The Causes of Regional Sea-Surface Temperature Anomalies During Wind Relaxation Events Off the U.S. West Coast in Summer

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The Causes of Regional Sea-Surface Temperature Anomalies During Wind Relaxation Events Off the U.S. West Coast in Summer

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The Causes of Regional Sea-Surface Temperature Anomalies During Wind Relaxation Events Off the U.S. West Coast in Summer

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Abstract

The term ‘wind relaxation’ describes weakening or reversal of the prevailing upwelling-favorable winds. Summertime wind relaxations along the U.S. West Coast exhibit an ‘event cycle’ spanning ~12 days. The winds first relax or reverse off the coast of Oregon. Next, the upwelling-favorable winds strengthen off the coast of central California; the strong winds move poleward and reach Oregon in ~3 days. Then the winds relax off central California. This previously known synoptic scale pattern in the momentum flux from atmosphere to ocean has led to two questions: 1) How does sea-surface temperature (SST) respond on scales of 100s to 1000s of km from the coast during the two wind relaxations, and 2) What drives the SST response? Satellite microwave radiometer data indicate the wind relaxations off Oregon result in anomalously warm SSTs, up to 1°C above climatology, with a spatial extent up to 2000 km offshore. To determine whether the net surface heat flux drives the SST response, we analyzed the net surface heat flux and the latent and sensible heat flux components from the Objectively Analyzed air-sea Fluxes (OAFlux) product and the shortwave and longwave radiative flux components from the International Satellite Cloud Climatology Project (ISCCP). During the wind relaxation off Oregon, the warm SST anomaly is largely a result of anomalous heating by the net surface heat flux, specifically from a decrease in cooling from the latent heat flux due to weaker winds. The net surface heat flux accounts for up to 90% of the warm anomaly, depending on spatial location. When the winds next strengthen off of California for ~4 days, the SSTs become anomalously cold. Finally, during the ~5 days between the end of the wind reintensification and the
end of the second wind relaxation off central California, the SSTs warm offshore of California, yet a cold SST anomaly persists from the preconditioned cold water. In contrast to the Oregon relaxation, the change in SST during this second wind relaxation is not primarily driven by the net surface heat flux. The wind stress and therefore cooling from the latent heat flux are reduced during the second wind relaxation, yet the net surface heat flux anomaly is small because there is increased cloudiness and reduced solar radiation. These effects (i.e., reductions in both latent cooling and solar warming) tend to cancel each other, so the net surface heat flux only accounts for up to 25% of the change in SST, depending on spatial location. The amount of penetrating solar radiation lost through the base of the mixed-layer is small (~10% of the shortwave radiation at the surface) for both wind relaxations. Estimates of the horizontal and vertical advection and mixing terms in a mixed-layer heat budget suggest that mixed-layer shoaling is the largest contributor to the ocean surface warming during the second wind relaxation. Using satellite vector winds and Argo float data, we determined that upwelling due to wind stress curl anomalies alone cannot explain the observed mixed-layer shoaling signal. Thus, we hypothesize that reduced wind-driven vertical mixing is the primary driver of the warming trend in the SST anomaly during the central California wind relaxation. To test this hypothesis would require long time series of vertical profiles of upper ocean structure with higher temporal resolution than the Argo floats. Future studies should focus on the importance of cloudiness in this region, as well as, in other eastern boundary upwelling systems. Whether the net surface heat flux is the driver of
SST anomalies during wind relaxations appears dependent on if cloud coverage increases or decreases, thus affecting the incoming solar radiation.
Chapter 1: Introduction & Background

The mean winds off the U.S. West Coast are upwelling-favorable (Fig. 1). The definition of a wind relaxation is a calming in upwelling-favorable winds that lasts for several days (e.g. Halliwell and Allen, 1987; Kosro, 1987; Melton et al., 2009); sometimes wind relaxations reverse the wind direction, resulting in downwelling. Wind relaxation literature along the U.S. West Coast focuses on resulting changes to circulation (Gan and Allen, 2004; Fewings et al., 2015), changes to larval transport (Dudas et al., 2009; Iles et al., 2012; Morgan et al., 2012), and resulting warm poleward ocean flows within 10s of km of the coast (Melton et al., 2009; Washburn et al., 2011). However, a recent study found the spatial extent of wind relaxations at Cape Blanco, OR and Point Conception, CA reaches as far as 2000 km offshore (Fig. 2; Fewings et al., 2016). Preliminary analyses show that the large-scale changes in winds also change the sea-surface temperature (SST) (Fig. 3). Thus, the purpose of this thesis is to determine the cause of the SST response to wind relaxations along the U.S. West Coast through an analysis of regional surface heat fluxes.

1. Air-Sea Interaction

Scientists observe the complicated air-sea system through fluxes. Fluxes are the vertical exchanges of heat, momentum, and mass between the ocean and atmosphere in the coupled boundary layer. It is sometimes difficult to distinguish between forcings and response, so we only focus on the SST response to atmospheric forcings.
On short time scales of hours to days, surface forcings have an inadequate amount of time to communicate with the deep ocean below the surface mixed layer (ML), so we focus on temperature changes in the ML. The ML is the surface layer where convection and turbulent mixing cause temperature, salinity, and other properties of seawater to be almost independent of depth. The ML becomes shallower in the summertime from a decrease in convection because warm surface water rests on denser, colder waters (Holte and Talley, 2009).

Three core fluxes cause changes in mass, momentum, and heat content that immediately impact the ocean surface ML. Evaporation and precipitation (i.e. mass flux) change the ML salinity and the resulting ocean surface buoyancy. For example, if the surface water becomes less buoyant (denser) than the subsurface water, then the result is convective overturning and mixing (Cronin and Sprintall, 2001). This type of mixing (convective mixing) cools the SST and deepens the ML (entrainment). Changes in the winds blowing over the ocean surface alter the surface stress, “wind stress”, which acts as a vertical flux of horizontal momentum. Wind stress (i.e. momentum flux) impacts the ML by generating shear-driven mixing, vertical and horizontal advection of SSTs, vertical or convective mixing, and evaporative cooling. Surface heating and cooling (hereafter net surface heat flux) warms the ML by downwelling solar radiation and infrared radiation and cools by infrared radiation emitted from the surface, by latent heat loss due to evaporation, and by conductive, or sensible, heat exchange. The sensible heat flux component of the net surface heat flux results in heat loss or gain depending on the air-sea temperature difference. Generally, the net latent heat flux is out of the
ocean, but latent heat flux into the ocean occurs in our study domain. This is when the air temperature is warmer than the sea temperature and the humidity is large enough to result in condensation of water vapor. This results in fog. All of these fluxes impact the temperature of the ocean surface ML.

Changes to ML ocean temperature result from many processes, as mentioned briefly above. We assume temperature is uniform from the surface to the base of the ML, so we use SST as a proxy for ML temperature. The vertically-integrated heat budget for the ML is similar to Moisan and Niiler (1998) but with a diffusion term added:

$$\frac{\partial \text{SST}}{\partial t} = \frac{Q_{\text{net}} - \text{SWR}(-h)}{\rho_w c_p h} - \bar{u} \cdot \nabla_h \text{SST} + \kappa_h \nabla^2 \text{SST} - \left(\frac{\text{SST} - T_{-h}}{h}\right) \left(\frac{\partial h}{\partial t} + \bar{u}_{-h} \cdot \nabla_h h + w_{-h}\right) - \frac{1}{h} \nabla_h \cdot \int_{-h}^{0} \tilde{z} \ \overline{\text{SST}} \ dz.$$  

(1)

The net surface heat flux is $Q_{\text{net}}$. SWR(-h) is the shortwave radiation (SWR) penetrating past the ML base at $z = -h$, where $z = 0$ at the sea surface and $t$ is time. The horizontal velocity is $\bar{u}$ with $u$ and $v$ defined as the eastward and northward components, respectively. The horizontal gradient operator is $\nabla_h = \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y}\right)$. The temperature just below the base of the ML is $T_{-h}$. The mean density of seawater $\rho_w$ is 1025 kg m$^{-3}$, the specific heat capacity of seawater $c_p$ is 4010 J kg$^{-1}$ °C$^{-1}$, and $\kappa_h$ is the horizontal eddy diffusivity term. Tilde indicates a vertical average over the mixed layer has been removed. The covariance between $\tilde{\bar{u}}$, the deviation from the vertically-averaged horizontal velocity, and $\overline{\text{SST}}$, the deviation from the vertically-averaged temperature, is included in term F (Eq. 1). We ignore the covariance term F because we assume the ML is completely mixed in both temperature and velocity.
The terms in Eq. 1 represent, from left to right: temperature trend (A); the net surface heat flux minus SWR penetration (B); horizontal advection due to geostrophic currents, Ekman transport, and other surface currents (C); horizontal eddy diffusion (D); vertical entrainment or ML shoaling (e.g., from wind-driven vertical mixing or Ekman pumping) and the horizontal advection of spatial variations in mixed layer depth (MLD) (E).

The wind stress affects most of the terms in the heat budget (Eq. 1). For example, evaporation is a wind-driven term (Eq. 1, B). The stronger the wind stress, the greater the evaporation for a given sea-air humidity difference, and thus the colder the surface waters will become. The wind stress drives ocean currents and thus drives horizontal advection (Eq. 1, C) and diffusion (Eq. 1, D) of SSTs. Wind stress also generates shear-driven mixing and results in entrainment of deeper, colder water to the surface ML (Eq.1, E). Curl in the wind stress results in horizontal convergence or divergence (Cronin and Sprintall, 2001) in the surface waters. Divergence of surface currents produce upwelled cold water from beneath the ML, labeled as a positive Ekman pumping velocity. Therefore, changes in the wind stress lead to hypothesized changes in the SST.

We will use a time-integrated version of the heat budget (Eq. 1) to analyze the resulting changes in SSTs:

\[ SST(t) = SST(t_0) + \int_{t_0}^{t} [B + C + D + E] \, dt \]  

(2)
1.1. Coastal Wind Relaxations

Upwelling occurs globally along eastern boundary currents. In our study domain, the eastern boundary current system is known as the California Current System (CCS). The CCS includes Washington to southern California and transports water equatorward (Hickey, 1979; Lynn and Simpson 1987). The upwelled water in the CCS fuels some of the highest productivity on Earth. For example, eastern boundary current systems only represent approximately 1% of the global ocean surface, but these regions yield more than 20% of all wild-marine-captured fish globally and are essential ecosystems for marine biodiversity (Dudas et al., 2009; Iles et al., 2012; Morgan et al., 2012).

Relaxations in upwelling-favorable winds occur year round at many locations around the world. Some locations include: the Canary Current upwelling system (García-Lafuente et al., 2006); the Benguela Current System at Cape Columbine, South Africa (Fawcett et al., 2008), and the CCS (Melton et al., 2009; Washburn et al., 2011). Along the U.S. West Coast, upwelling-favorable winds are dominant in the summer months (Fig. 1), yet summertime wind relaxations occur approximately every 2 weeks (Melton et al., 2009; Washburn et al., 2011). This study focuses on wind relaxations occurring off the coast of two locations: Cape Blanco, Oregon (hereafter, Northern Wind Relaxations) and Point Conception, California (hereafter, Southern Wind Relaxations) (Fig. 1).

Scientists typically use in situ measurements to identify the arrival of coastal wind relaxations and the ocean response (e.g., Send et al., 1987; Melton et al., 2009; Washburn et al., 2011). Melton et al. (2009) used zero crossings of the first empirical orthogonal function (EOF) of along-coast wind velocities from coastal buoys to index
wind relaxations at Point Conception, CA (hereafter the Melton index). The central California coastal wind relaxations result in warm poleward buoyant flows within ~20 km of the coast (Melton et al., 2009; Washburn et al., 2011). Melton et al. (2009) and Washburn et al. (2011) quantified the propagation speeds, velocity structure, and cross-shore spatial scales of the warm buoyant plumes as they moved poleward along the coast. The warm waters originate from the lee of the coastal headland of Point Conception, CA (Melton et al., 2009; Washburn et al., 2011). This periodic warm poleward flow appears to impact larval distribution and biodiversity along the coast (Joubert et al., in prep).

Low marine stratus clouds generally form during the Southern Wind Relaxations at Point Conception, CA (Kosro 1987; Rahn and Parish, 2008). The low marine clouds prevent solar heating, resulting in colder than normal SSTs. Also, cold SSTs can lead to stronger than normal atmospheric temperature inversion, producing low marine clouds. (Wang et al., 2004). Wind relaxations are important for marine weather forecasts due to these effects on cloudiness (Kosro 1987; Rahn and Parish, 2008).

1.2. Atmospheric Link Between Northern and Southern Wind Relaxations

The Northern Wind Relaxations (Halliwell and Allen, 1987; Bond et al., 1996; Nuss et al., 2007) and the Southern Wind Relaxations (Melton et al., 2009; Washburn et al., 2011) have been extensively studied. Using satellite wind observations over the ocean, the two regional wind relaxations have recently been connected through a ~12 day ‘event cycle’ (Fewings et al., 2016). The Southern Wind Relaxations are linked to
the Northern Wind Relaxations through atmospheric synoptic patterns in atmospheric sea level pressure, 500-hPa heights, and wind stress. Figure 2 shows the composite wind stress anomalies from 6 days before the Southern Wind Relaxations to 5 days after, capturing the entire ~12 day cycle.

Halliwell and Allen (1987) coined the term ‘event cycle’ to refer to the events in which the Northern Wind Relaxation (Fig. 2, blue) occurs from day -6 to -2 and then the winds re-intensify (Fig. 2, red) from day -1 to 2. Recently, the event cycle was amended to include a third step (Fewings et al., 2016). The Southern Wind Relaxations (Fig. 2, blue) occur from day 0 to 5, after the winds re-intensify. The complete ~12 day event cycle (Fewings et al., 2016) starts with an eastward-moving 500-hPa trough causing a Northern Wind Relaxation (Halliwell and Allen, 1987). Next, a northeastern extension of the North Pacific High atmospheric pressure system (NPH) intensifies the upwelling-favorable winds first near Point Conception and extending poleward off the coast of Oregon. The Southern Wind Relaxation occurs as a low sea-level pressure anomaly follows the NPH (Fewings et al., 2016). The spatial extent of the wind relaxations is much larger than just along the coast, with an extent of ~2000 km offshore (Fig. 2). The Northern Wind Relaxations are driven by an extra-tropical cyclone or weak trough and a southward shift of the jet stream (Halliwell and Allen, 1987; Bane et al., 2005, 2007); in contrast, the low-pressure system associated with the Southern Wind Relaxations occurs from the warm desert air advected by the NPH (Fewings et. al., 2016) similarly to the rarer wind reversals (Nuss, 2007).
This newly defined ~12 day event cycle indicates that the two identified wind relaxations in the CCS extend much farther, and are more related, than initially thought. That study led to the question: how does the ocean respond to this atmospheric synoptic ‘event cycle’? Specifically, how do regional wind relaxations change SSTs on daily time scales?

1.3. Regional SST Anomalies During 12-Day Event Cycle

SSTs during Southern Wind Relaxations have been examined using data from the satellite-borne Advanced Microwave Scanning Radiometer – Earth Observing System (AMSR-E). The AMSR-E polar-orbiting microwave radiometer, which was available from 2002—2011, is advantageous in studying the CCS in summer because it penetrates through clouds. The striking SST anomalies during Southern Wind Relaxations range from +/- 1°C (Fig. 3). There is a warm anomaly, extending from ~250 to 1000 km offshore of Northern California, during Northern Wind Relaxations (Fig. 3, day -6 to -1), which we hypothesize to occur due to the decreasing wind stress, which would result in a decrease of wind-driven evaporative cooling. A cold anomaly is present offshore of Southern California, extending ~2000 km southwest from the coast, when the upwelling-favorable winds re-intensify (Fig. 2, day -2 to 2). The cold anomaly may be a result of increasing wind-driven vertical mixing, wind-driven evaporative cooling, and upwelling from Ekman pumping (Eq. 2). During the Southern Wind Relaxations (Fig. 3, day 0 to 5) there is also a cold SST anomaly with the same spatial description as the previously mentioned cold anomaly. This is surprising because we would expect a warm
SST anomaly comparable to the Northern Wind Relaxations warm SST anomaly. To investigate these SST anomalies, we use the surface heat fluxes and Eq. 2.

1.4. Surface Heat Fluxes

The net surface heat flux is an important driver of SSTs in the ocean surface ML. In this thesis the sign convention for surface heat flux components is that the downward heat flux from the atmosphere to sea is positive, indicating ocean warming. The net heat flux \( Q_{\text{net}} \) is expressed as:

\[
Q_{\text{net}} = Q_{\text{SWR}} + Q_{\text{LWR}} + Q_{\text{SHF}} + Q_{\text{LHF}}
\]

where \( Q_{\text{net}} \) equals net shortwave (\( Q_{\text{SWR}} \)) and longwave (\( Q_{\text{LWR}} \)) radiative fluxes plus net sensible (\( Q_{\text{SHF}} \)) and latent (\( Q_{\text{LHF}} \)) heat fluxes. The net surface heat flux and components have units W m\(^{-2}\).

The shortwave radiation (SWR) is solar heating, which warms the ocean. SWR entering the Earth’s atmosphere is absorbed, scattered, and reflected by water in both its liquid and vapor forms (Zhang et al., 2004). The equation for SWR is (Yu et al., 2007)

\[
Q_{\text{SWR}} = SWR \downarrow - \alpha \ SWR \downarrow
\]

where SWR\( \downarrow \) is the downward SWR into the ocean and \( \alpha \) is a varying albedo based on solar elevation angle and cloud cover. Thus, cloudiness and cloud structure are important. When there is cloudiness, SWR is reflected from the cloud tops back into the atmosphere. SWR entering the ocean’s surface is distributed in the water column and some may penetrate through the bottom of the mixed layer. The penetrating solar
radiation is dependent on the mixed layer depth (MLD) and absorption characteristics of the water.

The sun emits longwave radiation (LWR), but Earth’s surface also emits longwave LWR as a near black-body. Clouds and atmospheric gases absorb longwave radiation and reemit a potion of the energy to outer space and a portion back toward the Earth’s surface. Therefore, net LWR is the sum of the outgoing and incoming LWR. The amount of LWR reflected back to the surface is dependent on cloud thickness, height, and the amount absorbed by water vapor content. The net LWR, $Q_{LWR}$, can be split into downward LWR ($LWR_\downarrow$, into the ocean) and upward LWR ($LWR_\uparrow$, out of the ocean) where

$$Q_{LWR} = \varepsilon LWR_\downarrow - \varepsilon \sigma T_s^4$$

(5)

and $LWR_\uparrow$ is defined as

$$LWR_\uparrow = \left[ \varepsilon \sigma T_s^4 + (1 - \varepsilon)LWR_\downarrow \right].$$

(6)

$LWR_\uparrow$ is the portion of $Q_{LWR}$ emitted from the ocean surface, which acts as a near black-body, plus the equivalent of the reflected part of the incoming radiation $LWR_\downarrow$ (e.g., Rapp, 2014; Yu, 2007). The Stefan-Boltzmann constant is $\sigma = 5.670 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$ and the infrared emissivity is $\varepsilon = 0.97$. Emissivity is the relative ability of a surface to absorb or emit heat, where a true black body would have an emissivity of 1.

The latent heat flux (LHF) is ocean surface cooling through evaporation and is driven by the wind and the sea-air humidity difference. Relative humidity of the air is usually less than 100%, as compared to the air-sea interface where the vapor pressure is saturated. Thus, except in foggy conditions, water vapor generally leaves the ocean
surface and cools the ocean. This evaporated moisture can condense into clouds, releasing heat to the atmosphere and affecting the large-scale wind patterns. The sensible heat flux (SHF) is the exchange of heat via conduction. This occurs when the ocean and atmosphere have different temperatures. As with the LHF, it increases with wind stress, but it is typically small in comparison to the SWR and LHF.

2. Research Question

This thesis aims to discover what drives the SST anomalies (Section 1.3) during the linked Northern and Southern U.S. West Coast wind relaxations (Fig. 3). This research expands on previous studies of poleward flows during wind relaxations by examining broader-scale SST responses to wind stress and atmospheric pressure anomalies. We focus on the summer months of Jun–Aug and the years 2002–2009. The study domain remains in the same region as Fewings et al. (2016): 140°W to 110°W longitude and 25°N to 50°N latitude, including the CCS (Fig. 2).

In this thesis, I analyze the SST response to wind relaxations by using the time-integrated ML heat budget (Eq. 2). The two main questions are as follows:

1. Is the \( Q_{\text{net}} \) the main driver of the SST anomalies during the 12-day event cycle?

2. Which components are the main contributors to the \( Q_{\text{net}} \) during the 12-day event cycle?

This thesis will flow as follows. The second chapter describes four datasets: Quick Scatterometer (QuikSCAT), Argo floats, Objectively Analyzed air-sea Fluxes (OAFlux; Yu et al., 2008) and International Satellite Cloud Climatology Project (ISCCP;
Zhang et al., 2004). I will also discuss wind stress curl and MLD calculations, and how I constructed composite anomalies for the ~12-day event cycle. Chapter 3 investigates the evolution of the regional composite anomalies for each term during the wind relaxations and the relative sizes of the heat flux components. In Chapter 4, I will discuss the main drivers of the SST anomalies during the Northern and Southern Wind Relaxations. I will summarize this thesis in the fifth chapter.
Chapter 2: Data and Methods

This study investigates changes to ML heating from changes in the wind stress and resulting changes to atmosphere-ocean heat exchange over the open ocean during wind relaxations in the CCS. We use the climatological MLD from Holte et al. (2010), which uses raw Argo float data and a hybrid method for MLD calculations (Holte and Talley, 2009). The QuikSCAT satellite measured changes in the wind stress. OAFlux + ISCCP determine the resulting changes in SST and atmosphere-ocean heat exchange.

1. Wind Stress Curl and Wind-Stress-Curl-Driven Upwelling From Satellite Data

Vector wind stress data is available from QuikSCAT during the study period Jun–Aug 2002–2009. QuikSCAT obtained global swath measurements of wind velocity by transmitting microwave pulses and measuring the power backscattered from the ocean surface (e.g., Tang et al., 2004). QuikSCAT Version 3 Level 2B data produced by NASA’s Jet propulsion Laboratory (Fore et al., 2014) provides 10-m equivalent neutral wind velocity. Equivalent neutral wind is the mean wind that would have been observed if there was neutral atmospheric stratification. The wind velocity vectors from QuikSCAT provide enough data to estimate the wind stress. We estimate the wind stress using the Coupled Ocean-Atmosphere Response Experiment (COARE) version 3.5 neutral drag parameterization (Edson et al., 2013). We follow the same methods as Fewings et al. (2016) modified for Jun–Aug 2002–2009. The wind stress at the ocean surface is (Large and Pond, 1980):

\[ |\vec{\tau}| \equiv \tau = -\rho_a \bar{u} \bar{w} = \rho_a C_d U^2. \]  

(7)
The air density is \( \rho_a \approx 1.225 \text{ kg m}^{-3} \), \( C_D \) is the neutral 10-m drag coefficient determined by COARE 3.5, \( U \) is wind speed, and \( \bar{\tau} = \bar{\tau}_x + \bar{\tau}_y \). The flux computed using the direct covariance method (Edson et al., 2013) is \( \rho_a \bar{u}'\bar{w}' \), where \( u' \) and \( w' \) are the fluctuating along-wind and vertical velocities, respectively, and the overbar represents a time average. The wind stress components \( \tau_x \) and \( \tau_y \) were calculated from the stress magnitude and wind direction. The resulting components are rotated into a coordinate system where \( x \) is the mean summer wind direction and then interpolated onto a 0.1° grid.

Wind stress curl can be estimated using

\[
\nabla \times \vec{\tau} = \frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}.
\]

(8)

Wind stress curl is in units of N m\(^{-3}\). The wind-stress-curl-driven upwelling (i.e. Ekman pumping velocity) \( w_{ek} \) is proportional to wind stress curl (Kraus and Businger 1994):

\[
\quad w_{ek} = \frac{1}{\rho_w f} \nabla \times \vec{\tau},
\]

(9)

where \( \rho_w \) is the mean density of seawater and \( f \) is the Coriolis parameter

\[
\quad f = 2\Omega \sin(\Phi),
\]

(10)

where two times the angular momentum of the earth (2\( \Omega \)) is 1.458x10\(^{-4} \) s\(^{-1} \) and \( \Phi \) is latitude. We converted the wind stress curl into an Ekman pumping velocity (Eq. 9) with units of m day\(^{-1} \).

2. Mixed-Layer Depth

Argo floats (http://www.argo.net) are quasi-Lagrangian, vertically-profiling instruments that sample the open ocean at a rate of approximately one observation
every 10 days for a ~3° latitude-longitude area (http://www.argo.ucsd.edu). This time frame allows for analysis of processes with time scales of months and longer, but not days. There are currently more than 3000 floats supplying subsurface conductivity, temperature, and depth (CTD) data. The temperature and salinity accuracy are 0.005°C and 0.01 psu, respectively. The floats descend to ~2000 m and drift with the current for 10 days before returning to the surface (http://www.argo.net) with descent and ascent speeds of ~10 cm s⁻¹. Measurements occur on the ascent with a vertical resolution of ~10 m (~200 measurements) and stop ~5 m below the surface (http://www.argo.ucsd.edu).

The Argo floats provide enough data to compute the mixed layer depth (MLD). We used an existing summer climatology of MLD from Jun–Aug 2000–2015 (Fig. 4) (http://mixedlayer.ucsd.edu). The MLD is calculated with a hybrid method (Holte and Talley, 2009). The hybrid method uses both density changes over the MLD and vertical gradients of density to determine a suite of MLDs. The method chooses a final MLD based on patterns in the suite (details in Holte and Talley, 2009). This study focuses on the depth of actual mixing of water parcels, so we use the density algorithm MLD, rather than the alternative temperature-based version.

3. Meteorological Data and Atmosphere-Ocean Heat Flux From Blended Data

We use OAFlux+ISCCP for the net surface (atmosphere-ocean) heat flux (Qₙₑₙ) analysis. OAFlux is a global flux product produced by an optimal blending of multi-platform satellite retrievals with three numerical weather prediction (NWP) reanalyses
from NCEP1, NCEP2, and ERA-40 (Yu et al., 2008). Buoy data determines the weight of each product. Daily SHF and LHF are available from 1985–present on a 1° latitude-longitude grid. OAFlux does not objectively analyze SWR and LWR, but provides access to the ISCCP SWR and LWR data from 1983–2009. A complete radiative transfer model from the Goddard Institute for Space Studies (GISS) General Circulation Model (GCM) computes ISCCP radiative fluxes. Global ISCCP fluxes for clear and cloudy skies are available every 3 h on a 2.5° latitude-longitude grid. For consistency, OAFlux provides daily ISCCP fluxes linearly interpolated onto a 1° latitude-longitude grid. OAFlux + ISCCP provides Qnet, SHF, LHF, SWR, LWR, and daily estimates of surface meteorological variables (wind speed, specific humidity, air temperature and SST) from 1985–2009. Further detail on OAFlux + ISCCP can be found in the Appendix. This study focuses on daily OAFlux + ISCCP for the summer months of Jun–Aug 2002–2009.

4. Mixed-Layer Heat Budget

To determine how much of the SST anomalies are a result of the Qnet anomalies, we simplify Equation 2 to test a 1-D heat budget. First, we rewrite the equation as:

\[
\int_{t_0}^{t} \frac{\partial SST}{\partial t} dt - \int_{t_0}^{t} \frac{Q_{net}}{\rho_w c_p h} dt = \int_{t_0}^{t} \left[ \frac{-SWR(-h)}{\rho_w c_p h} - \vec{u} \cdot \nabla h_{SST} + \kappa_h \nabla^2 SST - \frac{(SST - T_h)}{h} \left( \frac{\partial h}{\partial t} + \vec{u} \cdot \nabla h + w_h \right) \right] dt. \tag{11}
\]

Then, we simplify it further to:

\[
\int_{t_0}^{t} \frac{\partial SST}{\partial t} dt - \int_{t_0}^{t} \frac{Q_{net}}{\rho_w c_p h} dt = \int_{t_0}^{t} (\text{residual}) dt \tag{12}
\]

where SST is the composited SST anomaly at day t. The reference time t0 is day -6 for the Northern Wind Relaxation and day 0 for the Southern Wind Relaxation. This allows
for a clear picture of the changes in the SST over a 6 day period for each relaxation instead of a 12 day period for the entire event cycle. The MLD $h$ is held constant. The summer climatological spatial mean value of $h$ is smaller poleward of 37°N latitude ($h = 19.5$ m) and larger equatorward of 37°N latitude ($h = 25$ m) (Fig. 4). We use $h = 19.5$ m for the Northern Wind Relaxations and $h = 25$ m for the Southern Wind Relaxations. We will refer to the terms on the left-hand side of Equation 12 as $\Delta SST_{obs}$ and $\Delta SST_{Q_{net}}$, respectively, and the right-hand side of the equation as $\Delta SST_{Res}$. We estimate $\Delta SST_{obs}$, the observed change in SST anomalies using OAFlux SSTs. $\Delta SST_{Q_{net}}$ is the estimated change in SST anomalies if $Q_{net}$ was the sole driver of ML temperature. We estimate $\Delta SST_{Q_{net}}$ using OAFlux + ISCCP. $\Delta SST_{Res}$ is the estimated change in SST anomalies from the remaining heat budget terms (C,D,E in Eq. 2) and will equal 0 if $Q_{net}$ is the sole driver.

5. Constructing Composite Anomalies

The wind relaxations are identified using the Melton index, yet following previous studies we reject ~15% of the events because they are associated with landfalling fronts, which are not the focus of this study (see Fewings et al., 2016, for details). There are 44 remaining wind relaxations at Point Conception, CA during the summer months of Jun–Aug 2002–2009. We constructed composited anomalies for the following variables: wind stress (N m$^{-2}$ or Pa); $Q_{net}$, LHF, SHF, SWR, LWR (W m$^{-2}$); SST (°C), $\Delta SST_{obs}$ and $\Delta SST_{Q_{net}}$ (°C) from the 1-D heat budget equation (Eq. 12); wind stress
curl (N m$^{-3}$); and wind-stress-curl-driven upwelling (m day$^{-1}$). For the heat flux terms, positive values show anomalous ocean warming.

The method for calculating anomalies for OAFlux + ISCCP differs from the methods used for the wind stress components. The OAFlux + ISCCP data are 1° gridded daily means with no missing values. We calculated the anomalies at each latitude/longitude location by removing an annual cycle using a high-pass filter with a 50-day half-amplitude cutoff (Fig. 5, red line). In contrast, the wind stress swaths are on a 0.1° grid and have regions of missing data. The anomalies for the wind stress components were computed by removing the summer mean over all years from each latitude/longitude location.

In composite anomaly figures, gray shading indicates anomalies that are not significant at the 95% confidence level. We calculated the 95% confidence level of each anomaly using:

$$ X \pm t \left( \frac{S}{\sqrt{n}} \right) $$

where $X$ is the mean of the anomaly over all events, $t$ is the inverse of Student’s T cumulative distribution, $S$ is the standard deviation of the anomaly and $n$ is the sample size (equal to the number of events with data at each grid point). The inverse cumulative distribution calculation produces a t-value based on the sample size $n$ and the desired 95% confidence interval.
Chapter 3: Results

The results are split into two sections: Northern Wind Relaxation and Southern Wind Relaxation. For the purpose of this study the Northern Wind Relaxation is 1 to 6 days before the start of the Southern Wind Relaxation. The Southern Wind Relaxation begins with the arrival of the wind relaxation at Point Conception, CA (day 0) and ends on average 5 days later.

To answer the first thesis question: Is the $Q_{\text{net}}$ the main driver of the SST anomalies during the 12-day event cycle? (Ch. 1, Section 2), I mapped the evolution of composite anomalies of the $\Delta SST_{\text{obs}}$ and $\Delta SST_{Q_{\text{net}}}$ from the 1-D ML heat budget (Eq. 12, Ch. 2), and calculated $\Delta SST_{Res}$. To answer the second question: Which components are the main contributors to the $Q_{\text{net}}$ during the 12-day event cycle? (Ch. 1, Section 2), I mapped the daily evolution of the composited net surface heat flux $Q_{\text{net}}$ and its components: LHF, SHF, SWR, LWR. Next, I mapped the daily evolution of the composite anomalies of $Q_{\text{net}}$ and its terms to determine which components are the main contributors to $Q_{\text{net}}$. Last, I take a closer look at the net LWR, $Q_{LWR}$, by splitting it into its components, LWR downward flux and LWR upward flux. These steps are repeated for the Northern and Southern Wind Relaxations.

1. Northern Wind Relaxation

Consistent with the QuikSCAT (Fig. 2) and AMSR-E (Fig. 3) figures, the Northern Wind Relaxation is identified by anomalously weak wind stress (Fig. 6a, blue). The positive SST anomalies (Fig. 6b) visually appear correlated with changes in the wind
stress, perhaps with a time lag (Fig. 6a). The warm SST anomaly (Fig. 6b, red) is spatially located in the same region as where the winds relax (Fig. 6a, blue) and a cold SST anomaly (Fig. 6b, blue) occurs when the wind stress later intensifies (Fig. 6a, red).

1.1. Mixed-Layer Heating

I use a 1-D ML heat budget (Eq. 12) to determine whether the main driver of the SST anomalies during the Northern Wind Relaxations is the surface heat flux $Q_{net}$. The left-hand side of the ML heat budget is $\Delta SST_{obs} - \Delta SST_{Q_{net}}$, and the right-hand side is $\Delta SST_{Res}$ (Fig. 7). Over 6 days, both $\Delta SST_{obs}$ and $\Delta SST_{Q_{net}}$ have positive anomalies. The $\Delta SST_{Res}$ surrounding the warm SST anomaly is small (~10% of $\Delta SST_{obs}$), thus the $Q_{net}$ drives up to ~90% percent of the warm SST anomaly and ~45% of the strongest warm SST anomaly (i.e. the red area at left in Fig 7a). We conclude that vertical mixing and advection are not necessary to explain the majority of the SST anomaly during the Northern Wind Relaxations, but these terms are needed to explain the strong SST anomaly.

1.2. Net Surface Heat Flux and Components

We address the second question (Ch. 1, section 2) by using mapped composites of the 44 wind relaxations for $Q_{net}$, LHF, SHF, SWR, and LWR during the Northern Wind Relaxations. The results identify the magnitude and evolution over time of each component. All composited components (Fig. 9) look similar to the summer means (Fig.
Therefore, we look at the anomalies because the composites are dominated by the mean.

The composite anomalies create a clearer picture of the changes during wind relaxations (Fig. 10). $Q_{\text{net}}$ from day -6 to day -1 shows 3 anomalies: 1) a slight negative anomaly (less than average ocean heating, blue) off Oregon, 2) a negative anomaly off the coast of California that appears from day -2 to -1, and 3) a prominent positive anomaly (ocean warming, red) that appears from day -5 to -1 in a similar region to the relaxation in the wind stress (Fig. 7a, blue). The negative $Q_{\text{net}}$ anomaly to the north in Fig. 10a from day -6 to -3 is a result of the decrease in shortwave radiation, presumably from cloudiness (Fig. 10e, blue).

The positive $Q_{\text{net}}$ (Fig. 10a) anomaly associated with the Northern Wind Relaxation (in the center of the panels from day -6 to -1) is a result of an increase in SWR (Fig. 10e, red), a decrease in latent cooling (Fig. 10b, red), and a small increase in the downward SHF (Fig. 10c, red). The SST anomaly signal indicates ocean warming (Fig. 7b, red) where the $Q_{\text{net}}$ anomaly is heating the ocean (Fig. 10a, red). This result matches the finding in the ML heat budget (Eq. 12) mentioned above (Section 1.1) that changes in the $Q_{\text{net}}$ are the main drivers of changes in the SST during the Northern Wind Relaxation.

Calculations of LWR↑ and LWR↓ indicate LWR↓ anomalies have a larger magnitude than LWR↑ (Fig. 11). The net LWR anomaly is driven by changes in the incoming LWR (Eq. 5). LWR↑ is dependent on SST (Eq. 6), and thus mimics the SST anomalies (Fig. 11c, 11d).
2. Southern Wind Relaxation

The Southern Wind Relaxation also displays anomalously weak wind stress (Fig. 12a) as in the QuikSCAT (Fig. 2) and AMSR-E (Fig. 3) figures (Ch. 1). But, in contrast to the Northern Wind Relaxation, the Southern Wind Relaxation SST anomalies (Fig. 12b) do not agree spatially with changes in the wind stress (Fig. 12a). Instead, a cold SST anomaly (Fig. 12b, blue) is located in the same region as where the winds relax (Fig. 12a, blue) and a warm SST anomaly (Fig. 12b, red) occurs when the wind stress intensifies (Fig. 12a, red). The wind stress during the Southern Wind Relaxation is presumably coupled with the resulting SST through changes in atmospheric stratification and atmospheric boundary layer response (Chelton et al., 2007).

2.1. Mixed-Layer Heating

Again (as in Section 1.1), we use a 1-D ML heat budget (Eq. 12) to determine whether the main driver of the SST anomalies during the Southern Wind Relaxations is $Q_{\text{net}}$. Over 6 days, both the $\Delta SST_{\text{Obs}}$ and $\Delta SST_{Q_{\text{net}}}$ have warm SST signals in the region south of Pt. Conception (Fig. 12c,d). The $\Delta SST_{Q_{\text{net}}}$ is smaller during the Southern Wind Relaxation than the Northern Wind Relaxation. In contrast to the Northern Wind Relaxation, $\Delta SST_{Q_{\text{net}}}$ drives a small percentage (< 25%) of the warm $\Delta SST_{\text{Obs}}$ (Fig. 13). We must estimate terms in $\Delta SST_{\text{Res}}$ to explain what is driving the SST anomalies during the Southern Wind Relaxations (Ch. 4, Section 2).
2.2. Net Surface Heat Flux and Components

We now address the second question (Ch. 1, Section 2) for the Southern Wind Relaxations. We discuss the magnitude and evolution over time of each component of the net surface heat flux.

Similarly to above (Section 1.2), we will examine the composited components of the net surface heat flux during the Southern Wind Relaxation. We look at the anomalies because the composites are dominated by the mean, as is the case in the Northern Wind Relaxation. Yet, there is a striking signal at the bottom of the LWR panels from day 0 to 5. This signal indicates a decrease in the upward flux of net LWR and therefore an increase in the downward flux of LWR compared to the summer means. This is associated with the wind relaxation at Point Conception, CA. An increase in downward LWR is typically due to cloudiness (Ch. 1, Section 1.1) (Iacobellis and Cayan, 2013).

Q_{net} from day 0 to 5 shows 3 anomalies: 1) a positive anomaly offshore of Oregon (Fig. 15a, red) from day 0 to 2, 2) a negative anomaly (less than average ocean heating) off the coast of California, and 3) a slight positive anomaly (ocean warming) west of Baja California (Fig. 15a, red) from day 1 to 3. The positive Q_{net} anomaly (Fig. 15a) to the north from day 0 to day 3 is a result of an increase in SWR (Fig. 15e, red) and a decrease in latent cooling (Fig. 15b, red). This anomaly is associated with the end of the Northern Wind Relaxation and was discussed in Section 1. The slight positive Q_{net} anomaly (Fig. 15a, red) west of Baja California is a result of a decrease in latent cooling (Fig. 15b, red) from a decrease in wind velocity due to the Southern Wind Relaxations.
(Fig. 12a, blue). We will not focus on this anomaly because it extends farther south than our study domain.

The $Q_{\text{net}}$ anomaly off the coast of California is negative (Fig. 15a, blue) due to an increase in latent cooling (Fig. 15b, blue) and a decrease in SWR (Fig. 15e, blue), presumably due to cloudiness based on the positive LWR anomalies (Fig. 15d, red) observed in the same spatial region. However, the $Q_{\text{net}}$ anomaly is not strong, because the cloudiness is preventing a positive anomaly from the SWR, in contrast to the Northern Wind Relaxation (Section 1.2). The LHF and SWR anomalies oppose each other and lead to a small $Q_{\text{net}}$ anomaly. As discussed above (Section 2.1) this affects the 1-D ML heat budget equation and suggests that the SST anomalies are not solely driven by the changes in $Q_{\text{net}}$. The LWR anomalies for the Southern Wind Relaxations (Fig. 16) are similar in amplitude to those in the Northern Wind Relaxations (Fig. 11).

In contrast to the results during the Northern Wind Relaxation (Section 1.1), the SST anomalies during the Southern Wind Relaxations are not driven mainly by $Q_{\text{net}}$, and thus cannot be understood by using the time-integrated 1-D heat budget (Eq. 12). Instead, the $Q_{\text{net}}$ anomaly is small due to a cancellation between the shortwave radiation and latent heat flux anomalies. The wind stress and therefore cooling from the latent heat flux are reduced during the Southern Wind Relaxation, yet the net surface heat flux anomaly is small because there is increased cloudiness. So, the $Q_{\text{net}}$ drives $< 25\%$ of $\Delta SST_{\text{obs}}$, depending on spatial location.
Chapter 4: Discussion

The answer to question 1 (Ch.1 Section 2) differs from the Northern Wind Relaxations to the Southern Wind Relaxations, thus we split the discussion into two sections.

1. What causes the SST anomalies during the Northern Wind Relaxation?

The warm SST anomalies during the Northern Wind Relaxation result mostly (~90%) from changes in the net surface heat flux components, specifically the SWR and LHF. Thus, we do not need the residual to qualitatively explain the SST anomaly, and the 1-D heat budget (Eq. 2) is appropriate. Yet, where the warm SST anomaly is largest, Eq. 2 accounts for less (~45%) of the anomaly. In that region, the other terms in the heat budget equation (Eq. 2) must be more important, similarly to what is discussed below for the Southern Wind Relaxation.

2. What causes the SST anomalies during the Southern Wind Relaxation?

The results from Chapter 3 indicate the driver of the SST anomalies during the Southern Wind Relaxation is not $Q_{\text{net}}$. This led to the question: What terms in the time-integrated 1-D heat budget residual are driving the SST anomalies? Below, we estimate the remaining terms in Equation 2: penetrating solar radiation, horizontal advection, horizontal diffusion, and ML deepening and the Ekman pumping velocity.
2.1. Solar Penetration

We compute the downward radiative heat flux remaining below the ocean surface using an empirical formula (Paulson and Simpson, 1977):

\[
SWR(-h) = SWR_0 \left( Re^{-\frac{h}{\zeta_1}} + (1 - R)e^{-\frac{h}{\zeta_2}} \right),
\]

where \( SWR(-h) \) is the radiative flux anomaly at MLD \( h \), \( SWR_0 \) is the shortwave radiative flux anomaly at the sea surface. \( R = 0.62 \) and the attenuation lengths are \( \zeta_1 = 1.5 \) m and \( \zeta_2 = 20 \) m. The first term on the right-hand side is the longwave (red) spectral components that are mostly absorbed within the first few meters, and the second term is the visible (blue-green) spectral components. For the Southern Wind Relaxation, we use a MLD of 25 m (Section 2.2). From the time-integrated ML heat budget (Eq. 2, term B), the solar penetration equation becomes:

\[
-\frac{1}{\rho_w c_p h} \int_{t_0}^{t} SWR(-h) \, dt = -\frac{1}{\rho_w c_p h} \int_{t_0}^{t} SWR_0 (0.62 \ast e^{-25 \text{ m}} + 0.38 \ast e^{-25 \text{ m}}),
\]

This simplifies to give a penetrating radiative flux anomaly through the base of the ML of \( \sim 0.11 \ast SWR_0 \), or \( \sim 11\% \) of the composited SWR anomaly computed from OAFlux.

Therefore, we conclude that the solar penetration term can be neglected.

2.2. Horizontal Advection of SST Gradients

The time-integrated horizontal advection of SST gradients is calculated using Eq. 2, term C:

\[
- \int_{t_0}^{t} \bar{u} \cdot \nabla \mu \, SST \, dt
\]
An appropriate value for the horizontal advection is \( |u| \sim 0.1 \, \text{m s}^{-1} \). The SST changes by \( \sim 10^\circ\text{C} \) over a distance of \( \sim 2000 \, \text{km} \), based on the mean SST gradient (not shown here).

With a time change of 6 days, we write:

\[
\frac{10^\circ\text{C}}{2000 \, \text{km}} \times \frac{1 \, \text{km}}{1000 \, \text{m}} \times \frac{0.1 \, \text{m}}{1 \, \text{s}} \times \frac{86400 \, \text{s}}{1 \, \text{day}} \times 6 \, \text{days} \cong 0.26 \, \text{\circ C} \quad (17)
\]

The change in SST due to horizontal advection of SST gradients is at most \( \sim 25\% \) of \( \Delta SST_{\text{obs}} \) and is expected to make a negligible contribution to the SST anomalies during the Southern Wind Relaxation.

### 2.3. Eddy Diffusivity (Horizontal Heat Diffusion)

The time-integrated horizontal heat diffusion is calculated using Eq. 2, term D:

\[
\int_{t_0}^{t} \kappa_H \nabla^2 \, \text{SST} \, dt. \quad (18)
\]

A typical value for the eddy diffusivity \( \kappa_H \) is 500 m² s⁻¹ (Dong et al., 2007). As in Section 2.2, the SST changes by \( \sim 10^\circ\text{C} \) over a distance of \( \sim 2000 \, \text{km} \). With a time change of 6 days, we write:

\[
\frac{500 \, \text{m}^2}{1 \, \text{s}} \times \frac{10 \, ^\circ\text{C}}{(2000 \, \text{km})^2} \times \left( \frac{1 \, \text{km}}{1000 \, \text{m}} \right)^2 \times \frac{86400 \, \text{s}}{1 \, \text{day}} \times 6 \, \text{days} \cong 6 \times 10^{-4} \, ^\circ\text{C} \quad (19)
\]

Thus, the horizontal diffusion of heat is negligible.

### 2.4. Entrainment and Detrainment

The time-integrated entrainment is estimated using Eq. 2, term F:

\[
\int_{t_0}^{t} \frac{(\text{SST} - T_{-h})}{h} \left( \frac{\partial h}{\partial t} + \vec{u} \cdot \nabla h - w_{-h} \right) \, dt \quad (20)
\]
We assume the MLD $h$ is constant in space. First, we look at the ML
depening/shoaling term. The largest change in $h$ in one day we estimated as $\sim 20$ m based on variability in the Argo data (not shown). We use the summer mean maximum MLD $h = 50$ m (Fig. 4). With a time change of 6 days, we write:

$$\frac{(SST - T_{-h})}{50 \, m} \left( \frac{20 \, m}{1 \, day} \times 6 \, days \right) \approx 2.4 \ast (SST - T_{-h})$$  \hspace{1cm} (21)

For the ML shoaling term to be considered important, the temperature change from the base of the ML (SST) to right below the base of the ML must be $\geq 0.15^\circ$C. Then the change in SST from ML shoaling ($\approx 0.35^\circ$C) would account for $\sim 50\%$ of $\Delta SST_{obs}$ ($\sim 0.7$ $^\circ$C). This estimate of the vertical advection and mixing terms in the ML heat budget suggests that ML shoaling, presumably due to reduced vertical mixing, is a large contributor to changes in the SST during wind relaxations.

Though we do not have estimates of the total vertical velocity at the base of the ML, we can estimate the part of the entrainment term in Eq. 21 that is due to the Ekman pumping velocity. We use the time-integrated wind-stress-curl driven upwelling equation (Eq. 9):

$$\int_{t_0}^{t} w_{ek} = \int_{t_0}^{t} \frac{1}{\rho_w f} \nabla \times \tau, \hspace{1cm} (22)$$

where a positive wind stress curl ($\nabla \times \tau$) enhances upwelling locally through Ekman pumping. The magnitude of the wind stress curl based on satellite vector wind data is up to $10^{-6}$ Pa/m (C. Gotschalk and M. Fewings, unpublished data) so the Ekman pumping velocity is $\sim 1$ m/day, small compared to the observed changes in MLD of $\sim 20$ m/day in the Argo data. Thus, the contribution of Ekman pumping to the changes in MLD and the SST anomalies is negligible.
Chapter 5: Summary & Conclusions

The Northern and Southern Wind Relaxations along the U.S. West Coast exhibit an event cycle spanning 12 days, which result in surprisingly different SST anomalies. This thesis answers two questions:

1. Is the $Q_{net}$ the main driver of the SST anomalies during wind relaxations?
2. What are the main contributors to the $Q_{net}$ during wind relaxations?

To answer these questions, we analyzed net surface heat fluxes from OAFlux + ISCCP during the summer months Jun—Aug 2002—2009 in the CCS. The answer to question 2 was the most clear. The SHF and LWR make relatively small contributions to $Q_{net}$, and thus the mixed-layer heat balance, in comparison to LHF and SWR. As a result, the changes in the net surface heat flux are due primarily to changes in LHF and SWR. $Q_{net}$ explains up to 90% of the warm SST anomaly during Northern Wind Relaxations and at least 45% of the strongest warm SST anomalies. Therefore, the warm SST anomaly is mainly a result of enhanced shortwave radiation (reduced cloudiness) and a decrease in cooling from the latent heat flux (weaker winds). On the other hand, $Q_{net}$ during Southern Wind Relaxation explains less than 25% of the change in SST anomaly. The $Q_{net}$ is small during Southern Wind Relaxation as a result of decreased shortwave radiation and increased longwave radiation (increased cloudiness) and a decrease in cooling from the latent heat flux (weaker winds), which have opposing effects on $Q_{net}$. The change in SST with time displays a warming trend during Southern Wind Relaxation, yet a cold SST anomaly persists from the preconditioned cold water due to the preceding wind intensification. After estimating the
terms that contribute to the residual in the ML heat budget, we hypothesize that the warming trend in SST anomaly is from reduced wind-driven vertical mixing. To test this hypothesis would require long time series of vertical profiles of upper ocean structure with higher temporal resolution than the Argo floats. The factor determining whether the net surface heat flux is the driver of SST anomalies during wind relaxations appears dependent on cloudiness. Thus, the drivers of the contrasting changes in cloudiness in Northern vs. Southern Wind Relaxations should be explored further in the CCS, as well as in other eastern boundary upwelling systems.
References


Yu, L. (2009), Sea surface exchanges of momentum, heat, and freshwater determined by satellite remote sensing. Encyclopedia of Ocean Sciences, 2, 202-211.

Figure 1. QuikSCAT satellite mean wind stress at sea level May–Aug 2000–2009 in the study region of the California Current System. Warmer (cooler) colors indicate stronger (weaker) wind stress magnitude. The maximum wind stress values are in the California Current upwelling region. Arrows are subsampled and denote wind stress direction. Small yellow dots near Point Conception, CA indicate locations of surface wind measurements from NDBC buoys. The two large black dots indicate locations of the two wind relaxations in the 12 day event cycle. The dot at the top is the Northern wind relaxation that occurs off the coast of Cape Blanco, Oregon. The dot at the bottom is the Southern wind relaxation that occurs off the coast of Point Conception, California. Figure modified from Fewings et al. (2016).
**Figure 2.** Evolution of along-mean wind stress composite anomalies (relative to the mean wind stress in figure 1) based on 67 Southern Wind Relaxations from May–Aug 2000–2009. The number in each panel indicates the number of days since the onset of the Southern Wind Relaxation (Day 0). Blue (red) anomaly indicates weaker (stronger) than the mean upwelling-favorable wind stress (Fig. 1). Blue indicates downwelling for a small number of the composited Southern Wind Relaxations. Contours indicate a wind stress anomaly of +/- 0.03 Pa. Grey indicates the anomaly is not significant at the 95% confidence level. Figure and further detail are in Fewings et al. (2016).
Figure 3. Evolution of SST composite anomalies based on 67 Southern Wind Relaxations during May–Aug 2000–2009 from AMSR-E. The number in each panel indicates time in days relative to the onset of wind relaxation at Point Conception buoys (day 0). Color indicates the SST anomaly in °C. Blue (red) indicates colder (warmer) than the mean SST. A linear seasonal trend was removed at each grid point. Grey indicates the anomaly is not significant at the 95% confidence level. Figure and caption by M. Fewings and C. Gotschalk (unpublished).
Figure 4. Mixed layer depth (m) climatology for the summer months of Jun–Aug using the density algorithm from 2000–2015. The black line is 37°N and intersects the coast at San Francisco, CA. The mean mixed layer depth north (south) of the black line is 19.5 m (25 m).
Figure 5. Composite anomalies are calculated by removing a low-pass filtered annual cycle from each latitude/longitude location (red line). The four panels are shortwave radiation (SWR), longwave radiation (LWR), sensible heat flux (SHF), and latent heat flux (LHF). The red circles show all available data from OAFlux + ISCCP 2002–2009 with overlaid black circles showing the year 2005. Gray boxes indicate the summer months of Jun–Aug that are the focus of this study.
Figure 6. Evolution of composite anomalies based on 44 Northern Wind Relaxations from Jun–August 2002–2009. The composited terms are (a) wind stress from QuikSCAT, (b) SST(day t) – SST(day -6) from OAFlux, and (c,d) 1-D surface heat budget from OAFlux + ISCCP. Only the panels during the Northern wind relaxation are shown. The number in each panel indicates time in days t relative to the onset of wind relaxation at Point Conception buoys (day 0, not shown in figure). The 50-day low-passed-filtered daily mean was removed from each grid point. Color indicates the anomaly. Blue (red) indicates wind relaxation (intensification), colder (warmer) than the mean SSTs, and surface water cooling (heating) in panels (a), (b), (c,d), respectively. Grey indicates the anomaly is not significant at the 95% confidence level. Panel (a) is plotted on a 0.1° grid and panels (b,c,d) are plotted on a 1° grid.
Figure 7. Panels represent day -1 during the Northern Wind Relaxation. Panels (a,b,c) are anomalies for the composite wind relaxation event, based on 44 events during Jun–August 2002–2009. The composited terms are (a) SST(day t) – SST(day -6), (b) the right hand side of our 1-D mixed-layer heat budget (eq. 12), and (c) Residuals. The residuals include all the other terms from the heat equation (eq. 1). The residual is calculated by subtracting panel (b) from panel (a). Panel (a), blue (red) indicates colder (warmer) SSTs than the mean. In panel (b,c) blue (red) indicates surface water cooling (heating). Grey indicates the anomaly is not significant at the 95% confidence level.
Figure 8. Monthly means for Jun–Aug 2002–2009 of the net surface heat flux ($Q_{net}$) and its components. Positive (negative) indicates a downward (outward) flux of heat into (out of) the ocean.
Figure 9. Evolution of composites based on 44 Northern Wind Relaxations from Jun–August 2002–2009. Positive (negative) indicates a downward (outward) flux into (out of) the ocean. Map extends from 140° W to 110° W longitude and 25° N to 50° N latitude. The number in each panel indicates time in days relative to the onset of wind relaxation at Point Conception buoys (day 0, not shown in figure).
Figure 10. Evolution of composite anomalies based on 44 Northern Wind Relaxations from Jun–August 2002–2009. Positive (negative) indicates a downward (outward) flux into (out of) the ocean. Map extends from 140° W to 110° W longitude and 25° N to 50° N latitude. Only the panels during the Northern wind relaxation are shown. The number in each panel indicates time in days relative to the onset of wind relaxation at Point Conception buoys (day 0, not shown in figure). Color indicates the anomaly. The 50-day low-passed-filtered daily mean was removed from each grid point. Blue (red) indicates surface water cooling (heating). Grey indicates the anomaly is not significant at the 95% confidence level.
Figure 11. Evolution of composite anomalies based on 44 Northern Wind Relaxations from Jun–August 2002–2009. Positive (negative) indicates a downward (outward) flux into (out of) the ocean. The number in each panel indicates time in days relative to the arrival of the Southern Wind Relaxations (day 0, not shown in figure). Color indicates the anomaly. The 50-day low-passed-filtered daily mean was removed from each grid point. Blue (red) indicates surface water cooling (heating). Grey indicates the anomaly is not significant at the 95% confidence level.
Figure 12. Same as in Figure 6 for the Southern Wind Relaxation. The colorbars for (b) and (c) are changed to +/- 0.25 C. The number in each panel indicates time in days relative to the onset of wind relaxation at Point Conception buoys (day 0).

\[
\Delta SST_{Obs}^{(a)} - \Delta SST_{Q_{net}}^{(b)} = \Delta SST_{Resid}^{(c)}
\]

Figure 13. Same as in Figure 7 but for day 5 during the Southern Wind Relaxation. The colorbars are changed to +/- 0.25 C. The number in each panel indicates time in days relative to the onset of wind relaxation at Point Conception buoys (day 0).
Figure 14. Same as in Figure 9, but for the Southern Wind Relaxation.

Figure 15. Same as in Figure 10, but for the Southern Wind Relaxation.
Figure 16. Same as in Figure 11, but for the Southern Wind Relaxations.
Appendix

OAFlux products

1. NWP Model Reanalyses in OAFlux

OAFlux uses three NWP reanalyses to provide information that satellites alone cannot (i.e. air temperature and humidity) and to fill in gaps of missing data. The three NWP reanalyses used in OAFlux include two versions of the National Centers for Environmental Prediction (NCEP) plus the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al., 2005). The two NCEP versions are from the National Center for Atmospheric Research (NCEP1; Kalnay et al., 1996) and the Department of Energy (NCEP2; Kanamitsu et al., 2002).

NCEP1 dataset is available from 1948–present (Kalnay et al., 1996). The purpose of NCEP2 (1979–present) is to correct errors in NCEP1 and to improve parameterizations (Kanamitsu et al., 2002). NCEP1 does not receive data updates to prevent artificial variability in the model. NCEP2 is more automated and consistent than NCEP1 and incorporates solutions to documented errors in NCEP1. NCEP1 and NCEP2 has data every 6 hours on a 1.875° grid (Yu et al., 2008). The advantage to NCEP2 is that updates have allowed for less biased meteorological observations (i.e. 2-m air temperature, humidity, and 10-m wind speed; Sun et al., 2002). The drawback is the time series does not begin until 1979. OAFlux includes both NCEP1 and its updated version NCEP2 (hereafter NCEP) for analyses. The meteorological observation fields are daily averaged and interpolated linearly onto a 1° spatial grid.
The third NWP used in OAFlux is ERA-40. The ECMWF released a second-generation reanalysis product, ERA-40 that spans 1957–2002 (Uppala et al., 2005). ERA-40 ends during our study period of 2002-2009. The majority of OAFlux results used in this thesis are not influenced by ERA-40.

2. SST products in OAFlux

OAFlux uses the AVHRR package from NOAA Optimum Interpolation 0.25° daily SST analysis product by Reynolds et al. (2007) (hereafter Reynolds OI). Yet, many oceanographers prefer to use the combined AVHRR infrared with AMSR-E microwave (hereafter AVHRR + AMSR-E). Below, I explain the differences between AMSR-E and AVHRR and then explain the reason why Yu et al. (2008) chose the AVHRR SSTs.

AMSR-E began operating in 2002 and is capable of measuring SST through clouds (Chelton and Wentz, 2005). AMSR-E covers 89% of the globe daily, has a resolution of 56 km, and excludes rain-contaminated measurements (Chelton and Wentz, 2005). Besides SST, AMSR-E measures wind speed (discussed below).

Satellite measurements of SSTs with a finer spatial resolution (~1 km) come from the infrared (IR) Advanced Very High Resolution Radiometer (AVHRR). The AVHRR is a five-channel scanner that has flown on board the National Oceanic and Atmospheric Administration (NOAA) series of Polar Orbiting Environmental Satellites (POES) in sun-synchronous orbits since November 1982 (Yu 2008). AVHRR represents the longest global record of IR SST retrievals (Reynolds et al., 2007). Yet, the biggest challenge in
retrieving SST from an IR instrument is the cloud contamination problem, because clouds are opaque to IR and mask radiation from the ocean surface.

Even with limited SST fields due to a resolution of 56 km, AMSR-E is a better option than AVHRR in the CCS because it can measure SSTs through clouds. This is advantageous in the CCS during the study months of Jun–Aug, because cold-upwelled waters within ~200 km of the coast result in low marine stratus clouds (Kosro 1987; Rahn and Parish, 2008). AVHRR coverage drops to less than 30% due to cloudiness (Chelton et al., 2007), while the AMSR-E coverage is about 95% (Chelton and Wentz, 2005).

OAFlux uses the AVHRR package from Reynolds OI. Reynolds OI has two packages: one uses AVHRR only and the other combines AVHRR infrared with AMSR-E microwave (hereafter AVHRR + AMSR-E) SST data. Both products use in situ data from ships and buoys to adjust satellite biases. The daily SST analysis uses AVHRR from the Pathfinder reanalysis project that began in January 1985. There is a gap in Pathfinder data between November 1982 and January 1985 (during NOAA-7 flying period) due to sparse buoy data. OAFlux averages AVHRR from Reynolds OI onto a 1° grid for SST analysis. OAFlux uses satellite SST measurements from AVHRR, to ensure consistency throughout the OAFlux dataset (Yu et al., 2008).

For a consistent analysis of changes to the regional SSTs and net surface heat fluxes, we use SSTs from OAFlux instead of SSTs from AMSR-E alone (Fig. 3). The AMSR-E microwave SST is not used for two reasons. The first is that the mean SSTs from AVHRR and AMSR-E are different because AMSR-E can penetrate clouds and
thus provide more coverage over time. Thus, the mean differences between the two satellites would need to be resolved if AMSR-E is included. The second reason is the NWP reanalyses use Reynolds OI AVHRR weekly SST products as a lower boundary conditions. Thus, the consistency between air temperature and SST might worsen if OAFlux used AMSR-E SST (Yu et al., 2008).

3. Wind Velocity product in OAFlux

Estimates of 10-m equivalent neutral winds $U$ come from 3 satellite products (Table 1): SSM/I version 6 radiometer, AMSR-E version 5 radiometer, and QuikSCAT version 3.

SSM/I (July 1987-present) has a sun-synchronizing orbit with 1394-km swath that covers 82\% of the earth surface in a day, and covers the globe within three days (Wentz, 1997). The data are available at a resolution of 12 hours and at a swath resolution of 25 km. Flagged wind speeds occur if cloud/rain liquid water values exceed 18 mg cm\(^{-2}\), because the accuracy of the wind speed retrievals degrades in the presence of rain.

AMSR-E version 5 measures wind speed. AMSR-E agrees well with SSM/I, with a root-mean-square difference of 0.51 m s\(^{-1}\) during the summer months of Jun–Aug 2002 (Wentz et al., 2003). AMSR-E does not agree well with QuikSCAT (described in section 1) because their equator node times are about 6 hours different. The wind data (Table 1) are averaged onto the daily and 1° grid resolutions of OAFlux.
4. LHF + SHF from OAFlux

Scientists use bulk algorithms, such as COARE 3.0 (Fairall et al., 2003) to estimate SHF and LHF by using near-surface meteorological variables. It is important to estimating net surface heat fluxes within an accuracy of 10 W m\(^{-2}\) and requires resolution of diurnal solar warming of the ocean surface (the warm layer), and diffusive cooling of the viscous sub-layer known as the cool skin (Fairall et al., 1996b). The COARE bulk flux algorithm models the warm layer and cool skin corrections to the bulk ocean temperature. The cool skin correction makes a larger difference to the fluxes than the warm skin. Ignoring it can cause 10% overestimation of sensible and latent heat flux (Marion, 2014). Even though OAFlux ignores the cool skin when using the COARE algorithm, OAFlux estimates the R/V Revelle SHF and LHF to within 0.1% of the mean flux (Marion, 2014).

The COARE 3.0 bulk algorithm for the turbulent \((Q_{LH}, Q_{SH})\) heat fluxes needs inputs of 10-m wind speed \((U)\), 2-m air temperature \((T_a)\), \((SST)\), sea surface humidity \((q_s)\) and 2-m near-surface specific humidity \((q_a)\) from OAFlux. The bulk formulas for the turbulent fluxes follow the Monin-Obukhov similarity approach (Liu et al., 1979):

\[
Q_{LH} = -\rho_a L_E c_E U (q_s - q_a) \quad (B1)
\]

\[
Q_{SH} = -\rho_a c_p c_h U (SST - T_a) \quad (B2)
\]

where \(\rho_a\) is the density of air, \(c_p\) is the specific heat capacity (4000 kg\(^{-1}\) J K\(^{-1}\)) of the air at constant pressure, and \(c_E\) and \(c_h\) are the stability and depth-dependent turbulent exchange coefficients for latent and sensible heat, respectively. OAFlux uses COARE 3.0 instead of the updated 3.5 version. The difference between versions 3.0 and 3.5 is
greatest at higher wind speeds, thus does not affect the accuracy of this study. SST is the blend of Pathfinder AVHRR with NWP models (Table 1) mentioned above. OAFlux uses estimates of near surface air temperature ($T_a$) from NCEP. The latent heat of vaporization $L_E$ is a function of SST and approximated by the linear equation (Bolton, 1980, eq. 2):

$$L_E = (2.501 - 0.00237 \times SST^\circ C) \times 1.06 \times 10^6 J kg^{-1}. \quad (B3)$$

The surface humidity, $q_s$, is computed from the saturation humidity, $q_{sat}$, for pure water at $SST$,

$$q_s = 0.98 \times q_{sat} (SST) \quad (B4)$$

where a multiplier factor of 0.98 is used to take into account the reduction in vapor pressure caused by salt water. The Chou et al. (1997) empirical orthogonal function (EOF) technique computes the near-surface humidity $q_a$ based on total precipitable water from Special Sensor Microwave Imager (SSM/I) column water vapor retrievals and field humidity soundings. OAFlux uses 2-m height adjusted $q_a$ from Chou et al. (2001) 1° daily 10-m products (1988–2000) in the Goddard Satellite-Based Surface Turbulent Fluxes Version 2 dataset objectively blended with NWP models. This version ends in 2000, thus our study uses only the NWP modeled $q_a$ (Table 1).

5. OAFlux uses ISCCP LWR + SWR

ISCCP provides global LWR and SWR (1984—2009) and is constructed using the NASA GISS radiative transfer model and satellite data (Zhang et al., 2004). ISCCP began in 1983 to determine the physical properties of clouds from satellite
measurements. NASA GISS improved observations of the physical properties of the surface, atmosphere, and clouds based on the ISCCP D-series cloud climatology datasets (Rossow and Schiffer, 1999) and several other satellite data products. ISCCP data is available every 3 h on a 2.5° grid. For consistency, OAFlux uses ISCCP daily fluxes linearly interpolated onto a 1° grid.