horizontal, black dashed lines indicate the locations of Pt. Sal and Purisima Pt.

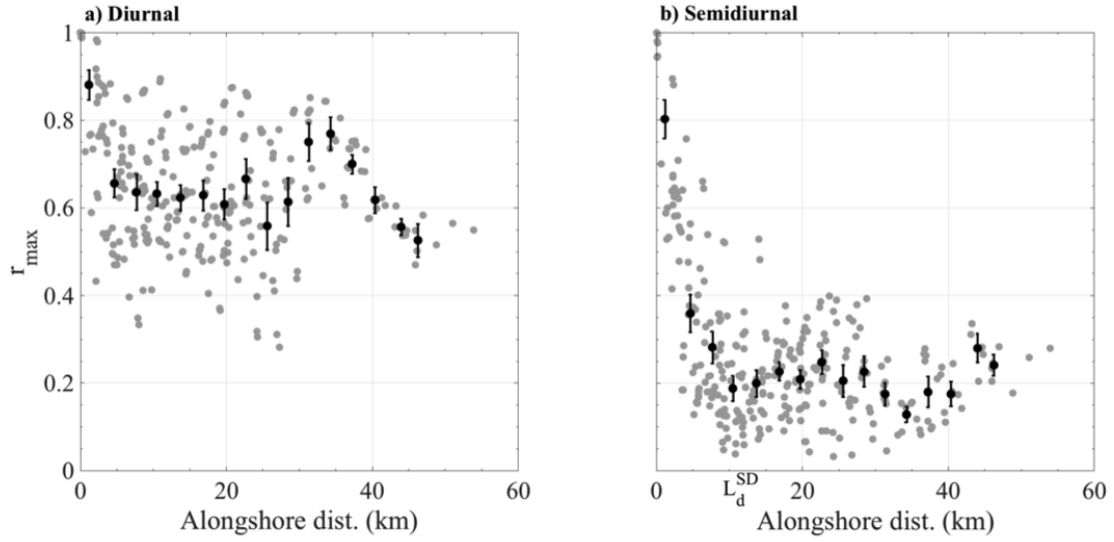


Figure 4.8. Maximum cross-correlation coefficient, r_{max} , plotted as a function of alongshore distance for a) T_{da}^{DU} and b) T_{da}^{SD} . Grey dots are the data, and black dots are bin-averaged over 3 km. Vertical black lines denote the 95% confidence interval. The SD-band decorrelation length scale L_d^{SD} is marked in (b).

Kaplan et al. (2003) found that diurnal winds could drive vertical mixing as deep as 15 m on the inner shelf. At SD-timescales, cold water associated with the run-up of internal tide bores can occupy the entire water column on the inner shelf, although the velocity structure

is still described as a mode-1 internal wave over the vertical (Sinnott et al. 2018).

Transitions between upwelling and relaxations drive the ST-temperature fluctuations as large as 4°C with the largest fluctuations occurring south of Purisima Pt. (Figure 4.4b). Concomitant with four wind relaxations, are four periods of warm water propagating poleward with speeds between 15 to 28 km day⁻¹. Cross-correlations suggest that fluctuations T_{da}^{ST} propagate from south to north at 25 km day⁻¹. Upwelling and relaxation patterns on the central California inner shelf described here are consistent with previous observations between Pt. Sal and the SBC (Melton et al. 2009; Washburn et al. 2011) and numerical model results (Suanda et al. 2016) of the thermal response to relaxations. Washburn et al. (2011) observed ΔT associated with poleward propagating plumes as large as 4°C and propagation speeds between 15-30 km day⁻¹ as far north as Pt. Sal.

Along the array, $\sigma_{DU}^2(T_{da})$ ranges from 0.03 to 0.20°C², and accounts for approximately as much as 13% of $\sigma^2(T_{da})$ at each mooring. The shape of $EOF0^{DU}(x)$ (Figure 4.6c) indicates the largest values of $\sigma_{DU}^2(T_{da})$ are along coastlines II and IV, and the smallest are along coastlines I, III, V, and VI (Figure 4.1). North of Pt. Sal, $\sigma_{DU}^2(T_{da})$ is greater percentage of $\sigma^2(T_{da})$ than south of Pt. Sal. There are similar trends in the SD-band, where $\sigma_{SD}^2(T_{da})$ ranges from 0.02 to 0.18°C² along the array, and accounts for as much as 11% of $\sigma^2(T_{da})$. Along coastlines II and IV, $\sigma_{SD}^2(T_{da})$ is largest. During the times of largest $W_{DU}(T_{da})$ and $W_{SD}(T_{da})$ (yeardays 250 - 264, Figure 4.5), $\sigma_{DU}^2(T_{da})$ and $\sigma_{SD}^2(T_{da})$ each account for as much as 34% of $\sigma^2(T_{da})$.

Patterns in the cross-correlations of T_{da}^{DU} and T_{da}^{SD} (Figure 4.7b,c) are suggestive of the shape of the coastline between Pismo and Pt. Arguello (Figure 4.1a). Temporal lags in the DU-band, where T_{da}^{DU} along coastline II, IV, and VI lags T_{da}^{DU} along coastlines I, III, and V; likely reflect the differences in the onset of the local sea breeze and the thermal response, however the fluctuations are coherent over scales > 54 km (Figure 4.8a). Differences in background stratification, both horizontal and vertical, can lead to differences in onset of vertical mixing on the inner shelf (Horwitz and Lentz 2014). Patterns in SD-correlations suggest T_{da}^{SD} at the headlands leads T_{da}^{SD} in the capes (Figure 4.7c). Colosi et al. (2018b) found that semidiurnal internal tide bores propagated onshore and shore normal at Pt. Sal during the summer of 2015 in water depths between 20 - 50 m. The patterns presented here are consistent with results of Colosi et al. (2018b). The SD-band temperature fluctuations

decorrelate at ~ 10 km scales. (Filonov et al. 2014) found that structure of internal waves varied over scales < 10 km within a bay owing to differences in background stratification, bathymetry, and local wind forcing.

The findings here are consistent with a study of DU- and SD-band temperature variability in the SBC (Aristizabal et al. 2016). They found near-bottom σ_{DU}^2 and σ_{SD}^2 between 0.3 and 0.6°C^2 along the mainland of the SBC and $< 0.1^\circ\text{C}^2$ near Pt. Sal and Purisima Pt., and the fluctuations were not coherent between Pt. Sal and Purisima Pt. and mainland SBC. Along the mainland, however, DU-band fluctuations were coherent over ~ 80 km. Aristizabal et al. (2016) attributed T_{da}^{DU} fluctuations to the sea breeze, and the spatial scales were representative of the differences in local diurnal wind amplitudes along the site. In the SD-band, Aristizabal et al. (2016) found that temperature fluctuations were not coherent along the mainland, whereas they were coherent around the Channel Islands (Figure 4.1b). In the SBC, semidiurnal internal tides are generated at the Channel Islands (Buijsman et al. 2012), whereas the semidiurnal internal tide along the SMB is likely generated along the continental shelf break based on observed propagation paths (Colosi et al. 2018b).

The SMB is characterized by upwelling and relaxation dynamics. It is well known that coastal upwelling is an important mechanism for supplying nutrients to the nearshore at ST-timescales. Additionally, internal tide bores are effective at transporting nutrients to the inner shelf (Lucas et al. 2011). At long timescales ($> 60hr$), the accumulated transport onto the inner shelf owing to SD internal tides can overwhelm the ST transport near Pt. Sal (Freismuth et al. submitted). Alongshore differences in temperature fluctuations at various time and spatial scales are indicative of processes that can impact the community structure and alongshore distribution of marine biota (Tapia et al. 2014), as well as DU-band processes impacting the vertical distribution of marine biota (Lagadeuc et al. 1997). Understanding hydrographic variations at ST, DU, and SD timescales and varying space scales of the inner shelf is important to understanding the ecological response to the variations.

4.5 Summary

Twenty five moorings with temperature sensors mounted over the vertical were deployed in 9 to 16-m water depths over 54 km of the central California inner shelf to measure the temporal and spatial (both vertical and horizontal) variability in temperature (Figure 4.1).

The goal was to evaluate how inter-connected temperature fluctuations occur along the coast.

Temperature variability, regardless of temporal scales, can be considered mostly depth uniform for these shallow-water depths. The depth-averaged, ST temperature variability describes over 80% of the total temperature variability (Figure 4.9), where the combined DU and SD variability account for 10%. The temporal structure varies amongst the frequency bands but suggests that much of the variability, when adjusted for temporal lags, is spatially coherent. The spatial coherence decreases with increasing frequency of temperature variability. Scale-averaged wavelet power reveals that DU- and SD-band fluctuations are greatest following upwelling favorable wind relaxations, and periods of DU-band fluctuations in T_{da} are consistent with times of diurnal winds.

The ST-band temperature is associated with alongshore propagation of warm plumes that originate from the SBC owing to pressure gradients induced during upwelling. The along-shore amplitude of the subtidal variability is related to the distance to the SBC, decreasing with increasing distance from the SBC.

For the DU-band, likely driven by the cross-shore sea breeze, as the temporal lags are associated with coastline distance westward, where the shorelines that protrude the farther west develop DU variability last. The r_{max}^{DU} values suggest the DU temperature variability is significantly correlated along the entire 54 km of observations. This implies that DU-band forcing is relatively uniform, however three general trends in the DU-band variance as measure by $EOF0(x)$ result from the headland-bay topography along this coast. First, the variance decreases from north to south along the experiment site. Second, variance decreases from north to south by $\sim 65\%$ along the headland stump. Third, variance increases from north to south by $\sim 65\%$ between adjacent headlands. These trends are likely a result of the local amplitudes of the sea breeze relative to coastline orientation near the headlands.

For the SD-band, this is also believed to be a cross-shore process driven by internal tide bores propagating onshore near normal to the shoreline. This results in temporal lags where the shorelines that protrude the farther west develop SD variability first. The SD-band variance exhibits a similar north-south trend as the DU-band variance. The spatial trends in SD-band variance are likely a result of differences in background properties, such as stratification and bathymetry, along the propagation path. Furthermore, the $r_{max}(\tau, x)$ values in the SD-band

suggest the variability decorrelates on ~ 10 km scales. This decorrelation scale is consistent with length scales of stretches of coastlines with similar orientation and geomorphology.

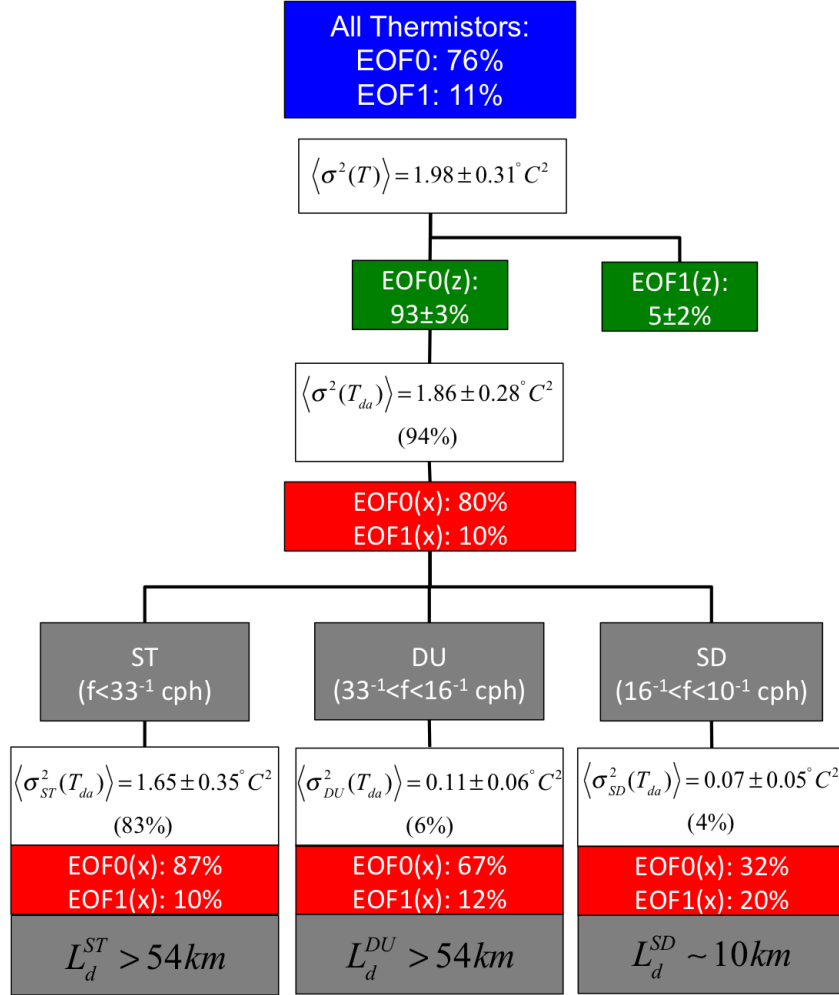


Figure 4.9. Chart summarizes $\langle \sigma \rangle$ (white boxes), percent of variance explained for vertical EOF modes (EOF(z) green boxes), and percent of variance explained for the horizontal EOF modes (red boxes). Spatial averaging of σ^2 is denoted by $\langle \cdot \rangle$, and the standard deviation of $\langle \sigma^2 \rangle$ reported. Subscripts *ST*, *DU*, and *SD* indicate subtidal, diurnal, and diurnal band variances, and *da* indicates depth-averaged temperature. ST-, DU-, and SD-band decorrelation scales (L_d^{ST} , L_d^{DU} , and L_d^{SD} , respectively) are noted on the bottom row.

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CHAPTER 5:

Conclusions

Temperature variability and cross-shore heat fluxes on the central California inner shelf are described from an unprecedented array of in situ field observations of seawater temperature and currents during two field campaigns, summer of 2015 and fall of 2017. Cross-shore heat fluxes at subtidal, diurnal, and semidiurnal timescales are estimated from the 50- to the 5-m isobath using the novel approach of scale-averaged wavelet co-scalograms that accounts for the non-stationarity of the processes. Previous studies of inner shelf, cross-shore heat flux have primarily focused solely on only either subtidal (e.g., Melanie R. Fewings and Lentz 2011), diurnal (e.g., Suanda et al. 2011), or semidiurnal (e.g., Lucas et al. 2011) timescales. In this study, relative contributions to heating and cooling on the inner shelf owing to cross-shore heat fluxes at all timescales are discussed for the first time. The first observations of warm internal tide bores at the edge of the surf zone are presented, and are the dominant mechanism for driving onshore heat fluxes on the central California inner shelf during the summer of 2015. The alongshore variability in temperature fluctuations at subtidal, diurnal, and semidiurnal timescales are investigated along 54 km of the inner shelf. This unique set of observations reveals the characteristic length scales of variability in a headland-and-cape topography during upwelling and relaxation conditions. These results are compared to diurnal and semidiurnal temperature fluctuations in the nearby Santa Barbara Channel (Aristizabal et al. 2016, 2017). Details of the results from this dissertation are summarized below.

In Chapter 2, the spatial and temporal contributions of cross-shelf heat flux on the inner shelf were investigated. Measures of co-located temperature and currents over the vertical were obtained for 35 days in a cross-shore array from the 50-m to 7-m isobaths near the rocky headland of Pt. Sal, California, during the summer of 2015. Cross-shore heat fluxes are estimated on the mid- and inner shelf at subtidal, diurnal, semidiurnal timescales using a wavelet co-scalogram technique to account for the non-stationarity of the forcing mechanisms. The width of the inner shelf averaged 1.9 km with variation up to ~6 km owing to changes in the background wind stress and stratification. Upwelling drives a subtidal cooling (negative flux) on the mid- and inner shelf, while wind relaxations result

in heating (positive fluxes). The sea breeze drives a negative diurnal flux on the mid-shelf and a positive flux on the inner shelf, as well as generated diurnal period internal waves. Semidiurnal internal tide bores typically drive a positive flux. Periods of negative flux indicate possible polarity reversal of the shoaling internal tide. A mean inner shelf heat budget reveals that a residual flux, likely associated with an alongshore heat flux, balances the cross-shore and surface heat flux contributions. Fluctuating heat budgets on the mid- and inner shelf reveal that the cross-shore and surface heat fluxes contribute more than 50% of the total flux to heat storage. At timescales of ~ 5 days, the diurnal and semidiurnal cross-shore heat fluxes contribute to a net input of energy on the inner shelf, while heat is being removed from the midshelf during upwelling. These results show that diurnal and semidiurnal processes are important for the cross-shore exchange of biota, nutrients, pollution, and heat at longer timescales.

In Chapter 3, the first observations of warm internal tide bores at the edge of the surf zone were presented. Nonlinear internal waves (NLIWs) were measured from the 50-m to the 5-m isobaths with dense arrays of thermistor strings and co-located Acoustic Doppler Current Profilers near Pt. Sal, California. NLIWs are observed either as cold boluses (waves of elevation), warm internal tide bores (ITBs, waves of depression), or as a mixture of both. Unlike for cold boluses, the fate and effect of ITBs has not been well quantified within the nearshore region. In this study, ITBs are shown to propagate to the surf zone edge and heat nearshore waters by 1.2°C which is comparable in magnitude to surf zone cooling by cold boluses observed in recent research. Observed vertical temperature overturns uniquely highlight the importance of ITBs for mixing as well as the alongshore heterogeneity in mixing owing to bathymetry. The findings have important implications regarding cross-shore nutrient transport and biological recruitment at the surf zone edge.

In Chapter 4, temperature variability along the central California inner shelf was discussed. Vertical profiles of water temperature, T , were obtained along 54 km of the central California inner shelf during the fall of 2017. Temperature fluctuations were studied in the subtidal (ST), diurnal (DU), and semidiurnal (SD) bands, at depths of 9 - 16 m, and over spatial scales from 150 m to 54 km. A mode 0 empirical orthogonal function (EOF) explains 96%, 78%, and 84% of the variance over the vertical at ST-, DU-, and SD-timescales, implying depth uniformity. Depth-averaged, subtidal variance $[\sigma_{ST}^2(T_{da})]$ accounts for as much as 83% of the total variance (σ^2), whereas $\sigma_{DU}^2(T_{da})$ and $\sigma_{SD}^2(T_{da})$ combined account for

10% of σ^2 . Mode 0 EOFs over the horizontal explain 87%, 67%, and 32% of the variance in the ST-, DU-, and SD-bands, implying spatial heterogeneities at larger frequencies. DU-band T_{da} does not decorrelate over the 54-km field site, whereas the SD-band T_{da} decorrelates on 10-km scales. Scale-averaged wavelet power reveals that DU- and SD-band fluctuations are greatest following upwelling favorable wind relaxations, and periods of DU-band fluctuations in T_{da} are consistent with times of diurnal winds. These new observations are compared to a recent study of temperature variability in the neighboring Santa Barbara Channel. These current findings have implications for the locations and timescales of marine biota migration, retention, and settlement.

The results presented in this current study detail the representativeness of seawater temperature variability and the transport mechanisms driving the variability on subtidal, diurnal, and semidiurnal timescales in a headland-bay topography on the eastern boundary of an oceanic basin. The semidiurnal internal tide dominates the temperature variability and cross-shore heat flux. This finding suggests that cross-shore exchange originating from the surf zone is second order to the internal wave driven exchange from the midshelf onto the inner shelf. Although not presented in this dissertation, observations of seawater temperature on the inner shelf in southern Monterey Bay yielded similar results where the semidiurnal internal tide dominated the temperature signal. This suggests that this ubiquitous phenomena can have significant impact on inner shelf temperature variability. It is important note there is an inherent connection between inner shelf processes at different time scales owing to changing stratification at varying timescales.

Since the Cudaback et al. (2005) statement that the inner shelf was too shallow for the oceanographers and too deep for the coastal scientists, there have been great advances in understanding inner shelf processes and cross-shore exchange lead by the recent pioneering work of Steven Lentz and his students. In fact, during the period of the Inner Shelf DRI field experiments, four important papers were published regarding inner shelf temperature variability and transport. There have been prolific inner shelf investigations on the Martha's Vineyard shelf (e.g., Lentz and Fewings 2012, and references therein), amongst a few other locations, as well as the findings from this current work and other emerging results from the Inner Shelf DRI will soon follow. I speculate that mixing and turbulence will remain the major unknown in inner shelf research. This is owing to the difficulty in separating turbulent velocities from surface gravity wave orbital velocities. However, new innovative

measurement schemes coupled with new instrumentation used during the Inner Shelf DRI may provide the valuable insight of eddy viscosity over the vertical. Mixing is critical as it can modify stratification, where the presence of a pycnocline is typically the difference between mixing and cross-shore transport. The diel modulation of a surface mixed layer owing to the diurnal sea breeze and surface heat flux may have a significant impact on the efficacy of cross-shore transport either by internal waves or wind-driven processes.

List of References

- Aghsaee, P., L. Boegman, and K. G. Lamb, 2010: Breaking of shoaling internal solitary waves. *J. Fluid Mech.*, **659**, 289–317, doi: 10.1017/S002211201000248X.
- Aristizabal, M. F., M. R. Fewings, and L. Washburn, 2016: Contrasting spatial patterns in the diurnal and semidiurnal temperature variability in the Santa Barbara Channel, California. *J. Geophys. Res.*, **121** (1), 427–440, doi: 10.1002/2015JC011239.
- Aristizabal, M. F., M. R. Fewings, and L. Washburn, 2017: Effects of the relaxation of upwelling — favorable winds on the diurnal and semidiurnal water temperature fluctuations in the Santa Barbara Channel, California. *J. Geophys. Res.*, **122** (10), 7958–7977, doi: 10.1002/2017JC013199.
- Austin, J. A., 1999: The role of the alongshore wind stress in the heat budget of the North Carolina inner shelf. *J. Geophys. Res.*, **104** (C8), 18 187–18 203, doi: 10.1029/1998JC900122.
- Austin, J. A., and S. J. Lentz, 2002: The inner shelf response to wind-driven upwelling and downwelling. *J. Phys. Oceanogr.*, **32** (7), 2171–2193, doi: 10.1175/1520-0485(2002)032<2171:TISRTW>2.0.CO;2.
- Beckenbach, E., and E. Terrill, 2008: Internal tides over abrupt topography in the Southern California Bight: Observations of diurnal waves poleward of the critical latitude. *J. Geophys. Res.*, **113** (C2), doi: 10.1029/2006JC003905.
- Black, K., J. Oldman, and T. Hume, 2005: Dynamics of a 3-dimensional, baroclinic, headland eddy. *New Zeal. Jour. Mar. Fresh. Res.*, **39** (1), 91–120, doi: 10.1080/00288330.2005.9517294.
- Blanchette, C. A., C. M. M. Miner, P. T. Raimondi, D. Lohse, K. E. K. Heady, and B. Bernardo R. Broitman, 2008: Biogeographical patterns of rocky intertidal communities along the Pacific coast of North America. *J. Biogeog.*, **35** (9), 1593–1607, doi: 10.1111/j.1365-2699.2008.01913.x.
- Boegman, L., G. N. Ivey, and J. Imberger, 2005: The degeneration of internal waves in lakes with sloping topography. *Limnol. and Oceanogr.*, **50** (5), 1620–1637, doi: 10.4319/lo.2005.50.5.1620.
- Boehm, A. B., D. B. Lluch-Cota, K. A. Davis, C. D. Winant, and S. G. Monismith, 2004: Covariation of coastal water temperature and microbial pollution at interannual to tidal periods. *Geophys. Res. Lett.*, **31** (6), doi: 10.1029/2003GL019122.

- Boehm, A. B., B. F. Sanders, and C. D. Winant, 2002: Cross-shelf transport at Huntington Beach. Implications for the fate of sewage discharged through an offshore ocean outfall. *Environ. Sci. & Tech*, **36** (9), 1899–1906, doi: 10.1021/es0111986.
- Bonicelli, J., C. Moffat, S. A. Navarrete, J. L. Largier, and F. J. Tapia, 2014a: Spatial differences in thermal structure and variability within a small bay: Interplay of diurnal winds and tides. *Cont. Shelf Res.*, **88**, 72–80, doi: 10.1016/j.csr.2014.07.009.
- Bonicelli, J., F. J. Tapia, and S. A. Navarrete, 2014b: Wind-driven diurnal temperature variability across a small bay and the spatial pattern of intertidal barnacle settlement. *J. Exper. Mar. Biol. Ecol.*, **461**, 350–356, doi: 10.1016/j.jembe.2014.09.003.
- Brink, K., 2016: Cross-shelf exchange. *Annu. Rev. Mar. Sci.*, **8** (1), 59–78, doi: 10.1146/annurev-marine-010814-015717.
- Buijsman, M. C., Y. Uchiyama, J. C. McWilliams, and C. R. Hill-Lindsay, 2012: Modeling semidiurnal internal tide variability in the Southern California Bight. *J. Phys. Oceanogr.*, **42** (1), 62–77, doi: 10.1175/2011JPO4597.1.
- Casey, K. S., and D. Adamec, 2002: Sea surface temperature and sea surface height variability in the North Pacific Ocean from 1993 to 1999. *J. Geophys. Res.*, **107** (C8), 14–1–14–12, doi: 10.1029/2001JC001060.
- Colosi, J. A., R. C. Beardsley, J. F. Lynch, G. Gawarkiewicz, C.-S. Chiu, and A. Scotti, 2001: Observations of nonlinear internal waves on the outer New England continental shelf during the summer shelfbreak primer study. *J. Geophys. Res.*, **106** (C5), 9587–9601, doi: 10.1029/2000JC900124.
- Colosi, J. A., N. Kumar, S. H. Suanda, T. M. Freismuth, and J. H. MacMahan, 2018a: Statistics of internal tide bores and internal solitary waves observed on the inner continental shelf off Point Sal, California. *Journal of Physical Oceanography*, **48** (1), 123–143, doi: 10.1175/JPO-D-17-0045.1.
- Colosi, J. A., N. Kumar, S. H. Suanda, T. M. Freismuth, and J. H. MacMahan, 2018b: Statistics of internal tide bores and internal solitary waves observed on the inner continental shelf off Point Sal, California. *Journal of Physical Oceanography*, **48** (1), 123–143, doi: 10.1175/JPO-D-17-0045.1.
- Connolly, T. P., and S. J. Lentz, 2014: Interannual variability of wintertime temperature on the inner continental shelf of the Middle Atlantic Bight. *J. Geophys. Res.*, **119** (9), 6269–6285, doi: 10.1002/2014JC010153.
- Cudaback, C. N., L. Washburn, and E. Dever, 2005: Subtidal inner-shelf circulation near Point Conception, California. *J. Geophys. Res.*, **110** (C10).

- Davies-Jones, R., 1993: Useful, formulas for computing divergence, vorticity, and their errors from three or more stations. *Mon. Wea. Rev.*, **121** (3), 713–725, doi: 10.1175/1520-0493(1993)121<0713:UFFCDV>2.0.CO;2.
- Dean, R. G., and R. A. Dalrymple, 1991: *Water Wave Mechanics for Engineers and Scientists*, Vol. 2. World Scientific Publishing Co Inc.
- Dever, E., and S. Lentz, 1994: Heat and salt balances over the northern California shelf in winter and spring. *J. Geophys. Res.*, **99** (C8), 16 001–16 017, doi: 10.1029/94JC01228.
- Dickey, T. D., D. V. Manov, R. A. Weller, and D. A. Siegel, 1994: Determination of long-wave heat flux at the air-sea interface using measurements from buoy platforms. *J. Atmos. Oceanic Technol.*, **11** (4), 1057–1078, doi: 10.1175/1520-0426(1994)011<1057:DOLHFA>2.0.CO;2.
- Fairall, C. W., E. F. Bradley, J. E. Hare, A. A. Grachev, and J. B. Edson, 2003: Bulk parameterization of air–sea fluxes: Updates and verification for the coare algorithm. *J. Climate*, **16** (4), 571–591, doi: 10.1175/1520-0442(2003)016<0571:BPOASF>2.0.CO;2.
- Fewings, M., S. J. Lentz, and J. Fredericks, 2008: Observations of cross-shelf flow driven by cross-shelf winds on the inner continental shelf. *J. Phys. Oceanogr.*, **38** (11), 2358–2378, doi: 10.1175/2008JPO3990.1.
- Fewings, M. R., L. Washburn, and J. C. Ohlmann, 2015: Coastal water circulation patterns around the Northern Channel Islands and Point Conception, California. *Prog. in Oceanogr.*, **138** (Part A), 283–304, doi: 10.1016/j.pocean.2015.10.001.
- Filonov, A., M. Lavan, L. Ladah, and I. Tereshchenko, 2014: Spatial variability of internal waves in an open bay with a narrow steep shelf in the Pacific off NW Mexico. *Cont. Shelf Res.*, **78**, 1–15, doi: 10.1016/j.csr.2014.01.015.
- Freismuth, T. M., J. H. MacMahan, J. A. Colosi, and M. K. Gough, submitted: Cross-shore observations of warm, internal tide bores within the nearshore. *J. Phys. Oceanogr.*, **n/a** (n/a), n/a–n/a, doi: n/a.
- Gan, J., and J. Allen, 2002: A modeling study of shelf circulation off northern California in the region of the Coastal Ocean Dynamics Experiment: Response to relaxation of upwelling winds. *J. Geophys. Res.*, **107** (C9), 6–1–6–31, doi: 10.1029/2000JC000768.
- Helfrich, K. R., 1992: Internal solitary wave breaking and run-up on a uniform slope. *J. Fluid Mech.*, **243**, 133–154, doi: 10.1017/S0022112092002660.
- Helfrich, K. R., and W. K. Melville, 2006: Long nonlinear internal waves. *Annu. Rev. Fluid Mech.*, **38** (1), 395–425, doi: 10.1146/annurev.fluid.38.050304.092129.

- Hendrickson, J., and J. MacMahan, 2009: Diurnal sea breeze effects on inner-shelf cross-shore exchange. *Cont. Shelf Res.*, **29** (18), 2195–2206.
- Herdman, L. M. M., J. L. Hench, and S. G. Monismith, 2015: Heat balances and thermally driven lagoon-ocean exchanges on a tropical coral reef system (Moorea, French Polynesia). *J. Geophys. Res.*, **120** (2), 1233–1252, doi: 10.1002/2014JC010145.
- Hickey, B. M., and N. E. Pola, 1983: The seasonal alongshore pressure gradient on the West Coast of the United States. *J. Geophys. Res.*, **88** (C12), 7623–7633, doi: 10.1029/JC088iC12p07623.
- Hodur, R. M., 1997: The Naval Research Laboratory's Coupled Ocean/Atmosphere Mesoscale Prediction System (COAMPS). *Mon. Wea. Rev.*, **125** (7), 1414–1430, doi: 10.1175/1520-0493(1997)125<1414:TNRLSC>2.0.CO;2.
- Holloway, P. E., 1987: Internal hydraulic jumps and solitons at a shelf break region on the Australian North West Shelf. *J. Geophys. Res.*, **92** (C5), 5405–5416, doi: 10.1029/JC092iC05p05405.
- Horwitz, R., and S. J. Lentz, 2014: Inner-shelf response to cross-shelf wind stress: The importance of the cross-shelf density gradient in an idealized numerical model and field observations. *J. Phys. Oceanogr.*, **44** (1), 86–103, doi: 10.1175/JPO-D-13-075.1.
- Horwitz, R. M., and S. J. Lentz, 2016: The effect of wind direction on cross-shelf transport on an initially stratified inner shelf. *J. Mar. Res.*, **74** (4), 201–227.
- Hudgins, L., C. A. Friehe, and M. E. Mayer, 1993: Wavelet transforms and atmospheric turbulence. *Phys. Rev. Lett.*, **71** (20), 3279.
- Huthnance, J. M., 1989: Internal tides and waves near the continental shelf edge. *Geophys. & Astrophys. Fluid Dyn.*, **48** (1-3), 81–106, doi: 10.1080/03091928908219527.
- Kao, T. W., F.-S. Pan, and D. Renouard, 1985: Internal solitons on the pycnocline: Generation, propagation, and shoaling and breaking over a slope. *J. Fluid Mech.*, **159**, 395–425, doi: 10.1017/S0022112085003081.
- Kaplan, D. M., J. L. Largier, S. Navarrete, R. Guíñez, and J. C. Castilla, 2003: Large diurnal temperature fluctuations in the nearshore water column. *Estuarine, Coastal and Shelf Sci.*, **57** (3), 385–398.
- Kumar, N., F. Feddersen, S. Suanda, Y. Uchiyama, and J. McWilliams, 2016: Mid- to inner-shelf coupled ROMS-SWAN model-data comparison of currents and temperature: Diurnal and semidiurnal variability. *J. Phys. Oceanogr.*, **46** (3), 841–862, doi: 10.1175/JPO-D-15-0103.1.

- Kundu, P. K., and J. S. Allen, 1976: Some three-dimensional characteristics of low-frequency current fluctuations near the Oregon coast. *J. Phys. Oceanogr.*, **6** (2), 181–199, doi: 10.1175/1520-0485(1976)006<0181:STDCOL>2.0.CO;2.
- Kunze, E., 1985: Near-inertial wave propagation in geostrophic shear. *J. Phys. Oceanogr.*, **15** (5), 544–565, doi: 10.1175/1520-0485(1985)015<0544:NIWPIG>2.0.CO;2.
- Lagadeuc, Y., M. Boute, and J. Dodson, 1997: Effect of vertical mixing on the vertical distribution of copepods in coastal waters. *J. Plank. Resear.*, **19** (9), 1183–1204, doi: 10.1093/plankt/19.9.1183.
- Lamb, K. G., 1994: Numerical experiments of internal wave generation by strong tidal flow across a finite amplitude bank edge. *J. Geophys. Res.*, **99** (C1), 843–864, doi: 10.1029/93JC02514.
- Lamb, K. G., 2002: A numerical investigation of solitary internal waves with trapped cores formed via shoaling. *J. Fluid Mech.*, **451**, 109–144, doi: 10.1017/S002211200100636X.
- Lamb, K. G., 2003: Shoaling solitary internal waves: on a criterion for the formation of waves with trapped cores. *J. Fluid Mech.*, **478**, 81–100, doi: 10.1017/S0022112002003269.
- Lamb, K. G., 2014: Internal wave breaking and dissipation mechanisms on the continental slope/shelf. *Annu. Rev. Fluid Mech.*, **46** (1), 231–254, doi: 10.1146/annurev-fluid-011212-140701.
- Large, W. G., and S. Pond, 1981: Open ocean momentum flux measurements in moderate to strong winds. *J. Phys. Oceanogr.*, **11** (3), 324–336, doi: 10.1175/1520-0485(1981)011<0324:OOMFMI>2.0.CO;2.
- Laxague, N. J., B. K. Haus, D. G. Ortiz-Suslow, and H. C. Graber, submitted: Unpacking observed air-sea momentum flux in frequency and time via wavelet analysis. *J. Geophys. Res.*, **n/a**, n/a–n/a, doi: n/a.
- Lentz, S. J., 1987: A heat budget for the Northern California shelf during CODE 2. *J. Geophys. Res.*, **92** (C13), 14 491–14 509, doi: 10.1029/JC092iC13p14491.
- Lentz, S. J., 1992: The surface boundary layer in coastal upwelling regions. *J. Phys. Oceanogr.*, **22** (12), 1517–1539, doi: 10.1175/1520-0485(1992)022<1517:TSBLIC>2.0.CO;2.
- Lentz, S. J., 2008: Seasonal variations in the circulation over the Middle Atlantic Bight continental shelf. *J. Phys. Oceanogr.*, **38** (7), 1486–1500, doi: 10.1175/2007JPO3767.1.

- Lentz, S. J., and M. R. Fewings, 2012: The wind- and wave-driven inner-shelf circulation. *Annu. Rev. Mar. Sci.*, **4**, 317–343, doi: 10.1146/annurev-marine-120709-142745.
- Lerczak, J. A., M. C. Hendershott, and C. D. Winant, 2001: Observations and modeling of coastal internal waves driven by a diurnal sea breeze. *J. Geophys. Res.*, **106** (C9), 19 715–19 729, doi: 10.1029/2001JC000811.
- Lucas, A. J., P. J. S. Franks, and C. L. Dupont, 2011: Horizontal internal-tide fluxes support elevated phytoplankton productivity over the inner continental shelf. *Limnol. Oceanogr.*, **1** (1), 56–74, doi: 10.1215/21573698-1258185.
- Melanie R. Fewings, M., and S. J. Lentz, 2011: Summertime cooling of the shallow continental shelf. *J. Geophys. Res.*, **116** (C7), doi: 10.1029/2010JC006744.
- Melton, C., L. Washburn, and C. Gotschalk, 2009: Wind relaxations and poleward flow events in a coastal upwelling system on the central California coast. *J. Geophys. Res.*, **114** (C11), doi: 10.1029/2009JC005397.
- Molina, L., G. Pawlak, J. R. Wells, S. G. Monismith, and M. A. Merrifield, 2014: Diurnal cross-shore thermal exchange on a tropical forereef. *J. Geophys. Res.*, **119** (9), 6101–6120, doi: 10.1002/2013JC009621.
- Molinari, R., and A. D. K. Jr., 1975: Calculations of differential kinematic properties from Lagrangian observations in the western Caribbean Sea. *J. Phys. Oceanogr.*, **5** (3), 483–491, doi: 10.1175/1520-0485(1975)005<0483:CODKPF>2.0.CO;2.
- Monismith, S. G., A. Genin, M. A. Reidenbach, G. Yahel, and J. R. Koseff, 2006: Thermally driven exchanges between a coral reef and the adjoining ocean. *J. Phys. Oceanogr.*, **36** (7), 1332–1347, doi: 10.1175/JPO2916.1.
- Nam, S., and U. Send, 2011: Direct evidence of deep water intrusions onto the continental shelf via surging internal tides. *J. Geophys. Res.*, **116** (C5), doi: 10.1029/2010JC006692.
- Nam, S., and U. Send, 2013: Resonant diurnal oscillations and mean alongshore flows driven by sea/land breeze forcing in the coastal Southern California Bight. *J. Phys. Oceanogr.*, **43** (3), 616–630, doi: 10.1175/JPO-D-11-0148.1.
- Nash, J. D., S. M. Kelly, E. L. Shroyer, J. N. Moum, and T. F. Duda, 2012: The unpredictable nature of internal tides on continental shelves. *J. Phys. Oceanogr.*, **42** (11), 1981–2000, doi: 10.1175/JPO-D-12-028.1.
- Noble, M., B. Jones, P. Hamilton, J. Xu, G. Robertson, L. Rosenfeld, and J. Largier, 2009: Cross-shelf transport into nearshore waters due to shoaling internal tides in San Pedro Bay, CA. *Cont. Shelf Res.*, **29** (15), 1768–1785.

- O'Connor, M. I., J. F. Bruno, S. D. Gaines, B. S. Halpern, S. E. Lester, B. P. Kinlan, and J. M. Weiss, 2007: Temperature control of larval dispersal and the implications for marine ecology, evolution, and conservation. *PNAS*, **104** (4), 1266–1271, doi: 10.1073/pnas.0603422104.
- Ohashi, K., and D.-P. Wang, 2004: Circulation in the Santa Maria Basin, California, during 1998. *J. Geophys. Res.*, **109** (C11), n/a–n/a, doi: 10.1029/2004JC002362.
- Omand, M. M., F. Feddersen, R. T. Guza, and P. J. S. Franks, 2012: Episodic vertical nutrient fluxes and nearshore phytoplankton blooms in Southern California. *Limnol. Oceanogr.*, **57** (6), 1673–1688, doi: 10.4319/lo.2012.57.6.1673.
- Pineda, J., 1994: Internal tidal bores in the nearshore: warm-water fronts, seaward gravity currents and the onshore transport of neustonic larvae. *J. Mar. Res.*, **52**, 427–458, doi: 10.1357/0022240943077046.
- Pineda, J., 1999: Circulation and larval distribution in internal tidal bore warm fronts. *Limnol. Oceanogr.*, **44** (6), 1400–1414, doi: 10.4319/lo.1999.44.6.1400.
- Pineda, J., and M. Lopez, 2002: Temperature, stratification and barnacle larval settlement in two Californian sites. *Cont. Shelf Res.*, **22** (8), 1183–1198, doi: 10.1016/S0278-4343(01)00098-X.
- Preisendorfer, R. U., 1988: *Principal Component Analysis in Meteorology and Oceanography* (Developments in Atmospheric Science), Vol. 17. Elsevier, 425 pp.
- Pritchard, M., and R. A. Weller, 2005: Observations of internal bores and waves of elevation on the New England inner continental shelf during summer 2001. *J. Geophys. Res.*, **110** (C3), doi: 10.1029/2004JC002377.
- Saucier, W. J., 1955: *Principles of Meteorological Analysis*. The University of Chicago Press, 438 pp.
- Send, U., R. C. Beardsley, and C. D. Winant, 1987: Relaxation from upwelling in the coastal ocean dynamics experiment. *J. Geophys. Res.*, **92** (C2), 1683–1698, doi: 10.1029/JC092iC02p01683.
- Shroyer, E. L., J. N. Moum, and J. D. Nash, 2009: Observations of polarity reversal in shoaling nonlinear internal waves. *J. Phys. Oceanogr.*, **39** (3), 691–701, doi: 10.1175/2008JPO3953.1.
- Signell, R. P., and W. R. Geyer, 1991: Transient eddy formation around headlands. *J. Geophys. Res.*, **96** (C2), 2561–2575, doi: 10.1029/90JC02029.

- Sinnett, G., and F. Feddersen, 2014: The surf zone heat budget: The effect of wave heating. *Geophys. Res. Lett.*, **41** (20), 7217–7226, doi: 10.1002/2014GL061398.
- Sinnett, G., F. Feddersen, A. J. Lucas, G. Pawlak, and E. Terrill, 2018: Observations of nonlinear internal wave run-up to the surfzone. *J. Phys. Oceanogr.*, **48** (3), 531–554, doi: 10.1175/JPO-D-17-0210.1.
- Smith, K. A., G. Rocheleau, M. A. Merrifield, S. Jaramillo, and G. Pawlak, 2016: Temperature variability caused by internal tides in the coral reef ecosystem of Hanauma Bay, Hawaii. *Cont. Shelf Res.*, **116**, 1 – 12.
- Suanda, S. H., J. A. Barth, and C. B. Woodson, 2011: Diurnal heat balance for the northern Monterey Bay inner shelf. *J. Geophys. Res.*, **116** (C9), doi: 10.1029/2010JC006894.
- Suanda, S. H., and Coauthors, 2016: Wind relaxation and a coastal buoyant plume north of Pt. Conception, CA: Observations, simulations, and scalings. *J. Geophys. Res.*, **121** (10), 7455–7475, doi: 10.1002/2016JC011919.
- Tapia, F., J. L. Largier, M. Castillo, E. Wieters, and S. A. Navarrete, 2014: Latitudinal discontinuity in thermal conditions along the nearshore of central-northern Chile. *PLOS ONE*, **9** (10), 1–11, doi: 10.1371/journal.pone.0110841.
- Thomas, A., D. Byrne, and R. Weatherbee, 2002: Coastal sea surface temperature variability from Landsat infrared data. *Remote Sens. Environ.*, **81** (2), 262–272, doi: 10.1016/S0034-4257(02)00004-4.
- Thomson, R. E., and W. J. Emery, 2014: *Data Analysis Methods in Physical Oceanography*. Newnes.
- Tilburg, C. E., 2003: Across-shelf transport on a continental shelf: Do across-shelf winds matter? *J. Phys. Oceanogr.*, **33** (12), 2675–2688, doi: 10.1175/1520-0485(2003)033<2675:ATOACS>2.0.CO;2.
- Torrence, C., and G. P. Compo, 1998: A practical guide to wavelet analysis. *Bull. Amer. Meteor. Soc.*, **79** (1), 61–78, doi: 10.1175/1520-0477(1998)079<0061:APGTWA>2.0.CO;2.
- Walter, R. K., and P. J. Phelan, 2016: Internal bore seasonality and tidal pumping of subthermocline waters at the head of the Monterey submarine canyon. *Cont. Shelf Res.*, **116**, 42–53, doi: 10.1016/j.csr.2016.01.015.
- Walter, R. K., E. C. Reid, K. A. Davis, K. J. Armenta, K. Merhoff, and N. J. Nidzieko, 2017: Local diurnal wind-driven variability and upwelling in a small coastal embayment. *J. Geophys. Res.*, **122** (2), 955–972, doi: 10.1002/2016JC012466.

- Walter, R. K., B. C. Woodson, P. R. Leary, and S. G. Monismith, 2014: Connecting wind-driven upwelling and offshore stratification to nearshore internal bores and oxygen variability. *J. Geophys. Res.*, **119** (6), 3517–3534, doi: 10.1002/2014JC009998.
- Walter, R. K., C. B. Woodson, R. S. Arthur, O. B. Fringer, and S. G. Monismith, 2012: Nearshore internal bores and turbulent mixing in southern monterey bay. *J. Geophys. Res.*, **117** (C7), doi: 10.1029/2012JC008115.
- Washburn, L., M. R. Fewings, C. Melton, and C. Gotschalk, 2011: The propagating response of coastal circulation due to wind relaxations along the central California coast. *J. Geophys. Res.*, **116** (C12), doi: 10.1029/2011JC007502.
- Weatherly, G. L., and P. J. Martin, 1978: On the structure and dynamics of the oceanic bottom boundary layer. *J. Phys. Oceanogr.*, **8** (4), 557–570, doi: 10.1175/1520-0485(1978)008<0557:OTSADO>2.0.CO;2.
- Witman, J. D., J. J. Leichter, S. J. Genovese, and D. A. Brooks, 1993: Pulsed phytoplankton supply to the rocky subtidal zone: Influence of internal waves. *Proc. Natl. Acad. Sci. (USA)*, **90** (5), 1686–1690.
- Woodson, C., 2018: The fate and impact of internal waves in nearshore ecosystems. *Annu. Rev. Mar. Sci.*, **10** (1), 421–441, doi: 10.1146/annurev-marine-121916-063619.
- Woodson, C., and Coauthors, 2007: Local diurnal upwelling driven by sea breezes in northern Monterey Bay. *Cont. Shelf Res.*, **27** (18), 2289 – 2302, doi: 10.1016/j..2007.05.014.

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