ABSTRACT

Two extremely sharp fronts with changes in sea surface temperature >0.4°C over lateral distances of ~1 m were observed in the equatorial Pacific at 0°, 140°W and at 0.75°N, 110°W. In both cases, layers of relatively warm and fresh water extending to ~30-m depth propagated to the southwest as gravity currents. Turbulent kinetic energy dissipation rates averaging $4.5 \times 10^{-5}$ W kg$^{-1}$ were measured with a microstructure profiler within the warm layer behind the leading edge of the fronts—1000 times greater than dissipation in the ambient water ahead of the fronts. From satellite images, these fronts were observed to propagate ahead of the trailing edge of a tropical instability wave (TIW) cold cusp. Results from an ocean model with 6-km grid resolution suggest that TIW fronts may release gravity currents through frontogenesis and loss of balance as the fronts approach the equator and the Coriolis parameter weakens. Sharp frontal features appear to be ubiquitous in the eastern tropical Pacific, have an influence on the distribution of biogeochemical tracers and organisms, and play a role in transferring energy out of the TIW field toward smaller scales and dissipation.

1. Introduction

Regions of strong lateral gradients in the upper ocean, known as fronts, are important locations for exchanges between the atmosphere, surface ocean, and ocean interior (Ferrari 2011). Because of the presence of strongly sloping isopycnals, vertical velocities at fronts can be an order of magnitude higher than elsewhere, reaching 100 m day$^{-1}$ at midlatitude submesoscale fronts (Thomas et al. 2008). Such large vertical velocities can drive the subduction of carbon and heat into the ocean’s interior or upwell nutrient-rich deep water into the surface layers leading to enhanced biological productivity (Lévy et al. 2012). This vertical exchange can be further enhanced by the vigorous turbulence frequently found at fronts (D’Asaro et al. 2011). Fronts also participate in scale interactions between different components of the ocean circulation. Frontogenesis, the intensification of fronts through lateral strain, is an important mechanism for transferring energy from...
the 100-km lateral scales of mesoscale eddies to the small scales at which energy is dissipated (Capet et al. 2008; Molemaker et al. 2010). The importance of fronts for vertical exchange processes and ocean energetics poses a challenge for climate models that cannot resolve them (Bryan et al. 2010; Soufflet et al. 2016). Thus, detailed process studies of fronts are critical to better understand and parameterize them.

In this article, we present observational and modeling work aimed at better understanding sharp fronts in the eastern equatorial Pacific Ocean. The two major fronts in this region are the north and south equatorial fronts that bound the equatorial cold tongue (NEF and SEF in Fig. 1). Far from being persistent, stationary features, both the north and south equatorial fronts are strongly shaped by eddy activity in the form of tropical instability waves (TIWs). TIWs typically form in boreal summer and autumn when the equatorial circulation intensifies in response to strengthening trade winds (De Szeoke et al. 2007) and becomes unstable to barotropic and baroclinic instabilities (Philander 1976; Cox 1980; Masina et al. 1999). TIWs strongly influence the horizontal and vertical circulation of the region (Qiao and Weisberg 1997; Perez et al. 2010), the upper-ocean heat budget (Menkes et al. 2006; Jochum and Murtugudde 2006), and the overall biological productivity of the equatorial cold tongue (Strutton et al. 2001; Menkes et al. 2002; Evans et al. 2009). TIWs shape the north equatorial front into a series of cold cusps and warm troughs that oscillate between the equator and 5°N with periods between 15 and 30 days (Chelton et al. 2000; Caltabiano et al. 2005). The frontal regions on either side of the TIW cold cusps can become extremely sharp, but little is known about how this sharpening occurs or how these sharp gradients later dissipate.

There are a number of previous observational studies of very sharp fronts associated with TIWs. Yoder et al. (1994) and Archer et al. (1997) describe observations of a strong, remarkably straight front visible from ships, aircraft, and satellites spanning from 2° to 4°N near 140°W. The front was characterized by strong north–south surface velocity convergence with a 70-m-thick layer of cold water subducting below a 40-m-thick layer of warm water. Yoder et al. (1994) proposed that the strong surface convergence was responsible for the accumulation of buoyant diatoms on the warm side of the front. Johnson (1996) presents a more detailed study of the same set of observations that highlights the strong downward vertical velocity associated with the front, reaching almost 1 cm s⁻¹, and the advection–pressure gradient force balance, suggesting that the front propagated as a gravity current. However, he also suggested that the front (characterized by a 1.5°C surface temperature change over ∼1 km) must be embedded in a larger-scale geostrophically balanced flow for which the Coriolis force is important. He does not discuss the mechanism through which the front may have initially formed. The intermediate values of momentum and density created in the lee of the front imply intense levels of turbulence dissipation and mixing within the front, but there were no direct measurements of turbulence available to corroborate this. Flament et al. (1996) discuss the connections between the large-scale circulation of a tropical instability vortex (TIV) and the frontal regions on its fringes.

The observations of Yoder et al. (1994), Johnson (1996), and Archer et al. (1997) were of a front on the leading edge of a TIW cold cusp separating the warm water within the North Equatorial Counter Current (NECC) from the cold equatorial water to the southeast (see Fig. 1). Strong fronts can also occur on the trailing edge of TIW cold cusps (Fig. 1). These trailing edge fronts can approach the equator, where the Coriolis force weakens and the dynamics change. Model results suggest that at the ∼100-km scale these fronts obey a cyclogeostrophic force balance, with the centrifugal force associated with the strongly curved nature of the flow playing a significant role (Holmes et al. 2014). The process of frontogenesis and sharpening along such a front can differ from that at straight fronts (Shakespeare 2016) and may contribute to the generation of anticyclonic vorticity within the TIVs (Holmes et al. 2014). Trailing edge fronts are also interesting because of their proximity to the complex and turbulent subsurface region above the Equatorial Undercurrent (EUC; Lien et al. 2002; Smyth et al. 2013; Moom et al. 2013). The trailing edge front is located in the transition phase between northward and southward TIW flow, thought to be a phase in which the TIWs enhance turbulence levels in the upper EUC (Lien et al. 2008; Moom et al. 2009; Inoue et al. 2012; Holmes and Thomas 2015). It remains an open question as to how the fronts in this region influence and interact with the existing turbulent mixing.

The formation of sharp fronts could also play a role in the dissipation of TIW energy. Marchesiello et al. (2011) examined the submesoscale activity associated with TIWs and showed that a forward energy cascade emerges at larger scales than in the midlatitudes, constituting a potential route to small-scale dissipation for TIWs. Previous proposals for the dissipation mechanism of TIWs include reabsorption by the mean flow (Qiao and Weisberg 1998), the meridional radiation of Rossby waves (Farrar 2011; Holmes and Thomas 2016), and the generation of internal waves that propagate into the thermocline and deep ocean (Tanaka et al. 2015). However, many TIW energetics studies are based on models that do not resolve the finescale processes, and there have been few direct observations of the small-scale (<10 km) dynamics associated with TIWs.

The small length scales and location close to the equator where the effect of planetary rotation is weak suggests that
Advection may play an important role in the force balance of these sharp features and that they may propagate as gravity currents (Johnson 1996). Fronts that propagate as gravity currents are found in many natural and industrial contexts (Simpson 1997; Huppert 2006). River plumes can propagate as gravity currents (e.g., Garvine and Monk 1974; Luketina and Imberger 1987), induce shear instabilities (Tedford et al. 2009), and are highly turbulent at frontal convergences (Nash and Moun 2005; Kilcher et al. 2012; Jurisa et al. 2016). Fronts have also been observed far from land following rain events that create freshwater lenses on the ocean’s surface whose edges propagate as gravity currents (Moulin 2016). The theoretical speed of a steady-state gravity current propagating into a homogeneous ambient was first quantified by Benjamin (1968). Further studies have added the effect of a stratified
ambient (Maxworthy et al. 2002; Ungarish 2006; White and Helfrich 2008) and two-layer stratified ambients (Xu 1992; White and Helfrich 2012). These previous studies provide a framework to understand the dynamics of the observed equatorial fronts.

In this article, we describe observations of two very sharp temperature fronts extending to \( \sim 30-40\) m depth found near the trailing edge of two TIW cold cusps. The first of these occurred on 2 November 2008 at \( 0^\circ, 140^\circ \) W and the second occurred on 24 November 2014 at \( 0.75^\circ \) N, \( 110^\circ \) W. Both fronts were oriented northwest–southwest and propagated to the southwest as gravity currents. We refer to these as small-scale fronts (SSFs) to differentiate them from the leading edge and trailing edge fronts that delineate the TIWs. We address large-scale aspects of the dynamics with a high-resolution ocean model used to investigate how SSFs may have been generated and how they relate to TIW circulation.

This paper is organized as follows: In section 2, we introduce the shipboard observations and the numerical model setup, followed by detailed descriptions of observed temperature, salinity, velocity, microstructure turbulence, and biological matter found in the vicinity of the SSFs in section 3. The propagation kinematics and energetics of the observed SSFs are examined in section 4. The model results are discussed in section 5 and related to the observations. We discuss the prevalence of sharp fronts in the eastern equatorial Pacific in section 6. Finally, we summarize our conclusions in section 7.

2. Methods

a. Shipboard and moored observations: \( 0^\circ, 140^\circ \) W, 2 November 2008

The first of two SSFs was observed on 2 November 2008 at 0142 UTC during the Equatorial Internal Wave Experiment (EQUIX; Moum et al. 2009; Inoue et al. 2012; Smyth et al. 2013), which was a collaborative effort designed to study links between turbulence and internal waves in the equatorial cold tongue at \( 0^\circ, 140^\circ \) W.

Three measurement platforms were used during EQUIX: a ship and two moorings. The R/V Wecoma was roughly statically positioned allowing continuous temperature, salinity, 880-nm optical backscatter, and turbulence measurements to 200 m using the Chameleon microstructure profiler (Moum et al. 1995), and velocity was measured with a suite of three hull-mounted acoustic Doppler current profilers (ADCPs). For our purposes, we use the 300-kHz ADCP with 2-m vertical bin size. The blanking distance of the ADCP means that the shallowest velocity measurement was at 10 m. We estimate the velocity between 0 and 8 m by taking a quadratic fit over the depth range 10–48 m and extrapolating to the surface.

A moored upper-ocean buoy provided by the University of Washington’s Applied Physics Laboratory (EQUIX mooring; Inoue et al. 2012) provided temperature and salinity measurements from 20 Sea-Bird 37 MicroCATs and turbulence measurements from 10 \( \chi \) pods (Perlin and Moum 2012). The majority of instruments were tightly clustered at \( \sim 1\) m spacing from 21- to 50-m depth. Four ADCPs provide velocity measurements in the upper 500 m. The National Oceanic Atmospheric Association’s (NOAA) Tropical Atmosphere–Ocean (TAO) mooring (McPhaden et al. 1998) has been almost continuously occupied at \( 0^\circ, 140^\circ \) W since 1983, measuring temperature, salinity, and velocity. Since 2005, \( \chi \) pods have also been included on this mooring at 29-, 39-, 49-, 59-, and 84-m depths.

b. Shipboard observations: \( 0.75^\circ \) N, \( 110^\circ \) W, 24 November 2014

The second SSF was observed from the R/V Oceaneus on 24 November 2014 at 0259 UTC, while the ship steamed eastward at 2 \( \text{kt} \) \( (1 \text{ kt} = 0.51 \text{ m s}^{-1}) \). After the first front crossing, the ship turned around and crossed the SSF two additional times, about 2 and 4 h later.

The ship was equipped with a variety of instrumentation, including Chameleon, to provide detailed observations of the SSF. During the first crossing of the front, Chameleon was deployed to 200 m about once every 5 min. Subsequently, Chameleon was deployed only to 50 m, allowing a profile every minute with the ship traveling at 2 \( \text{kt} \). Chameleon profiles were taken at nominal 300-m spacing during the first crossing and 60-m spacing during the second and third crossings. Throughout this paper, only the first crossing is plotted because of the deeper range of Chameleon, but each crossing shows the same general characteristics of the SSF.

A hull-mounted 300-kHz RD Instruments ADCP measured velocity over a vertical range of 10–80 m with a bin size of 2 m. The same method as for the 2008 front is used to extrapolate velocities to the surface. During the second and third front crossings, the ship’s X-band (\( \sim 3\) cm) radar was tuned to detect Bragg scattering off ripples that indicate surface convergences; 56 images of X-band radar during a 1-h time period were recorded and used to map the orientation and propagation speed of the SSF over a radius of about 5 km from the ship. A chain of 21 high-frequency thermistors was mounted off the ship’s bow in undisturbed water to about 10-m depth. A 100-Hz thermistor at 2.25 m provided a well-resolved record of near-surface temperature. The ship’s flowthrough system provided additional temperature, salinity, and fluorescence measurements at an intake located in the hull at 3-m depth.
c. Numerical model

Output from a set of nested simulations of the Pacific Ocean performed using the Regional Ocean Modeling System (ROMS; Shchepetkin and McWilliams 2005) is used to investigate how SSFs may form and propagate. The outer nest consists of a Pacific basinwide simulation over the region 30°S to 30°N, 240° to 70°W, with 0.25° horizontal resolution, 50 vertical levels, and a time step of 10 min. This simulation was initialized from the 20 vertical level Pacific basinwide simulation of Holmes et al. (2014) and spun up for an additional 5 years to equilibrate to the increased vertical resolution. Daily climatological surface forcing was taken from the Common Ocean Reference Experiment Normal Year Forcing field (CORE-NYF; Large and Yeager 2004). At the meridional boundaries, temperature and salinity were nudged to climatological values from the World Ocean Atlas (Locarnini et al. 2010; Antonov et al. 2010), while zonal and meridional velocities were nudged to zero. The K-profile parameterization (KPP) vertical mixing scheme was used to parameterize subgrid-scale vertical mixing processes (Large et al. 1994).

In the last year of the basinwide simulation, a 0.05° (~6 km) resolution simulation was nested inside the larger domain over the region 5°S to 10°N, 155° to 120°W. A combination of nudging and radiation boundary conditions was used to transfer fields from the basinwide to the high-resolution simulation, with the exception of a clamped condition on the eastern boundary tracers. Horizontal diffusion of momentum is achieved with a biharmonic viscosity with coefficients of $1 \times 10^{10}$ (high resolution) and $1 \times 10^{11}$ m$^3$ s$^{-1}$ (basinwide), and horizontal diffusion of salinity and temperature is achieved with a harmonic diffusion with a coefficient of 100 m$^2$ s$^{-1}$. ROMS has been successfully used for process studies of TIWs under similar configurations (Marchesiello et al. 2011; Holmes et al. 2014; Holmes and Thomas 2015).

3. Observed equatorial fronts

a. Satellite observations

The north equatorial front and its TIW-driven meanders are clearly visible from satellites. However, the spatial resolution of satellite data is low relative to the scale of the SSFs that are the focus of this paper. The location and orientation of the SSFs relative to the TIW trailing edge front are shown in images of 3-day composite averages of sea surface temperature (SST) from the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI; Wentz et al. 2015) at the time of the two encounters in November 2008 and 2014 (Fig. 2). During November 2008, the Oceanic Niño index (ONI; http://www.cpc.noaa.gov/products/analysis_monitoring/enso/years.shtml) was negative (~0.6) at the beginning of a weak La Niña, whereas during November 2014, the ONI was positive (0.6) during the first month of the weak 2014 El Niño that would later grow into the strong 2015–16 El Niño. Typically, TIWs are much stronger during La Niña and neutral years than during El Niño years (Philander 1990), which may be why the TIWs observed during 2008 were stronger than in 2014 (Fig. 2).

Here, we denote the northern edges of the TIW cold cusps with the 25.3° and 26.2°C isotherms (Fig. 2), which highlight the leading edge fronts and trailing edge fronts as depicted in Fig. 1. Trailing edge fronts are observed to come close to—and sometimes traverse—the equator. TIW cold cusps propagate westward at a rate of ~0.4 m s$^{-1}$. Both of the SSFs discussed herein were observed to the southwest of a trailing edge front.

b. Shipboard observations

Both SSFs were identified by abrupt increases in SST. On 2 November 2008, SST increased by 0.45°C within 7 s, as measured by R/V Wecoma’s flowthrough system (Fig. 3a). The same front was detected about an hour later at the EQUIX mooring by a χpod at 21 m with two temperature sensors 1.1 m apart (Fig. 3b). The χpod measures temperature at 100 Hz and therefore shows much sharper variations in temperature than the shipboard flowthrough measurement. At 21 m, the total change in temperature was smaller (0.34°C) and the front was fragmented. On 24 November 2014, a high-speed thermistor attached to the bow chain on the R/V Oceania at 22.5 m and detected a temperature increase of 0.44°C in 0.88 s, equivalent to ~1-m lateral distance at the nominal ship speed of 2 kt (Figs. 3c,d). The reaction time of the flowthrough temperature sensors is comparably much slower than the thermistors on the bow chain (Fig. 3c), and detailed features of the front’s structure can only be seen with high-speed thermistors (Figs. 3b,d).

Chameleon provides a detailed look at the 2D structure of the SSFs and surrounding waters. The abrupt increase in surface temperature (Figs. 4a,h) was accompanied by a decrease in salinity of 0.26 and 0.10 psu, respectively, in 2008 and 2014 (Figs. 4b,i). The decreases in density across the fronts were ~0.35 kg m$^{-3}$ in 2008 and ~0.20 kg m$^{-3}$ in 2014 (Fig. 4, contours). In general, we will refer to the relatively warm and fresh water behind the front as the warm layer and the colder region ahead of the front and below the warm layer as the ambient water. The depth of the warm layer is taken to be the average depth of isopycnals that outcrop at the
front once they deepen to a near-constant depth. In 2008, the warm layer was $37.9 \pm 4.4$ m deep calculated from the $1023.57-1023.67 \text{ kg m}^{-3}$ isopycnals (Figs. 4a,b), and in 2014, the warm layer was $28.3 \pm 3.7$ m deep calculated from the $1023.40-1023.50 \text{ kg m}^{-3}$ isopycnals (Figs. 4h,i). These depths were shallower than the front observed by Johnson (1996), who showed the warm layer extending below 60 m.

**Fig. 2.** TMI 3-day-averaged satellite SST ending on (a) 3 Nov 2008 and (b) 25 Nov 2014. TIWs are visible along the north equatorial front, which is shown with the 25.3°C and 26.3°C isotherms (thin black contours). The location, orientation, and propagation direction of the observed small-scale fronts that cannot be seen with the satellite data are plotted with thick black lines/arrow.

**Fig. 3.** Observed temperature in 2008 from (a) the ship’s forward intake at 3 m (0.5 Hz) and (b) a χpod on the EQUIX mooring at 21-m depth, which had two temperature sensors 1.1 m apart (10 Hz). The forward intake temperature does not have the necessary temporal resolution to see the sharpness of the front in the way that the χpod does; however, the measurements by the χpod were much deeper and do not show as big of an overall temperature jump. (c) Observed temperature for the 2014 front from the ship’s forward intake at 3 m (0.25 Hz; gray) and from a high-speed thermistor at 2.25 m on a bow chain (100 Hz; black). (d) An enlarged image of high-speed thermistor temperature shows the sharpness and exact arrival time of the 2014 front.
FIG. 4. Time-depth slices for the (left) 2 Nov 2008 and (right) 24 Nov 2014 fronts. (a),(h) Potential temperature and (b),(i) salinity measured by Chameleon; color axes do not have the same limits but cover the same range. (c),(j) Four times the squared buoyancy frequency. ADCP Velocity in the (d),(k) eastward $u$ and (e),(l) northward $v$ directions. (f),(m) Squared shear. (g),(n) Turbulent kinetic energy dissipation rate (log scale) as measured by shear probes on Chameleon. Isopycnals are contoured at intervals of 0.2 kg m$^{-3}$ (thin black) with the 1024, 1025, and 1026 kg m$^{-3}$ isopycnals in thicker black and labeled in (a) and (h).
In 2008, Chameleon profiles were taken continuously for many days at 0°, 140°W. The warm layer was observed to persist after the passage of the SSF for at least 7 days until the end of the experiment (Fig. 4a in Inoue et al. 2012). Over the 8 days prior to the SSF’s crossing, the mean temperature in the upper 10 m was 24.90°C with a standard deviation of 0.22°C. This is nearly 0.6°C cooler than the mean temperature over the subsequent 7 days, which was 25.48° ± 0.18°C.

Elevated values of shear and stratification were observed at the base of the warm layer especially in the region right behind the front where the warm layer deepened. Buoyancy frequency ($N^2 = -g \rho_0 \rho$, where $\rho_0 \rho$ and gravity $g$ are small, and the subscript $z$ indicates a partial derivative in the vertical direction. We find high stratification with $N^2 > 10^{-3} \text{s}^{-2}$ at the base of the warm layer compared to lower $N^2 < 10^{-4} \text{s}^{-2}$ in the ambient water (Figs. 4c,j). Velocity in the zonal direction $u$ changed abruptly with the passage of the front (Figs. 4d,k), whereas changes in the meridional velocity $v$ are smaller (Figs. 4e,l). In both years, shear ($S^2 = u^2 + v^2$) was elevated within the warm layer and was especially strong at the base of the warm layer in the vicinity of the 1023.6 and 1023.4 kg m$^{-3}$ isopycnals in 2008 and 2014, respectively, where $S^2$ consistently exceeded values ($N^2 = -g \rho_0 \rho > 10^{-3} \text{s}^{-2}$) (Figs. 4f,m).

The measured values of turbulence kinetic energy dissipation rates $\varepsilon$ within the warm layer were found to be at least $10^{-6} \text{W kg}^{-1}$, with many instances exceeding $10^{-4} \text{W kg}^{-1}$ (Figs. 4g,n). In contrast, much of the ambient water extending from the surface to the depth of the warm layer tended to have $\varepsilon$ below $10^{-8} \text{W kg}^{-1}$ (Figs. 4g,n). To put the 2008 warm layer dissipation rates in context, they were similar to other nighttime values measured during the EQUIX experiment but maintained uncharacteristically high during the subsequent daytime when surface $\varepsilon$ typically dropped by two orders of magnitude (Fig. 5a in Inoue et al. 2012). Compared to a river plume, average $\varepsilon$ in the warm layer was an order of magnitude smaller than those measured within the Fraser (MacDonald and Geyer 2004) and Columbia (Kilcher et al. 2012) river plumes at peak ebb.

Looking deeper, the impact of the EUC must be considered. The region above the EUC core is typically characterized by high $S^2$ and $\varepsilon$ (Smyth and Moum 2013), and this was observed during both field campaigns despite the difference in EUC depth and strength at the two locations (Figs. 4f,g,m,n). In 2008, at 140°W, the core of the strong eastward-flowing EUC was at 110 m and extended vertically between 75 and 150 m (Fig. 4d). In 2014, at 110°W, the core of the EUC was at 75 m and extended vertically between 55 and 110 m (Fig. 4k). The observations in 2008 were made right at the equator, whereas in 2014, they were made at 0.75°N where EUC velocities were weaker (Figs. 4d,k). High dissipation was present prior to the passage of the fronts on the upper flank of the EUC. During 2008 there was a region from 50 to 70 m of low- $\varepsilon$ water between the base of the warm layer and the top of the EUC (Fig. 4g). In 2014, because of the shallower nature of the EUC at 110°W, the highly dissipative warm layer intersected the highly sheared and turbulent region on the upper flank of the EUC (Fig. 4n). The low- $\varepsilon$ water ahead of the front appeared to be subducted under the warm layer over the first 2 km but eventually disappeared, or was mixed, between these two layers of high dissipation.

c. Biological contrasts across the front

In both cases, higher levels of 880-nm optical backscatter were observed in the ambient water on the cold side of the front than within the warm layer (Figs. 5a,c), indicating differences in biological productivity in the two water masses. It is expected that the cold tongue water would have higher biological productivity because of the increased nutrient concentration from upwelling (Murray et al. 1994). In 2014, an additional sharp peak in fluorescence was observed from the ship’s flowthrough system (Fig. 5d), suggesting the possibility of the accumulation of fluorescent particles at the convergent front. Yoder et al. (1994) observed an accumulation of diatoms at the leading edge front separating the cold, nutrient-rich, upwelled water of the South Equatorial Current (SEC) and the warmer water of the NECC.

4. Front kinematics and energetics

a. Front propagation and orientation

The propagation speed and direction of the 2008 front were found by comparing arrival times of the SST front at the three components of the EQUIX array—the R/V Wecoma, the EQUIX mooring, and the TAO mooring (Fig. 6)—in a method similar to that used by Scotti et al. (2005) to estimate internal wave speeds through the four beams of an ADCP. To estimate the front’s velocity with three measurements of SST, it is assumed that the velocity remained constant and that the front formed a straight line that propagated perpendicularly to its orientation.

The front can be defined in Cartesian space as $ux + vy = 0$, where $(u, v)$ are the propagation velocities in the $(x, y)$ directions, respectively. The perpendicular distance $d$ between a point $(x_0, y_0)$ and the front is

$$
d = \frac{|ux_0 + vy_0|}{\sqrt{u^2 + v^2}}. \quad (1)
$$

Using the fact that distance is the product of velocity and time ($d = \sqrt{u^2 + v^2}t$), we can calculate the time it would take for the moving front to intersect two stationary points $(x_1, y_1, t_1)$ and $(x_2, y_2, t_2)$ as

$$
t_1 = 0, \quad t_2 = \frac{d}{\sqrt{u^2 + v^2}}.
$$
\[(t_2 - t_1) = \frac{|u(x_2 - x_1) + v(y_2 - y_1)|}{u^2 + v^2}, \tag{2}\]

which in vector form is

\[\Delta t_k + \Delta r_k = \frac{\Delta x_k \cdot u_{fo}}{u_{fo} \cdot u_{fo}}, \tag{3}\]

where the subscript \(k\) denotes location pairs, \(\Delta t_k\) is the difference in arrival times between location pairs, \(\Delta x_k = (\Delta x_k, \Delta y_k)\) is the distance between location pairs, and \(u_{fo} = (u_{fo}, v_{fo})\) is the velocity of the front in Earth’s reference frame (denoted by subscript \(fo\)). An additional error term \(\Delta r\) has been added to (3) to allow a least squares error solution to be computed following the method detailed in appendix A.

Using the three locations and arrival times for the 2008 front, it is found that the front propagated at \(u_{fo} = 0.56 \text{ m s}^{-1}\) at compass heading 246.5° (Fig. 6). With only three measurements, no error can be computed because all three measurements are used to constrain the propagation speed and direction.

In 2014, the shipboard X-band (~3 cm) radar was tuned to detect Bragg scattering off of surface ripples that form at convergences, thereby allowing the front orientation and propagation velocity to be mapped over a 5-km radius (Fig. 7). Similar to the 2008 front, a least squares error solution is used to calculate the propagation speed and direction using (3). The biggest difference is that in 2008 only three crossing times and position are known, whereas during 2014, 56 radar images give many hundreds of time and location pairs of the moving front. It is found that the front propagated at \(u_{fo} = 0.25 \pm 0.04 \text{ m s}^{-1}\) at compass heading 226.7° ± 0.2°.

To find the speed of the front in the reference frame of the moving fluid, the ambient velocity of the underlying water must be removed. The ambient velocity \((u_{bkgrnd}, v_{bkgrnd})\) is calculated by taking the mean shipboard ADCP velocity over 10- to 30-m depth, 45 to 15 min before each front’s crossing. It is assumed that \(u_{bkgrnd}\) and \(v_{bkgrnd}\) are in steady state. The ambient velocity is then rotated into the reference frame of the front \((u'_{bkgrnd}, v'_{bkgrnd})\), where primes denote the rotated reference frame for all variables with positive \(u'_{bkgrnd}\) in the across-front direction to the northeast and \(v'_{bkgrnd}\) in the along-front direction to the northwest. Then, \(u'_{bkgrnd}\) is subtracted from the observed front velocity in the fixed reference frame \(u'_{fo}\) to find the velocity of the fronts in the reference frame of the moving fluid \(u'_{front}\):

\[u'_{front} = u'_{fo} - u'_{bkgrnd}. \tag{4}\]

In 2008, \(u'_{fo} = -0.56 \text{ m s}^{-1}\) and \(u'_{bkgrnd} = -0.35 \text{ m s}^{-1}\), resulting in a net frontal propagation speed of \(u'_{front} = -0.21 \text{ m s}^{-1}\). In 2014, \(u'_{fo} = -0.25 \pm 0.04 \text{ m s}^{-1}\) and \(u'_{bkgrnd} = -0.08 \text{ m s}^{-1}\), leading to a net frontal propagation speed of \(u'_{front} = -0.17 \pm 0.04 \text{ m s}^{-1}\).

b. Velocities within the front

The across- and alongfront velocity \(u'\) and \(v'\) in the vicinity of the fronts (Fig. 8) can be calculated by
rotating the ship’s ADCP velocity into the reference frame of the fronts \( u_{0}^{\text{obs}} \) and \( y_{0}^{\text{obs}} \) and subtracting the ambient velocity \( u_{0}^{\text{bkgrnd}} \) and \( y_{0}^{\text{bkgrnd}} \):

\[
(u', y') = (u_{0}^{\text{obs}} - u_{0}^{\text{bkgrnd}}, y_{0}^{\text{obs}} - y_{0}^{\text{bkgrnd}}).
\]  

(5)

A strong alongfront velocity to the northwest is present in both fronts \( y' \) (Figs. 8b,e). We suggest in section 5 that this alongfront velocity originates from the cyclogeostrophic balance of the TIW frontal flow (Holmes et al. 2014).

In the across-front direction, it is useful to remove the propagation speed of the fronts \( u'_{\text{frnt}} \) to view the warm layer in the reference frame traveling with the front (Figs. 8c,f). In addition, we add streamlines that show the cross-front circulation pattern. This reveals that in the upper half of the warm layer, the fluid was moving in the across-front direction at a velocity faster than the propagation speed of the front. In the lower half of the warm layer, the velocity was near zero or slightly opposite to the direction of front propagation, setting up a recirculation pattern within the warm layer, similar to that observed by Johnson (1996).

From the velocity measurements, we can estimate how deep the effects of the SSFs are “felt,” which we take to be the depth at which the across-front velocity within the warm layer is compensated by return flow beneath the warm layer. We assume continuity and calculate the depth at which the difference between before and after across-front velocities integrates vertically to zero. In other words, this is the depth where volume is conserved for 2D circulation. In 2008, this depth was 103 m, and in 2014, this depth was 67 m. These depths correspond closely to the depths of the EUC core (i.e., maximum zonal velocity; Figs. 9b,h), distinct peaks in \( 4N^2 \) (Figs. 9e,k), and a 100-fold drop in dissipation rates (Figs. 4g,n).

c. Interpretation as gravity currents

Gravity currents are driven by the across-front pressure gradient between two fluids of different densities.
The main force acting on such flows is reduced gravity, which is distributed unequally between the two adjacent fluids and results in the descent of the denser fluid \( \rho_0 \) and the lateral spread of the buoyant fluid \( \rho_c \) along the surface, drawn schematically in Fig. 1c. Thus, the difference in potential energy between the gravity current and the ambient fluid \( [PE = (\rho_0 - \rho_c)g(h/2)] \) is converted to kinetic energy \( [KE = (1/2)\rho_c c_0^2] \), where \( g \) is gravity and \( h \) is the thickness of the gravity current. The upper limit of the speed of this current is achieved when there is

![Figure 8](image.png)

**Fig. 8.** (a),(d) ADCP velocity in the across-front direction \( u' = u'_\text{obs} - u'_\text{back} \) and (b),(e) alongfront direction \( v' = v'_\text{obs} - v'_\text{back} \). (c),(f) Across-front velocity with the frontal propagation speed removed \( u' - u'_\text{front} \), that is, flow around and within a stationary warm layer. Streamlines, calculated by integrating \( u' - u'_\text{front} \) down from the surface, are shown with dashed lines with the zero streamline in bold. Isopycnals (solid black lines) are shown in all subplots.

![Figure 9](image.png)

**Fig. 9.** Before and after profiles of (a),(g) density, (b),(h) zonal velocity, (c),(i) across- and (d),(j) alongfront velocities as defined by Eq. (5), (e),(k) buoyancy frequency, and (f),(l) squared shear for the (top) 2 Nov 2008 and (bottom) 24 Nov 2014 fronts. Temporal averages are taken over −4 to 0 km before and from 2 to 6 km after the fronts’ arrivals. Depth of the warm layer \( h \) plotted as blue horizontal lines.
perfect energy transfer PE = KE, yielding \( c_0 = \sqrt{g' h} \) (Simpson 1997; Ungarish 2009), where \( g' = g(\rho_0 - \rho_c)/\rho_c \) is the reduced gravity.

Averages of density, speed, buoyancy frequency, and shear are calculated over \(-4\) to \(2\) km before the front and \(2\) to \(6\) km after the front (Fig. 9), which avoids the region where the frontal interface is rapidly deepening.

We take \( h \) to be the average depth of isopycnals that outcrop at the front once they deepen to a near-constant depth: \( h = 37.9 \pm 4.4\) m in 2008 using the 1023.57–1023.67 kg m\(^{-3}\) isopycnals and \( h = 28.3 \pm 3.7\) m in 2014 using the 1023.40–1023.50 kg m\(^{-3}\) isopycnals. (See Table 1 for a summary comparison.) These estimates of \( h \) are close to the depths where \( u' \) were equal and where there were local maxima of \( S^2 \) and \( 4N^2 \) (Fig. 9). Before and after densities are averaged from \(0\) to \(h\) to get \( \rho_0 \) and \( \rho_c \); in 2008, \( \rho_0 = 1023.73 \pm 0.03\) kg m\(^{-3}\) and \( \rho_c = 1023.45 \pm 0.08\) kg m\(^{-3}\), and in 2014, \( \rho_0 = 1023.53 \pm 0.02\) kg m\(^{-3}\) and \( \rho_c = 1023.36 \pm 0.04\) kg m\(^{-3}\). These values yield \( c_0 = 0.31 \pm 0.08\) m s\(^{-1}\) in 2008 and \( c_0 = 0.24 \pm 0.06\) m s\(^{-1}\) in 2014 (Table 1), which are larger than \( u'_{\text{front}} \) in both cases. The ratio of measured front speed to \( c_0 \) can be considered as a Froude number \( Fr = u'_{\text{front}}/c_0 \). In both cases, \( Fr < 1 \) (\( \approx 0.7\)), suggesting imperfect energy transfer from PE to KE. One possibility is that significant dissipative losses occur due to turbulence, as suggested in section 4d. For the front observed on the leading edge of a TTW by Johnson (1996), \( Fr \) was estimated to be 0.72, albeit using the unstratified gravity current model of Benjamin (1968).

Alternately, instead of comparing \( u'_f \) to \( c_0 \), we compare to the intrinsic first baroclinic mode speed of internal waves propagating in the ambient fluid \( c \). For this purpose, we use vertical profiles of \( N^2 \) and the eigenvalue technique described by Chelton et al. (1998). From the full water column profile of stratification to 3700 m, \( c \) is found to be 2.5 m s\(^{-1}\). If, however, we assume that the strongly stratified and fast-flowing EUC acts as a lower barrier to the surface-trapped current (103 m in 2008 and 67 m in 2014, discussed in section 4b), we find \( c \) in the surface layer of 0.33 m s\(^{-1}\) in 2008 and 0.20 m s\(^{-1}\) in 2014. These first-mode internal wave speeds are similar to \( c_0 \) and yield comparable values of \( Fr \).

d. The life-span and dissipation of the observed SSFs

To estimate how long it might take for the observed SSFs to dissipate, we calculate a spindown time \( \tau \) for the fronts from the ratio of their energy to \( \varepsilon \). Here, we assume that the SSFs lose energy solely to turbulence. They may also lose energy through radiation of internal waves (e.g., Tanaka et al. 2015), which would further reduce the spindown time scales calculated here. We isolate the energetics of the front by calculating KE and available potential energy (APE) from differences in velocity and density between the warm layer behind the front and the ambient fluid ahead of the front. The total available energy (TE) contained within the front is therefore

\[
\begin{align*}
\text{TE} &= \int_{-h}^{0} \left( \frac{\rho_A(z)}{\rho_A} \left[ u'_A(z) - u'_B(z) \right]^2 + \frac{1}{2} \rho_B \left[ \rho_A(z) - \rho_B(z) \right] \right) \, dz, \\
&\quad \text{KE in across-front dir.} \\
&\quad - \frac{g z (\rho_A(z) - \rho_B(z))}{\eta_A - \eta_B} \, dz, \\
&\quad \text{baroclinic part of APE} \\
&\quad + \rho_0 g (\eta_A - \eta_B) \, dz, \\
&\quad \text{barotropic part of APE}
\end{align*}
\]

where \( u'(z), u'(z), \rho(z), \) and \( \eta \) are the horizontally averaged across-front velocities, alongfront velocities, densities, and sea surface heights, with subscripts indicating the values before \( (B) \) and after \( (A) \) the front. The vertical integral is taken from the surface to the depth of the warm layer \( z = -h \). Horizontal averages are taken from \( \pm 0.5 \) to \( \pm 6 \) km before and after the front. The sea surface heights are found by assuming a depth of no motion at \( H \). We presume that KE in the alongfront direction is independent of APE since the Coriolis force is too small for geostrophy to hold.

The TE contained within the warm layer is 6687 J m\(^{-2}\) for the 2008 front of which 1754 J m\(^{-2}\) is APE, 2232 J m\(^{-2}\) is KE in the across-front direction, and 2701 J m\(^{-2}\) is KE in the alongfront direction (summarized in Table 2). In 2014, TE is 1704 J m\(^{-2}\), of which 1160 J m\(^{-2}\) is APE, and the KE in the across- and alongfront directions are 265 and 279 J m\(^{-2}\). We calculate \( \tau \) for the SSFs by dividing TE by the product of \( h \) and \( \varepsilon \) and find that \( \tau \) is 11 h in 2008 and 4 h in 2014. These spindown times are short, suggesting that the SSFs are strongly damped, particularly given that we have neglected energy loss via wave radiation. However, this calculation also assumes there is nothing supplying energy to the front while it is dissipating. It is possible that
the TIW continues to sustain the SSF through an advective flux of energy.

5. Model perspective of equatorial front evolution

To more thoroughly understand the generation and propagation mechanisms of SSFs, we turn to a numerical model. It should be noted that the resolution of this model (~6 km) is not sufficiently fine to model the observed SSFs that change on scales of meters. Instead, we use the model as a guide for understanding phenomena similar to our observations.

a. TIW-driven frontogenesis

TIWs exert horizontal strain on the north equatorial front driving both frontogenesis (frontal intensification) and frontolysis (frontal weakening). Frontogenesis occurs when the strain in the velocity field compresses the horizontal distance between lines of constant buoyancy $b$, and frontolysis occurs when this distance expands. Frontogenesis can be quantified by considering the change in magnitude of the horizontal buoyancy gradient $|\nabla_h b|^2$ caused by the advection of buoyancy by horizontal $u_h = (u, v)$ and vertical $w$ velocities. The evolution of $|\nabla_h b|^2$ following a fluid parcel is known as the frontogenesis function (Miller 1948; Hoskins 1982):

$$
\frac{D|\nabla_h b|^2}{Dt} = -2\nabla_h b \cdot \frac{\partial u_h}{\partial x} \nabla_h b, \quad \frac{\partial u_h}{\partial y} \nabla_h b, \quad \frac{\partial u_h}{\partial y} \nabla_h b
$$

$$
\times -2\nabla_h b \cdot (N^2 \nabla_h w), \quad (7)
$$

where $D/Dt$ is the material derivative, and $\nabla_h$ is the horizontal gradient. The magnitude of the horizontal buoyancy gradient $|\nabla_h b|^2$ can change because of the horizontal strain [HS in Eq. (7)], differential vertical advection in a stratified fluid [DVADV in Eq. (7)], and diabatic processes, which have been neglected. Regions where the horizontal strain increases $|\nabla_h b|^2$ (i.e., HS > 0) are referred to as frontogenetic regions, and regions where HS < 0 are referred to as frontolytic.

There are a number of frontogenetic and frontolytic regions along the north equatorial front (Fig. 10). The strongest frontogenesis occurs along the leading edge fronts where the SEC curves northward into the NECC (e.g., near 6°N, 144°W, 135°W, and 124°W in Figs. 10a,b). However, there is also frontogenesis along the trailing edge fronts (e.g., near 2°N, 128–134°W in the black box in Figs. 10a,b). Northeast of this front, the westward flow of the SEC is strengthened in the southern part of a TIV and strains the front against the relatively weak northward flow to the west (velocity vectors in Figs. 10a and 10c and shown conceptually in Fig. 1).

Along the trailing edge front, frontogenesis is often sporadic with frontolysis and frontogenesis alternating in time and space. On a number of occasions, a thin front with high HS appears to detach from the trailing edge front at its eastern end and propagate toward the equator (e.g., compare Figs. 10a,b with Figs. 10c,d near 132°W within the black boxes). In the following section we examine the frontal dynamics of the trailing edge front during one example of this kind of event and show that the detached portion behaves as a SSF.

b. Gravity current formation

Here, we examine the behavior of a strong trailing edge front in the southern portion of a TIV (Fig. 11). Initially, the front is characterized by strong alongfront velocities with a change in alongfront velocity of more than 1 m s$^{-1}$ across the frontal outcrop (Fig. 11c). Convergence in the across-front velocity at the surface drives frontogenesis and is accompanied by a downward vertical velocity (Fig. 11c). This front shares similarities with submesoscale midlatitude fronts, with stronger downward motion on the cold side of the front than upward motion on the warm side (Shcherbina et al. 2013). As time progresses, the front moves toward the equator (Fig. 11e), and turns to orient northwest–southeast. The across-front velocity strengthens and the alongfront velocity weakens (cf. Figs. 11c and 11g). By yearday 314, the buoyancy gradient is focused at the frontal outcrop and is followed by a 50-m-thick warm layer that appears to flow directly toward the frontal outcrop (Fig. 11f). The front now looks like a gravity current with many features in common with the observed SSFs. The change in surface temperature across the front is 2°C in 15 km.

To examine the dynamics acting as the SSF moves toward the equator, we consider the following across-front force balance:
In all subpanels, the black box highlights the region where an SSF was released from the trailing edge front.

Fig. 10. (a),(c) The magnitude of the horizontal buoyancy gradient \( \log_{10}(\nabla b) \), 1°C SST contours, and surface velocity vectors from the 6-km horizontal resolution ROMS simulations. (b),(d) The frontogenesis function \( HS [\text{Eq. (7)}] \). Fields are shown on yearday (left) 309 and (right) 315. In all subpanels, the black box highlights the region where an SSF was released from the trailing edge front.

\[
\frac{\partial \mathbf{u}'}{\partial t} + \frac{\partial \mathbf{u}'}{\partial z} = -\frac{1}{\rho_0} \frac{\partial P}{\partial x} + f \mathbf{u}' + \text{Proj}_{k=\mathbf{u}}(\mathbf{u}' \cdot \nabla \mathbf{u}') \cdot \hat{z} + \text{Other terms},
\]

where \( \mathbf{u}' = u' \hat{x} + v' \hat{y} + w' \hat{z} \) defines the across-front, alongfront, and vertical velocities in the \((x', y', z)\) directions, and \( P \) is pressure. On the left-hand side is the advection of across-front momentum by itself and by the vertical velocity, calculated in the frame of the moving front. On the right-hand side, the Coriolis force involves the Coriolis parameter \( f = 2\Omega \sin \theta \), where \( \Omega \) is the rotation rate of Earth and \( \theta \) is latitude. Also included are the pressure gradient force and the across-front component of the centrifugal force or the component of momentum advection perpendicular to the horizontal velocity (Holmes et al. 2014). The projection operator is defined as \( \text{Proj}_b(a) = (a \cdot b) \hat{b} \), where \( b = b/|b| \). Each term is calculated from snapshots of the model state variables once per day using finite differences. The pressure gradient force is reproduced using the model’s online density Jacobian method. The residual includes processes such as friction, acceleration in the frontal frame, and other advection terms. Averages are taken over different across-front distances on yeardays 309 and 314 because of the compression of across-front scales (Figs. 11c.g).

On yearday 309, the front is in cyclogeostrophic balance: pressure gradient + centrifugal \( \approx \) Coriolis (Fig. 11d). As the front moves toward the equator (from a mean latitude of 1.6°N on day 309 to 0.7°N on day 314), the Coriolis force weakens, breaking the cyclogeostrophic balance. The centrifugal and pressure gradient forces then drive the front to accelerate forward. On yearday 314, both centrifugal and Coriolis forces within the front are small (Fig. 11h). The barotropic component of the pressure gradient force is also significantly smaller (surface value of the pressure gradient is \( \sim 1 \times 10^{-5} \text{ m s}^{-2} \) in Fig. 11d vs \( \sim 0.5 \times 10^{-5} \text{ m s}^{-2} \) in Fig. 11h) as a result of changes in the sea surface height gradient across the front. However, the baroclinic component of the pressure gradient is large on yearday 314 and works to accelerate the ambient water under the base of the warm layer. This acceleration manifests in the advective terms (blue line in Fig. 11h), suggesting that the front is no longer in cyclogeostrophic balance and has dynamics consistent with a gravity current.

This force balance is similar to that diagnosed by Johnson (1996, his Fig. 17) along a leading edge front. He showed that at the surface, the sum of advection (1.5 \( \times \) 10^{-5} m s^{-2}) and Coriolis force (0.4 \( \times \) 10^{-5} m s^{-2}) balanced the pressure gradient force (\( -1.9 \times 10^{-5} \text{ m s}^{-2} \)). At depth, between \( \sim 30 \text{ m} \) and the chosen reference depth of 100 m, the advection and pressure gradient changed sign, indicating acceleration of the ambient water beneath the warm layer, as found in our model. The biggest difference between our model’s force balance and that of Johnson (1996) is the ratio in size between forces at the surface and those at depth. In the model, the magnitude of the pressure gradient force is about twice as big at 80 m than it is at the surface, whereas in Johnson (1996) the pressure gradient force is about twice as big at the surface than at 80 m. One reason for this could be the 100-m reference depth chosen.
by Johnson (1996). The model would indicate that a depth of no motion should be deeper than 100 m. If Johnson (1996) had chosen 140 m as a reference depth, then the surface value of the pressure gradient force would indeed be twice as big at 80 m than at the surface. Alternately, the model has a finite vertical resolution that may limit its ability to resolve sharp vertical gradients in the density field and allow the pressure gradient signature of the front to extend deeper in the water column.

c. Gravity current propagation

The force balance in the previous section suggests that the SSF in the model may indeed be a gravity current. This conclusion is further supported by comparing the propagation speed of the SSF to \( c_0 \).

To calculate \( c_0 \) for the modeled SSF, we estimate that \( h \) ranges between 35- and 70-m depth, as by these depths the across-front temperature gradient changes sign (Figs. 11b,f). We take \( \rho_0 \) and \( \rho_c \) as the average surface densities on the cold and warm sides of the front, respectively, out to distances between 15 and 80 km from the frontal outcrop, with the spread in results between these two values and across the range of \( h \) used to estimate the uncertainty of \( c_0 \) in Fig. 12.

The speed of the front is estimated from Eq. (4). The frontal speed in Earth’s reference frame \( u_{f0} \) is found by tracking the position of the maximum horizontal buoyancy gradient and taking the component of the motion perpendicular to the frontal line. The background flow speed \( u_{bgm} \) is taken as the average across-front velocity \( \pm 80 \) km on either side of the front from 0 to \( h \).
The difference between these is the frontal propagation speed in the reference frame of the moving fluid $u'_{\text{front}}$. Figure 12 shows an acceleration of $u'_{\text{front}}$ between yeardays 309 and 312, consistent with the loss of clogeostrophic balance discussed in section 5b. After this acceleration, $u'_{\text{front}}$ compares well with $c_0$, suggesting that the front has indeed transformed into a gravity current. These results are robust to the wide range of thicknesses and widths used to calculate $c_0$.

**d. Comparison to satellite SST**

Similar to the fronts found in the model, features that detach from the TIW trailing edge front are detected in satellite SST (to the northeast of 0°, 140°W in Fig. 13). In both the model and satellite, the trailing edge front has a sharp leading edge (Figs. 10a,b, 13a,b) that a few days later is released and propagates to the southwest (Figs. 10c,d, 13c,d). The Moderate Resolution Imaging Spectroradiometer (MODIS) satellite (Werdell et al. 2013) has a resolution of 4.6 km, which is similar to the model’s ~6-km horizontal resolution; therefore, we presume we are detecting fronts of similar sharpness. Theoretically, propagation speeds could be calculated for the fronts detected in the satellite SST (Fig. 13) and compared to those from the model. However, clouds prevent detection of the fronts at enough intervals for both the 2008 (shown in Fig. 13) and 2014 fronts.

**6. Prevalence of sharp fronts in the equatorial Pacific**

In this section, we quantify the prevalence of fronts over the entire eastern equatorial Pacific. We used near-surface (1-m depth) temperature at 10-min temporal resolution from the TAO moorings (McPhaden et al. 1998) to quantify both the frequency and spatial extent of fronts. All of the TAO moorings between 165°E and 95°W from 8°S to 8°N were used. A comparison to the Optimum Interpolation SST (OISST) dataset (Reynolds et al. 2007; Banzon et al. 2016) allows us to examine how fronts fit within the spatial structure of TIWs. Throughout this section we use the term “front” to describe any front that encountered the TAO array since we cannot diagnose whether these fronts are dynamically similar to the SSFs discussed above.

Fronts were identified from the temperature time series using the following three criteria:

1) Change of $T$ by at least 0.2°C between consecutive data points separated by 10 min; alternatively, $|T'(t_0)| \geq 3.33 \times 10^{-4}$ K s$^{-1}$, where subscript $t$ represents a time derivative, and $t_0$ is the time that the front occurred. This criterion ensures that we identify features with magnitudes and propagation speeds similar to or greater than our two observed SSFs. For instance, when the 2008 fronts crossed the TAO mooring at 0°, 140°W, $T$ increased by 0.27°C between consecutive measurements, which is the largest value of $T_t$ at that location over a 3-week period (Fig. 14).

2) The 48-h-averaged temperature following the front is at least 0.3°C warmer (or cooler) than the 48-h-averaged temperature before the front (gray lines show daily averages in Fig. 14). These criteria ensure that the change in temperature was long lasting and not related to the diurnal temperature cycle.

3) We required that there were no decreases (or increases) in $T$ with magnitudes greater than 30% of $-T'(t_0)$ within 1 h before or after the observed front. This criterion was necessary to remove instances where temperature spiked and then quickly returned to its original value one or two data points later.

These criteria were tuned to be sure the algorithm identified fronts but did not count false positives. By visually inspecting time series from 0°, 140°W and 2°N, 110°W, it was found that the algorithm produced at most a 10% error in the number of fronts and false positives identified.

The total number of fronts found at each location is shown in Fig. 15, with each line normalized by the number of days of good observations at each location. As expected, the greatest number of fronts was found in the eastern cold tongue, 95° to 140°W, and slightly north of the equator. Almost no fronts were found south of the equator except at 95°W. At most locations, fronts were most common during August through January when TIW energy is typically greatest.
In Fig. 15, fronts are categorized by whether SST increased ($T_i > 0$; red bars) or decreased ($T_i < 0$; blue bars) as they passed the stationary TAO moorings. The fact that both warm-to-cold and cold-to-warm transitions are observed is because the TAO moorings observe temperature changes that are controlled by advection in Earth’s reference frame $u_{fo}$ in Eq. (4). In the absence of background flow, buoyant gravity currents will propagate warm water over colder ambient water and appear as $T_i > 0$ in the TAO SST time series [i.e., $u_{bkgnd} = 0$ and $u_{front} = u_{fo}$ in Eq. (4)]. However, if $u_{bkgnd}$ advects the front in the opposite direction of $u_{front}$, it can appear as $T_i < 0$ in the TAO SST time series. Fronts with $T_i < 0$ occur most frequently at $5^\circ$N between 140$^\circ$W and 110$^\circ$W, where the TIW flow drives leading edge fronts northwestward. For instance, the front observed by Johnson (1996) along a leading edge front at $2.1^\circ$N, 140$^\circ$W was observed to propagate westward in Earth’s reference frame even though the gravity current in the frame of the ambient fluid was propagating southeastward.

Figure 15 suggests that we were lucky to have observed two sharp fronts. At 0$^\circ$, 140$^\circ$W, on average only two sharp warm fronts are observed in the TAO data each year. In fact, only two fronts were observed at this location in 2008: on 29 August 2008 with $\Delta T = 0.31^\circ$C and on 2 November 2008 with $\Delta T = 0.27^\circ$C. Both occurred to the south of a trailing edge front (not shown). At 0$^\circ$, 110$^\circ$W and 2$^\circ$N, 110$^\circ$W, fronts are more common, occurring 8.5 and 6.2 times per year on average, respectively. No obvious signal from the 2014 observed front was found in the nearest TAO data at 0$^\circ$, 110$^\circ$W or 2$^\circ$N, 110$^\circ$W, suggesting that many of these features may be quite localized, unlike the Yoder et al. (1994) leading edge front and the modeled fronts in this study (Fig. 11c), which span many hundreds of kilometers.

It is unknown why some trailing edge fronts sharpen significantly when they approach the equator and others...
do not. It can be seen in Fig. 14 that the 2008 front occurred near the start of a 10-day period from 1 to 10 November 2008, where SST was elevated corresponding to the passage of a TIW warm trough at $0^\circ$, $140^\circ$W, as seen in TAO SST and OISST. However, sometimes the OISST shows a trailing edge front approaching $0^\circ$, $140^\circ$W, and the TAO temperature rises gradually over a $\sim$10-day period, but no sharp fronts are detected. The model results suggest that frontogenesis is quite variable in this region, depending on the mesoscale and submesoscale variability driving the strain field. Therefore, it is likely that sharp fronts form and dissipate between moorings without being counted using our metric.

7. Conclusions

We have described two very sharp fronts observed in the equatorial Pacific on 2 November 2008 at $0^\circ$, $140^\circ$W and on 24 November 2014 at $0.75^\circ$N, $110^\circ$W. A temperature increase of $>0.4^\circ$C accompanied by a decrease in salinity occurred over less than 1 m lateral distance in both cases. The water on the two sides of the fronts differed in biological makeup with higher levels of 880-nm optical backscatter and fluorescence found on the cold side. Both fronts propagated to the southwest with speeds relative to the background flow of 0.21 m s$^{-1}$ in 2008 and 0.17 m s$^{-1}$ in 2014. Comparisons to $c_0$ showed that these SSFs had Froude numbers of $\sim$0.7, suggesting that they were propagating as subcritical gravity currents. It is possible that these gravity currents were still transitioning away from cyclogeostrophic balance and had not achieved a full gravity current force balance when observed.

We measured turbulent kinetic energy dissipation rates directly with the Chameleon microstructure profiler. Within the warm layer, observed dissipation rates reached values as high as $10^{-4}$ W kg$^{-1}$, and the average dissipation, $4.5 \times 10^{-6}$ W kg$^{-1}$, was 1000 times greater than on the cold side of the fronts. The total energy content of the 2008 and 2014 fronts was 6687 and 1704 J m$^{-2}$, respectively, with much higher kinetic energy in 2008 accounting for the bulk of the difference between the 2 years. A spindown time for the fronts was calculated, assuming that they dissipate via turbulent mixing, from the ratio of energy to dissipation found to be short (11 h in 2008 and 4 h in 2014). This indicates that the SSFs were strongly dissipative and that, if they were to exist for longer, they must be sustained by the TIWs through a continuous advective flux of energy.

The fronts were characterized by remarkably similar temperature jumps, dissipation rates, and propagation characteristics, yet there were a few key differences. In 2008, the warm layer was thicker ($38 \text{ vs } 28 \text{ m}$) with stronger velocities within the warm layer than in 2014. This is reflected in the respective spindown times and ratios of APE to KE. Possibly these differences were due to the latitude at which the two fronts were observed ($0^\circ$ in 2008 and $0.75^\circ$N in 2014) or the overall energy of the TIW field caused by the difference in ENSO state (ONI negative in 2008 and positive in 2014). Or it may simply be due to variability in the flow at sub-TIW scales.

A numerical model of the equatorial Pacific with 6-km grid resolution suggests that these fronts may have formed via TIW strain-induced frontogenesis from trailing edge fronts initially in cyclogeostrophic balance (e.g., Fig. 11d). In one example, a trailing edge front propagates toward the equator resulting in a reduction in the Coriolis parameter $f$. The reduction in the Coriolis force allows the centrifugal and pressure gradient forces to accelerate the across-front flow and increase the propagation speed of the front. The across-front advective accelerations $\frac{\partial u}{\partial x}$ and $\frac{\partial w}{\partial z}$ then play a more important role in the frontal force balance (Fig. 11h), suggesting that the feature has transformed into a gravity current. The propagation speed of the front agrees with theoretical estimates for the propagation speed of a gravity current (Fig. 12), adding further support to our hypothesis.

In Fig. 1a, we show the SSF as a detached feature that is separate from the trailing edge front. However, it may...
be more accurate to consider the SSF as the sharpened leading edge of the trailing edge front. As the SSF accelerates away from a trailing edge front as a gravity current, presumably it draws with it warm water from the trailing edge front, keeping the two features connected. Conversely, it cannot be said that every trailing edge front is an SSF. In section 6, we show that there are many trailing edge fronts that do not sharpen to the point of forming an SSF. Therefore, natural variability of the TIW cold cusp shape and the strength of frontogenesis along the trailing edge front must both play important roles in the creation of SSFs.

The dynamics of SSFs that form along the trailing edge front likely differ from those created along the leading edge front. Because of the strong, persistent convergence along the leading edge front, and their more northern latitude where the Coriolis parameter is larger, it may be more difficult for SSFs to detach from a leading edge front than from a trailing edge front. For instance, the leading edge front observed by Yoder et al. (1994) and Johnson (1996) was remarkably straight and stable, even though the force balance indicated that it was a gravity current.

This study represents one of the few direct observations of the small-scale dynamics associated with TIWs. The SSFs studied here constitute a mechanism for the downscaling of energy from the large, predominantly cyclogeostrophically balanced scales of the TIWs to the small scales at which energy dissipation occurs. Furthermore, the intense, near-surface turbulence and injection of warm surface water that accompanies the SSFs may impact the flux of heat across the air–sea interface within the equatorial cold tongue (Moum et al. 2009) and whether that heat penetrates deeper to warm the ocean interior. The downwelling of ambient water below the warm layer also has biochemical implications, as suggested by the observed accumulation of fluorescence along the front similar to that seen by Yoder et al. (1994).

A future observational study to track an SSF from its generation to its eventual dissipation would shed light on the alongfront variability of these features and the evolution of these sharp fronts as they propagate and dissipate. However, given the statistical occurrence rate of these features from TAO analysis, it might be difficult to locate. As observational techniques improve through the use of higher spatial and temporal resolution satellite sensors and high-speed thermistors on moorings and ships, oceanographers are finding that sharp fronts are more prevalent in the ocean than was once thought. This paper gives insight into the generation and propagation of sharp fronts in the equatorial Pacific, which may turn out to have similarities to small-scale fronts elsewhere.

Acknowledgments. This work was funded by the National Science Foundation (1256620). We acknowledge the support of NOAA in helping us to maintain χpod measurements on the TAO moorings and R.C. Lien of APL/UW for providing EQUIX mooring data. The numerical modeling computations were performed at the Stanford Center for Computational Earth and Environmental Science. R. Holmes was supported by a Robert and Marvel Kirby Stanford Graduate Fellowship while undertaking this study. TMI data were produced by Remote Sensing Systems and sponsored by the NASA Earth Sciences Program. (Data are available at http://www.remss.com/missions/tmi/.)

APPENDIX

Calculating the Front Propagation Velocity Using the Method of Least Squared Error

In section 4a, we showed that the velocity of the front traveling through an observation array can be calculated from

$$
\Delta t_k + \Delta \tau_k = \frac{\Delta x_k \cdot u_{k0}}{u_{k0} \cdot u_{k0}},
$$

(A1)

where the subscript $k$ denotes location pairs with values of $k = 1, \ldots, K$, where $K = N(N-1)/2$ and $N$ represents the total number of measurements. In Eq. (A1), $\Delta t_k$ is the difference in arrival times between location pairs, $\Delta x_k = (\Delta x_k, \Delta y_k)$ is the distance between location pairs, and $u_{k0} = (u_{x0}, u_{y0})$ is the estimated velocity of the front. An additional error term $\Delta \tau$ has been added to allow a least squares error with multivariate regression to be computed following the method described by Emery and Thomson (2004, 239–242).

First, create matrices that include all location, time, and error pairs:

$$
X = \begin{pmatrix}
\Delta x_{k=1} & \Delta y_{k=1} \\
\Delta x_{k=2} & \Delta y_{k=2} \\
\vdots & \vdots \\
\Delta x_{k=K} & \Delta y_{k=K}
\end{pmatrix},
$$

$$
T = \begin{pmatrix}
\Delta t_{k=1} \\
\Delta t_{k=2} \\
\vdots \\
\Delta t_{k=K}
\end{pmatrix},
$$

$$
E = \begin{pmatrix}
\Delta \tau_{k=1} \\
\Delta \tau_{k=2} \\
\vdots \\
\Delta \tau_{k=K}
\end{pmatrix}
$$

(A2)
For simplicity, we introduce a reduced velocity with units of seconds per meter:

\[
U = \frac{u_s - u_0}{u_s}.
\]  

(A3)

Combining Eqs. (A2) and (A3) allows Eq. (A1) to be written as a matrix equation:

\[
T = X \cdot U + E,
\]  

(A4)

which, according to Emery and Thomson (2004), has a least squares error equation of

\[
(X^T \cdot X) \cdot \hat{U} = X^T \cdot T
\]  

(A5)

and a solution

\[
U = (X^T \cdot X)^{-1} X^T \cdot T.
\]  

(A6)

where \(X^T\) is the transpose of \(X\), and \((X^T \cdot X)^{-1}\) is the inverse of \(X^T \cdot X\). The velocity of the front can then be calculated from the reduced velocity as

\[
u_s = \frac{U}{U_0}.
\]  

(A7)

Provided there are more than three measurements—which was the case in 2014 but not in 2008—the standard deviation \(\sigma\) of the error can be calculated as

\[
\sigma = \sqrt{\frac{T^T \cdot T - (X^T \cdot X) \cdot T}{K - 3}}.
\]  

(A8)

Note that for the error calculation in 2014, instead of using the total number of location pairs, we instead use the number of radar images as an estimate for our independent measurements of the front locations.

REFERENCES


