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3 Properties of an Upper Ocean Front Associated with Water Mass Boundaries at the
4 Entrance to the Gulf of California, November 2004
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ABSTRACT

The structure of an upper ocean front at the entrance to the Gulf of California is described. The front occurred in a region of strong cyclonic shear between fresh Tropical Pacific waters, which flowed into the Gulf along Sinaloa, and salty Gulf of California waters, which flowed out of the Gulf along Baja California. Observations included two high-resolution hydrographic sections across the entrance to the Gulf. One section used a towed CTD that profiled from near surface to 125 m. A second across-Gulf section was obtained using CTDs spaced about every 10 km. Both sections indicated a nearly vertical salinity front near 108.6°W, which extended to a depth of 120 m at the northern front. The surface salinity change across the front was $\Delta S \sim 0.4$ and the width of the front 10-15 km. The surface horizontal mixing ratio, -0.15 to -0.7, indicated that salinity dominated the density change. The mixed layer was strongly developed to a depth of ~ 45 m. Features that were observed at the front included a surface jet (10 km wide, 0.25 m s^{-1}), subduction from the mixed layer to the top of the pycnocline, evidence of symmetric instability driven by atmospheric cooling of surface waters, and a deformation rate of $-0.5 \times 10^{-5} \text{ s}^{-1}$. Northward ageostrophic currents in the upper 60 m extended inflow 55 km to the west of the surface front. Immediately below the bottom of the mixed layer, the seasonal pycnocline was characterized by a 15-m thick and 50-km wide salinity maximum (minimum), $S=34.9-35$ ($S=34.3-34.4$) for the outflow (inflow). These features were not distinct in the frontal zone. Temperature flux and diffusivity were estimated to be about $7.6 \times 10^{-2} \text{ }^\circ\text{C m s}^{-1}$ and $8600 \text{ m}^2 \text{ s}^{-1}$, respectively.

1. Introduction

The entrance to the Gulf of California (GC) marks the locus of exchange between waters of the Gulf and the adjacent Pacific Ocean (PO), which are critical for heat and salt balances in the Gulf. Here, waters from the Gulf lie adjacent to, co-mingle with, and mix with waters from the Pacific Ocean. The boundaries between the surface waters often form sharp fronts which are easily seen from space and are persistent along the continental slope as well as across the entrance to the Gulf, the latter marking where Gulf and Pacific waters meet (Belkin, 2005; Belkin and Cornillon, 2007, Belkin et al., 2009).

In late fall, large scale ocean conditions favor frontogenesis. Strong horizontal shear exists between inflowing Pacific and outflowing Gulf waters. These currents form a cyclonic circulation of narrow (20-30 km), deep flows (0.1 m s^{-1} at 1000 m) on either side of the mouth of the Gulf. They are baroclinic in structure and easily seen in high resolution (10-km) CTD sections (Roden, 1972) or direct current measurements (Collins *et al.*, 1997). The exchanges are driven by wind and

buoyancy and result in cooling and freshening of the Gulf (Castro *et al.*, 1994; Berón-Vera and Ripa, 2000). The November exchange is marked by the transport of Tropical Surface Waters (TSW) into the Gulf (Marcarenhas *et al.*, 2004).

Seasonal meteorological conditions also promote frontogenesis at the entrance to the Gulf in late fall. The atmosphere is generally cooler than the ocean in November and the resulting air-sea buoyancy flux contributes to deepening of the mixed layer, convective overturning and symmetric instability (Taylor and Ferrari, 2010). Although northerly winds dominate from late summer through early spring in the Gulf (Parés-Sierra *et al.*, 2003, Marinone *et al.*, 2003), southerly winds occur six percent of time (National Geospatial Intelligence Agency, 1994). These southerly winds provide a down front component of wind stress which transport more dense surface water toward less dense surface water, initiating vertical motions in the upper ocean including symmetric instabilities (Taylor and Ferrari, 2010).

Existing hydrographic data provide evidence of strong currents in fall and associated frontal structures. Roden (1971) studied hydrographic data from late fall of 1969 and suggested that the border between TSW, California Current Water (CCW) and the Gulf of California Water (GCW) was characterized by a front consisting of a strong horizontal salinity gradient off the southwestern tip of the Baja California peninsula. Collins *et al.* (1997) pointed out the presence of a frontal eddy vortex in a satellite image from November 1990 that indicated a front between cold Pacific waters and either warm tropical waters or GCW at the Gulf entrance.

An example of these flow patterns is given by an AVHRR image for November 11, 2004 (Fig. 1). Surface currents measured by a vessel mounted Acoustic Doppler current profiler (VMADCP) during November 15-17, 2004, (described further below) have been superimposed on Fig. 1. Here a tongue of warm waters appeared to extend poleward along the Mexican coast into the Gulf of California. The VMADCP and AVHRR data show that the surface waters circulated cyclonically across the entrance to the Gulf, and left the Gulf along the Baja California Peninsula. Upon exiting the Gulf, the AVHRR data show a tongue of warm water extended poleward along the Pacific coast of Baja California. Pacific Ocean fronts that are suggested by this circulation pattern include boundaries (1) between the Tropical waters carried

by the Mexican Coastal Current and Pacific Waters and (2) between Gulf Waters and Pacific Waters along the Pacific Coast of Baja California. Within the Gulf, a third front, the focus of the paper, was formed at the entrance to the Gulf in a region of strong horizontal shear between currents flowing into and out of the Gulf.

In November 2004, two high resolution hydrographic sections were occupied across the entrance to the Gulf (Fig. 2) that provided a detailed view of the structure of the upper ocean. Note that the entrance to the Gulf lies in Pescadero Basin, which has a northern sill depth of ~ 2000 m and at its southern limit freely communicates with the Pacific Ocean above 3000 m. The two sections met at the 500 m isobath near Sinaloa and were separated by 50 km at their western origins along the Baja California coast. The southern section was obtained from traditional CTD casts spaced 10 km from one another and provided a mesoscale resolving view of the upper 1000 m of the water column. The northern section was sampled continuously to a depth of about 120 m using a CTD mounted on a towed profiler and provided a higher resolution view of the small-scale (1 m vertically, 1 km horizontally) structure of the upper layer. While these sections showed a number of frontal zones, the strongest frontal zone was the one that separated in- and out-flowing waters. The location of this frontal zone is shown by the trapezoidal region along 108.5°W between the two hydrographic sections shown in Figs. 1 and 2.

The purpose of this study is to describe the properties of the upper ocean front observed at this water mass boundary in the context of regional circulation patterns and global estimates of frontal properties (Fedorov, 1986). The paper is organized in the following manner. Observations and methods are described in the second section. Results are described in the third section, proceeding from large scale to mesoscale to sea surface conditions, before smaller scale frontal features are examined in more detail. A summary of frontal conditions is given in the last section.

2. Methods

Measurements and data processing methods are described below. Continuous measurements of sea surface water properties and meteorological

conditions made on both legs are described first. A description of methods used for measurements to a depth of 1000 m on the southern leg are described next, followed by details of methods used for shallow upper ocean measurements on the northern leg. The final section describes methods used to process the CTD data from both legs.

2.1 Meteorological and sea surface measurements.

Meteorological and sea surface conditions were observed every 24 s while the ship was underway. Meteorological measurements (wind speed, air temperature and humidity, long and short-wave radiation) were made from the ship's mast, about 18.9 m above the sea surface. Sea surface conditions were measured using seawater which was pumped from a sea chest located in the ship's bottom at a depth of 3 m to a SeaBird 21 thermosalinograph located in a laboratory about 10 m away. The accuracy (precision) of the conductivity measurements was 0.001 (0.0001) S m⁻¹, and for temperature measurements was 0.01 (0.001) °C. Underway temperature and salinity measurements were compared with corresponding values of sea surface CTD measurements and corrected if persistent offsets or drift were noted.

Wind stress, $\vec{\tau}$ (N m⁻²), was computed with the bulk aerodynamic formulae given by Trenberth *et al.* (1989), $\vec{\tau} = \rho C_D |\vec{W}| W$, where ρ is a constant air density of 1.2 kg m⁻³, $\vec{W}$ is the wind velocity vector, and C_D is the drag coefficient given by .001(0.49+0.065 W) for $W > 10$ m s⁻¹, .000114 for $3 \leq W \leq 10$ m s⁻¹, or 0.001(0.62+1.56 W^{-1}) for $W < 3$ m s⁻¹. Buoyancy flux, B_0 (m² s⁻³), was estimated as $B_0 = -g \alpha Q_0 / \rho C_p + g \beta (E - P) S_0$, where g is gravity, ρ is the density of the sea surface, C_p is specific heat of water, S_0 is surface salinity, α (β) is the thermal expansion (haline contraction) coefficient, Q_0 is the net heat flux (W m⁻²), E (m s⁻¹) is evaporation and P (m s⁻¹) is precipitation (Cronin and Sprintall, 2009). A positive (i.e., downward) buoyancy flux, due to either surface cooling or evaporation, would make the ocean surface more dense.

2.2 Southern Section

The southern section consisted of 20 CTD stations spaced at ~ 10 km intervals between $24^{\circ} 09' \text{ N}$, $107^{\circ} 39' \text{ W}$ and $23^{\circ} 27' \text{ N}$, $109^{\circ} 24' \text{ W}$ (triangles in Fig. 2). The southern section is a repeated section which has been occupied twenty times over a fifteen-year period. (Results were used to determine seasonal thermohaline properties and geostrophic flow [Castro *et al.*, 2000, Mascarenhas *et al.*, 2004]). These stations were occupied between 20:30 UTC November 15 and 03:00 UTC November 17, 2004. CTD observations extended to near bottom or to 1000 m (where water depth permitted). CTD salinities were corrected using Niskin bottle water samples that were processed with a laboratory salinometer. The CTD, a SeaBird 911plus, had accuracy (precision) of 0.0003 (0.00004) S m^{-1} for conductivity, 0.001 (0.0002) $^{\circ}\text{C}$ for temperature, and 0.9 (0.06) dbars for pressure. The CTD was sampled at 24 Hz and data were processed using software provided by the manufacturer.

A 75 kHz ADCP was used in the generally deeper water along the southern line except in the shallower waters that overlie the Sinaloa and BCS shelves, where the 300 kHz ADCP was used as described below for the northern section. The vertical resolution of the 75 kHz ADCP was 16 m and the center of the shallowest depth bin was 35 m. The accuracy of 5 minute ensembles from these instruments was about 0.01 m s^{-1} (Teledyne RD Instruments 2009a, b). Vessel heading was corrected using GPS measurements, and ADCP data were processed using software provided by the University of Hawaii.

2.3 Northern Section.

Profiling CTD measurements were obtained along the northern section primarily at night, beginning (ending) at 00:00 (13:30) UTC on November 15, 2004. (See dashed line on Fig. 2) These measurements began at the 300 m isobath off Baja California Peninsula ($23^{\circ} 49.6' \text{ N}$, $109^{\circ} 39.5' \text{ W}$) and ended near the 500 m isobath on the eastern side of Pescadero Basin ($24^{\circ} 0.8' \text{ N}$, $108^{\circ} 3.2' \text{ W}$). Water properties were measured along this section using a towed vehicle (TRIAXUS, manufactured by MacArtney Underwater Technology) which sampled continuously at 24 Hz while cycling between 3 and 124 dbars as the vessel proceeded at a speed of 3.75 m s^{-1} .

The TRIAXUS incorporated SeaBird modules for measurement of temperature, conductivity, and dissolved oxygen. A Paroscientific Digiquartz® pressure gauge and a Turner SCUFA fluorometer were also mounted on the TRIAXUS. A total of 244 profiles were collected along the 164 km section, about one profile every 0.7 km. TRIAXUS was towed at a distance of 430-450 m behind the ship.

A ship-mounted 300 kHz ADCP was used along this section for observing ocean currents. Data were binned into 5-minute ensembles with spatial resolution somewhat less than that of the TRIAXUS, about 1.2 km, and a vertical resolution of 4 m. Vessel heading was corrected using GPS measurements. ADCP data were processed using software provided by the University of Hawaii.

2.4 CTD Data Processing

CTD data from both legs were processed in a similar manner. Potential temperature (θ , °C), potential density anomaly (γ_θ , kg m⁻³), and other parameters (e.g., thermal expansion and saline contraction coefficients, buoyancy frequency, density ratio, etc.) were computed following UNESCO (1991). The mixed layer depth was estimated for each CTD profile as the depth where the potential density exceeded that at the surface by 0.125 kg m⁻³ (Kara *et al.*, 2000). The Turner angle was used to show regions where double diffusive processes could affect thermohaline stability (Ruddick, 1983) and Richardson Number to identify turbulent regions.

3. Results

Results are discussed from larger to smaller scale and from deep to shallow. The upper 500 m of the southern section are first examined and then the higher resolution data from the northern section, including small- (1-10 m) and medium-scale (10-100 m) vertical structures. Finally, sea surface and mixed layer properties are described.

3.1 Southern Section

In November 2004, strong horizontal shear associated with seasonal cooling of the Gulf provided the basis for frontogenesis (Hoskins and Bretherton, 1972). This was manifested by sharp gradients of water properties and cyclonic horizontal shear which extended throughout the water column and was easily seen in sections of water properties and velocities along the southern section (Fig. 3). The location of the front was most apparent in the upper ocean salinity field shown in Fig. 3a despite the variation of salinity with depth being relatively small at the entrance to the Gulf compared to other ocean locations: salinity varied by only $\Delta S=0.9$. The salinity field was marked by east-west asymmetry in the upper 100 m, with the transition occurring near 108.6°W. The inflowing upper ocean waters off Sinaloa were clearly fresher (minimum salinity 34.4 just beneath the base of the mixed layer) than waters to the west (maximum salinity > 34.9). Although the 10 km station spacing was too large to resolve details of the frontal structure along the southern section, evidence of subduction near the surface front was seen at 108.5°W (the eastern edge of the surface front) which indicated a 3 m thick lense of $S = 34.6$ at the base of the mixed layer.

The upper water column was highly stratified by temperature (Fig. 3b). Between the bottom of the mixed layer (about 26°C and 50 m) and 100 m, the temperature decreased about 10°C. Unlike temperature, there wasn't a simple monotonic change associated with the vertical salinity transition below the mixed layer (Fig. 3a), in part due to the strong stabilizing effect of the vertical temperature gradient. If the flow were in geostrophic equilibrium, a front associated with cyclonic shear between inflowing and outflowing waters should be marked by a doming of isotherms. This is clearly seen in Fig. 3b below about 200 m, where the 8-12°C isotherms were shallowest near 108.5°W, e.g. the 9°C isotherm shoaled from 420 m to 330 m at 108.6°W. In other observations (Collins *et al.*, 1997; Castro *et al.*, 2000; Mascarenhas *et al.*, 2004), this doming was observed 50 km to the west of the location shown here.

The potential density anomaly (Fig. 3c) was similar to the temperature section. The largest density gradient occurred between the bottom of the mixed layer and about 100 m, corresponding to a density change of 23 to 25.4 kg m⁻³. In

the layer above 100 m, there was baroclinic compensation for the deeper flow, e.g. the shape of the shallow pycnocline appeared to be the mirror image of the deeper pycnocline. Shallow upwelling (downwelling) was observed at the Sinaloa (Baja California) coast, consistent with the observed southward winds and westward surface currents near Baja California (Fig. 1).

Dissolved oxygen concentrations confirmed the difference in water characteristics in the upper 100 m (Fig. 3d). In the upper ocean, the inflowing waters above 60 m (potential density $\leq 24 \text{ kg m}^{-3}$) to the east of 108.7°W contained about $20\text{--}30 \text{ }\mu\text{mol kg}^{-1}$ more dissolved oxygen than waters to the west. Maximum dissolved oxygen ($206 \text{ }\mu\text{mol kg}^{-1}$) was associated with the lowest salinity (34.26) water at 54 m, 108.4°W . On density surfaces (not shown), these differences between dissolved oxygen for inflowing and outflowing waters disappeared for potential densities greater than 25 kg m^{-3} (70–90 m). The strong oxycline and low oxygen ($>10 \text{ }\mu\text{mol kg}^{-1}$) below about 250 m are a distinctive characteristic of these waters (Castro *et al.*, 2001).

The observed velocities are shown in the two bottom panels of Fig. 3. The east-west velocity field (Fig. 3e) indicated a region of convergence near 108.3°W (indicated by the zero isotach at depths shallower than 250 m). The north-south velocity field was dominated by the strong shear between the outflow and inflow near this longitude (Fig. 3f). Here the zero isotach rose vertically from 300 m to 70 m above the center of the deep thermal dome, separating the core of inflowing water (peak speed of 0.4 m s^{-1} centered at 150 m) from outflowing water (peak speed -0.2 m s^{-1} also at 150 m). Above 70 m, the zero isotach leveled off, and intersected the mixed layer somewhat eastward at about 108.7°W ; this feature was also seen in geostrophic velocity fields referenced to 1000 dbars (not shown) and was due to the westward shoaling of the shallow pycnocline at the base of the mixed layer. Other features of the north-south velocity field that matched the patterns of density and temperature change along isobars were inflow over the narrow Baja California shelf and upper slope and shallow outflow associated with near surface upwelling over the Sinaloa shelf.

3.2 Northern Section

Data from the northern section provided greater horizontal resolution of the upper ocean hydrographic structure than did the southern section and clearly resolved small-scale frontal structures. Fig. 4 shows the temperature-salinity relationship for the TRIAXUS data. Water mass characteristics have been adopted from Castro *et al.* (2006) and Cepeda-Morales *et al.* (2013) and include Tropical Surface water (TSW), Gulf of California Water (GCW), Pacific Waters (PW) originating in either the California Current (CCW) or Mexican Coastal Current, and Subsurface Subtropical Water (SsStW), each of which is identified in Fig. 4. Fresh Pacific waters entering the Gulf could originate either from southward flow of Subarctic waters in the California Current or by poleward flow of tropical Pacific waters in the Mexican Coastal Current. As discussed above, CCW is typically observed at the Gulf's entrance, but Fig. 1 and historical analyses suggest that tropical surface waters (TSW) can find their way to the Gulf's entrance in autumn (Lavín and Marinone, 2003). Near surface waters (shown at the top of Fig. 4) of GCW and TSW characteristics were mixed immediately to the west of 108.6°W. Between density anomalies of 24.5 to 22.7 kg m⁻³, the water mass structure was bimodal, with TSW/PW to the east and GCW generally to the west; e.g. the upper ocean front was clearly defined on these isopycnal surfaces, with colder and fresher water to the east and warmer and saltier water to the west. At densities greater than 25.5 kg m⁻³, the distinction between eastern and western waters became ambiguous; the salinity of waters next to Sinaloa increased from 34.6 to 34.8 while those next to Baja California ranged from 34.62 to 34.95. The greater variability of salinity along Baja California at these deeper density surfaces could be due to a variety of processes, including east-to-west recirculation in northern Pescadero Basin, variability in Subsurface Subtropical Water characteristics, or occasional intrusions of Pacific Water into the Gulf along Baja California.

Fig. 5 shows high resolution upper ocean sections for the northern section. Note that this section ended at a water depth of about 500 m over the Sinaloa slope so did not resolve inshore water properties. The temperature panel (Fig. 5a) included the mixed layer depth, which is discussed below. As for the southern

section, the front between the western and eastern regions of the northern section was marked by salinity change (Fig. 5b). The 34.7 to 34.9 isolines were nearly vertical above 80 m (where density anomaly was less than 25 kg m^{-3}) at 108.6°W .

Within the higher salinity western area and lower salinity eastern region, extreme values (maxima, minima) of salinity occurred in 10-20 m thick intrusive layers which were located just beneath the bottom of the mixed layer. The fresh intrusion to the east of 108.6°W was marked by $S < 34.4$ and was centered at 50 m along the 23.6 kg m^{-3} density surface. This layer extended to the eastern boundary of Fig. 5b and, using the CTD data shown in Fig. 3a, to the edge of the Sinaloa shelf at 108°W , a distance of 80 km.

The salty intrusion to the west of 108.6°W was marked by $S > 34.9$. This intrusion appeared less intense, less thick, and slightly shallower than the salinity minimum intrusion and was centered along a lighter density surface, 23.2 kg m^{-3} . The salinity maximum could be traced east along the base of the mixed layer to 109.45°W , a distance of about 70 km. This high salinity water was remarkable for its small scale structures. For example, a 2 m thick intrusion of $S = 34.8$ can be traced from 109.45°W to 109.2°W , dividing the salinity maximum into upper and lower halves.

Other features were collocated with the surface front. The temperature (density anomaly) of the mixed layer increased (decreased) from 26°C to 26.5°C (Fig. 5a) (22.8 to 22.4 kg m^{-3} , Fig 5c) to the east of the front. Below 60 m, the zero isotach of the northward flow was located at 108.5°W (Fig. 5f), roughly the same meridian as for the southern section (Fig. 3f). At both the northern and southern sections, the 108.5°W meridian marked a region of strong cyclonic shear between a maximum of inflow ($>0.3 \text{ m s}^{-1}$) to the east and a weaker maximum outflow ($>0.1 \text{ m s}^{-1}$). Strong cyclonic shear also occurred at 109.2°W where the zero isotach of meridional velocity bisected the mixed layer (Fig. 5f). Above 70 m, waters between the surface front at 108.5°W and 109.2°W were flowing northward into the Gulf. At depths above 70 m, weak outflow occurred to the east of 108.4°W over the Sinaloa slope and shelf and stronger outflow ($>0.3 \text{ m s}^{-1}$) next to Baja California.

Fluorometer measurements are shown in Fig. 5d. For each profile, maximum values occurred just beneath the bottom of the mixed layer and were centered between isopycnals 23.5-23.8 kg m⁻³. In this zone of vertical maximum fluorescence, largest values (greater than 2.4 µg l⁻¹) were found on the western side of Pescadero Basin, ~109.3°W. For isopycnals 23.5-23.8 kg m⁻³, minimum fluorescence was observed near the base of the surface frontal zone at about 108.5°W.

The zero isotach of the east-west flow was located at about 109°W and marked a region of surface divergence (Fig. 5e, Fig. 1). Minimum surface temperature (Fig. 5a) and maximum fluorometer values (Fig. 5d) occurred immediately to the east of this divergence. The location of the surface front (108.6°W) was a region of weak eastward flow, $u \sim 0$ m s⁻¹.

To determine the geostrophic flow across the section, the thermal wind was calculated from TRIAXUS measurements of the density field and then added to the smoothed (~ 40 km, see Collins *et al.*, 1997) across section ADCP measurements of velocity at 100 m. The resulting geostrophic currents are shown in Fig. 5g. The character of inflow near the front at 108.6°W and outflow next to Baja California were well represented by the geostrophic flow.

Ageostrophic currents contribute strongly to frontogenesis (Hoskins and Bretherton, 1972, Pallàs-Sanz *et al.*, 2010) and were calculated by subtracting the geostrophic velocities (Fig. 5g) from the observed across section flow (similar to Fig. 5e). Above 50 m and between 109.2° and 108.8°W, ageostrophic currents of ~ 0.2 m s⁻¹ (Fig. 5h) dominated weaker geostrophic flow. This extended inflow associated with the 108.6°W frontal zone to the west.

3.3 Surface and Mixed Layer Properties

Continuous underway data measured at about 3 m depth along the northern and southern sections are shown in Fig. 6. A front divided the cooler, saline and heavier waters in the west from warmer, less saline and lighter waters to the east for both sections (Fig. 6a-c, respectively), consistent with data shown in Figs. 3 and 5. The southern front was observed 28.4 hours later than, 34.2 km south of, and 5.1

km west of the northern front. Surface water with $S \sim 35$ was observed for both north and south sections at 108.7°W , and it decreased to 34.6 at 108.6°W . Temperature also transitioned from 26°C to $26.3\text{--}26.6^\circ\text{C}$ across this longitude band, but the rate of change was much greater for the northern than the southern section.

The surface frontal region was chosen using sea surface salinity change, i.e. the region from the salinity maximum (35) which marked the western side of the front to the minimum (34.6) on the eastern side. Gradients were estimated using a linear fit to the temperature, salinity, and density data shown in Fig. 6. Results are presented in Table 1 and quantify the differences between the northern and southern surface fronts. The width of the surface frontal zone in the southern (northern) section was 14.6 km (9.8 km), and this area is superimposed on Fig. 1 and 2 as a trapezoid. The front decreased its width by a third while the density gradient doubled. The mixing ratio was estimated as $R_\rho = \alpha(\delta T / \delta x) / \beta(\delta S / \delta x)$, where α is the thermal expansion of seawater and β is the haline contraction. For both fronts, the density change associated with salinity variability exceeded that due to temperature change.

Fig. 6c shows the extent to which across-Gulf density changes were dominated by the front near 108.6°W . Other regions with large density changes were located adjacent to the coast. One was associated with upwelled (cooler, saltier) waters over the Sinaloa shelf. Off BC, surface waters warmed 0.3°C .

The east-west and north-south velocity components at 13 m for the northern section are shown in Fig. 6d. Largest velocity changes occurred near 109.8°W and did not appear well correlated with temperature, salinity or density variability. A northward flowing surface jet (expected at a surface front and discussed below) was centered at the eastern edge of the frontal zone (108.6°W) with maximum velocity of 0.3 m s^{-1} . The frontal zone was a region of cross-frontal convergence as indicated by the eastward decrease of the east-west velocity, $\delta u / \delta x = -0.5 \times 10^{-5} \text{ s}^{-1}$, to a minimum, $\sim 0 \text{ m s}^{-1}$, at 108.5°W . For the southern front, east-west velocities at 35 m (not shown) also indicated cross frontal convergence, $-1.3 \times 10^{-5} \text{ s}^{-1}$, to the east of the front. Note that the east-west direction was approximately normal to the front

and along-front changes were small, so $\delta u/\delta x$ corresponded to the deformation rate, D , for the northern front (Fedorov, 1986).

The mixed layer depth (MLD) was obtained for each section (N and S) following the criteria of Kara *et al.* (2000), and was included in temperature, salinity, density, fluorescence, and oxygen panels of Figs. 3 and 5. The MLD fluctuated between 30 and 45 m depth and had temperatures, salinities, and densities of sea surface waters as shown in Fig. 6.

Frontal stability parameters can be estimated for the northern front using these observations. The frontal Rossby number, Ro , was estimated using $Ro = \Delta U/fL \sim 0.3$, where L is the width of the front and was indicative of the geostrophic and semi-geostrophic character of the surface front (Fedorov, 1986). The frontal Rossby number for the northern front can be compared to frontal Rossby numbers for coastal upwelling off Oregon, 0.5, and for the New England shelf/slope water, 0.02 to 0.13 (Flagg and Beardsley, 1978).

The frontal Richardson number (FRN) was estimated using

$$FRN = \frac{gH(\rho_1 - \rho_2)}{\rho|(U_2 - U_1)^2|} = 1.9$$

where the subscript 1 refers to the lighter fluid, H is the mixed layer depth, U to the velocity along the front, and ρ to the density (Fedorov, 1986). This was less than values of 5-10 for upwelling fronts off Oregon (Mooers *et al.*, 1976) or values of 20-120 for the New England shelf/slope water front (Flagg and Beardsley, 1978). Baroclinic and barotropic instability may be expected for fronts with frontal Richardson and Rossby numbers similar to those for the northern front (Fedorov, 1986).

Observed atmospheric forcing conditions are shown in the lower two panels of Fig. 6. Some caution is warranted when examining these data, as spatial and temporal patterns may not be resolved due to ship movement. For example, meteorological conditions over the Sinaloa shelf were visibly affected by the evening buildup of thunder clouds over the western slope of the Sierra Madre, with booming thunder, diffuse flashes of light, and occasional light rain experienced on the ship.

Fig. 6e indicated a maximum wind stress, 0.07 N m^{-2} , near the Sinaloa shelf break at 108.0°W which was associated with weak ($0.1 - 0.2 \text{ ms}^{-1}$) onshore flow in the upper 100 m over the shelf (Fig. 3e). But neither isopycnals nor isotherms over the shelf downwell toward the coast (Fig. 3 b, c) so the wind stress maximum was likely due to temporal variability.

At the southern front, a minimum of the along front component of wind stress, τ_y , was observed, -0.08 N m^{-2} , at 108.6°W and the wind stress subsequently increased eastward to near zero at 108.5°W , indicating Ekman divergence at the eastern edge of the frontal zone. At the northern front, the north-south wind stress was near zero with an average wind stress of -0.0055 N m^{-2} (Table 1). For both surface fronts, the wind stress was directed up front and would push the lighter surface waters to the east over more dense surface waters to the west, and hence not induce convection, symmetric instability, etc.

The surface buoyancy flux, B_0 , is shown in Fig. 6f. The average value observed in the southern frontal zone, $1.8 \times 10^{-7} \text{ m}^2 \text{ s}^{-3}$ was 50% greater than that observed at the northern frontal zone, $1.2 \times 10^{-7} \text{ m}^2 \text{ s}^{-3}$ (Table 1). The positive surface buoyancies indicated that the surface ocean waters were becoming more dense, a condition which could contribute to symmetric instability.

3.4 Frontal Features

Ocean fronts have a number of kinematic and dynamic features. In order to examine these in greater detail, the region in the area of the northern surface front described above has been enlarged and water properties, velocities, and derived quantities (Turner angles, Richardson Numbers) examined (Fig. 7). Frontal features are described below.

3.4.1 Frontal Slope

Fig. 7a shows the density anomaly superimposed upon the salinity field. The surface front occurred where the 22.7 kg m^{-3} isopycnal shoaled from 40 m to 10 m near $108^\circ 38' \text{ W}$, while the salinity transitioned from 34.9 in the west to 34.6 in the

east. The slope, s , of a front in geostrophic equilibrium can be determined from Margules equation, $s = \frac{f\bar{\rho}(u_1 - u_2)}{g(\rho_1 - \rho_2)}$, where u is along-front velocity, subscript 1 refers to the lighter water, ρ is density, g is gravity, and f is the Coriolis parameter (Fedorov, 1986). These quantities can be estimated from near surface data shown in Fig. 7, where $\rho_1 = 1022.55 \text{ kg m}^{-3}$, $\rho_2 = 1023 \text{ kg m}^{-3}$, $u_1 = 0.33 \text{ m s}^{-1}$, $u_2 = 0.06 \text{ m s}^{-1}$, $g = 9.79 \text{ m s}^{-2}$, and $f = 5.932 \times 10^{-5} \text{ s}^{-1}$ yield $s = -0.0037$.

The slope of the 22.7 kg m^{-3} isopycnal can also be estimated directly from TRIAXUS data. For the region near the front shown in Fig. 7a, there were three data points between 30 and 40 m that intersected the sloping isopycnal. The second and third data points were 0.25 and 1.25 km from the first data point, respectively. Using a least squares fit, $\hat{s} = -0.0084 \pm 0.00015$. Differences between the observed and geostrophic slope can be attributed to the curvature of the front with depth as well as ageostrophic currents.

3.4.2 Communication of the Mixed Layer with the Pycnocline

Surface fronts are regions where the mixed layer communicates directly with the pycnocline (Fedorov, 1986). This is also shown in Fig. 7a where the salinity maximum, $S > 35.1$, coincided with the 23 kg m^{-3} isopycnal at a depth of 35 m at $108^\circ 42' \text{W}$ and sank to the east, forming an intrusion of 15 m thick salinity maximum at the bottom of the mixed layer along the 50 m depth which extended eastward under the surface front. The region $\sim 108^\circ 40' \text{W}$ where the salinity maximum sank was marked by the largest static stability, E , observed along the bottom of the mixed layer ($E = 4.75 \times 10^{-4} \text{ m}^{-1}$ at 52.5 m depth, $108^\circ 40' \text{W}$). The salinity maximum extended to the east ($108^\circ 40' \text{W}$ to $108^\circ 36' \text{W}$) where it met the salinity minimum characteristic of inflowing Pacific waters.

The subduction of these near surface waters under the front is also shown by the fluorometer data (Fig. 7b). Here the maximum fluorescence values ($> 1.4 \mu\text{g l}^{-1}$) observed immediately under the mixed layer in the higher salinity Gulf waters (see also Fig. 5d) was also observed to coincide with the salinity maximum. The tongue

of $>1.4 \mu\text{g l}^{-1}$ fluorescence extended eastward under the front to $108^{\circ}34'\text{W}$, where the core of salinity minimum water ($S<34.3$) was located.

The residence time of the subducted salinity maximum can be estimated. Fig. 7a indicates that it was about 15 m thick and about 5 km in length. East of $108^{\circ}38'\text{W}$, the eastward component of velocity was near zero in the layer of the salinity minimum/maximum at about 45 m - 50 meter's depth; but at the base of the $S>34.9$ waters at $108^{\circ}40'\text{W}$ it is flowing eastward at $\sim 0.05 \text{ m s}^{-1}$. At this rate of eastward flow, the observed salinity maximum would form in about 1.2 days.

3.4.3 Surface Jet

Surface fronts are regions of strong along front-flow, referred to as surface jets, which are directed so that the lighter water is to the right of the direction of the flow in the northern hemisphere. The along-front (north-south) flow is shown in Fig. 7d and the surface jet is readily apparent as peak northward velocities ($>0.25 \text{ m s}^{-1}$) between 10 and 30 m depth centered at about $108^{\circ}37'\text{W}$. At 10 m, the width of the flow $>0.25 \text{ m s}^{-1}$ was 10 km. Note that in both Figs. 6 and 7 the observed peak northward flow was located at the eastern edge of the frontal zone.

3.4.4 Small Scale Structures

Another distinguishing feature of the front shown in Fig. 7a was that the vertical small scale structure differed from that of adjoining water masses. Nowhere else along the northern section did a 20 m thick salinity maximum occur. (It was thinner to the west, and a salinity minimum was to the east.) In the area near $108^{\circ}36'\text{W}$, the isohalines were closer to vertical than to the east or west and with smaller scale (2-5 m) perturbations suggestive of interleaving processes.

The Turner angle (Tu) was determined using one meter TRIAXUS data and is used to describe the stability of the water column to double-diffusive convection (Ruddick, 1983) (Fig. 7e). Three regions were identified:

(i) Tu values between 45° and 90° (colored blue in Fig. 7e) typically appeared as layers 2-15 m thick and 5-10 km wide below 30 m. These waters are unstable to salt fingering.

(ii) Tu values between -45° and -90° (colored red in Fig. 7e) mostly occurred in the upper 40 m of the water column. These were smaller features than (i), typically oriented vertically, less than a km wide, and 5-10 m thick. These were regions unstable to diffusive convection where interleaving could occur.

(iii) Tu values between -45° and 45° (colored yellow in Fig. 7e) were most common, especially in the surface layer near the surface front (108°38'W) and below 50 m to the east of 108°34'W. These waters were stable.

3.4.6 Horizontal Mixing Rates

The fine structure was used to quantify horizontal mixing rates using the method described by Joyce (1977). A steady state balance between temperature and salinity exchanges and small scale vertical mixing was assumed. Across-front fluxes of salt and temperature estimated from

$$-\overline{\tilde{v}\tilde{T}} = \frac{\overline{\tilde{A}_T^v (\partial \tilde{T} / \partial z)^2}}{\partial \bar{T} / \partial y} \quad \text{and} \quad -\overline{\tilde{v}\tilde{S}} = \frac{\overline{\tilde{A}_S^v (\partial \tilde{S} / \partial z)^2}}{\partial \bar{S} / \partial y},$$

where the vertical diffusivities $\tilde{A}_T^v$ and $\tilde{A}_S^v$ were assumed equal to $5 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$.

The lateral diffusivity was estimated as

$$\bar{A}^H = \frac{\overline{\tilde{A}_\phi^v (\partial \tilde{\phi} / \partial z)^2}}{(\partial \bar{\phi} / \partial y)^2},$$

where the bar stands for the mean property field corresponding to the medium scale and the tilde is the small scale departure from the mean field. As pointed out by Joyce (1977), two difficulties arise when applying this model to oceanographic observations: (1) the vertical diffusivities are unknown, as are (2) the location of the boundaries between medium and small scale.

To calculate the lateral heat and salt flux, it was necessary to estimate the vertical T/S gradient of the medium scale field. The approach used here followed Bianchi *et al.*, (1993). Each vertical cycle of TRIAXUS was separated into an up and a down cast. Observations were then interpolated to fixed depths in order to compute the vertical temperature and salinity gradients. For observations in the

frontal region, the spectra of the detrended vertical T and S gradients were estimated for each temperature and salinity profile. Inspection of the spectra revealed a change in regime centered on wavelengths of 8, 4 and 3 m. Similar results were obtained using coherence. The location of the eastern edge of the frontal system was given by the location of where the 34.6 isohaline sloped almost vertically from a depth of 40 m to the surface (108°36'W). The mean large scale cross-front T and S gradients were computed using profiles in the frontal zone, yielding $\partial \bar{T} / \partial z = 24 \times 10^{-5} \text{ }^{\circ}\text{C m}^{-1}$ and $\partial \bar{S} / \partial z = 0.255 \times 10^{-5} \text{ m}^{-1}$. The estimated values of the vertical temperature (salinity) gradient at the above scales ranged from 9.65 to $9.90 \times 10^{-2} \text{ }^{\circ}\text{C m}^{-1}$ ($2.1 \times 10^{-3} \text{ m}^{-1}$), about the same for all scales. The fluxes were then estimated using the value for vertical diffusivities given by Joyce (1977) of $5 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$. Results are shown in Table 2. It should be noted that the salt fluxes were similar for all three scales and the thermal and salt diffusivities were one order of magnitude different.

Table 2

Scales (m)	T flux ($^{\circ}\text{C m s}^{-1}$)	Diffusivity ($\text{m}^2 \text{ s}^{-1}$)	S flux (m s^{-1})	Diffusivity ($\text{m}^2 \text{ s}^{-1}$)
3	7.15×10^{-2}	8200	0.865×10^{-3}	340
4	7.23×10^{-2}	8200	0.865×10^{-3}	340
8	7.56×10^{-2}	8600	0.865×10^{-3}	340

For comparison, the values found by Bianchi *et al.*, (1993) for the Brazil-Malvinas confluence were $1 \times 10^{-2} \text{ }^{\circ}\text{C m s}^{-1}$ and $1 \times 10^{-3} \text{ m s}^{-1}$ for temperature and salinity flux, respectively. Note that Bianchi used a diffusivity value of $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$. Had this value for diffusivity been used with the northern front TRIAXUS data, flux estimates would have been $1.43 \times 10^{-2} \text{ }^{\circ}\text{C m s}^{-1}$ and $1.73 \times 10^{-4} \text{ m s}^{-1}$, 40-70% greater than those for the South Atlantic.

3.4.5 Turbulent Mixing

High spatial resolution TRIAXUS and broadband ADCP measurements provided the opportunity to look for regions in the water column which were susceptible to turbulent mixing. To do this, a gradient Richardson number, Ri , was estimated

$$Ri = \frac{N^2}{\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2},$$

where N is the Brunt Väisälä frequency and the denominator is the square of the vertical shear of the horizontal velocity. $Ri < 0.25$ is a necessary condition for velocity shear to overcome the buoyancy forces of a stable stratified fluid. The one meter density data were smoothed so that they had the same 4 m resolution as the current measurements. Richardson numbers for the frontal region are shown in Fig. 7f. In the upper water column, lowest Ri occurred in the region of the frontal jet, where stability was low in the mixed layer and vertical shear was relatively large. Here the lowest observed Ri was 0.6, more than twice as high as that necessary for turbulent mixing but close to values ($Ri \sim 1$) indicative of symmetric instability (Taylor and Ferrari, 2010).

3.4.6 Relative Vorticity

Mesoscale fronts are marked by positive relative vorticity, $\zeta = v_x - u_y$ (Pallàs-Sanz *et al.*, 2010). Although gradients along the front were not measured, the front appeared to be oriented in a north-south direction and meridional velocities were much greater than zonal velocities. So relative vorticity was estimated from meridional velocity alone, e.g. $\zeta \sim v_x$, and scaled (divided) by the local Coriolis parameter, f . Results are shown in Fig. 8 along with isotachs for the meridional velocity. The largest scaled relative vorticity, $\zeta/f = 0.86$, occurred at 108.5°W at 100 m, and this local maximum of scaled relative vorticity could be followed upward through the bottom of the mixed layer to 13 m depth, where maximum ζ/f had

decreased to 0.55 and was located at 108.64°W. This was the location of the northern surface front described above.

The western edge of the northward flow in the mixed layer was also a maximum of scaled relative vorticity, 0.44, which was observed at 13 m, 109.2°W. This was noted above as a region of large shear in the surface currents and was collocated with the zero isotach of both the meridional and zonal flow (Fig. 6d). Although this front was marked by large velocity shear, it was difficult to discern distinct water mass changes similar to those associated with the front to the west. On November 11, sea surface temperatures increased from west to east along the ship track and the frontal location was close to the 27°C isotherm at 109.2°W (Fig. 1). But on 15 November, temperature and salinity gradients varied near 109.2°W (Fig. 6a,b). At this location, TRIAXUS detected high fluorescence, 0.7 µg/l (Fig. 5b), and salinity, 34.9 (Fig. 5c), in the mixed layer.

4.0 Summary and Discussion

A mesoscale surface front between warm and fresh Pacific waters and Gulf of California waters that were about 0.5°C cooler, 0.4 saltier, and 0.4 kg m⁻³ denser was observed along 108.65°W at about 23.6°N and 23.9°N at the entrance to the Gulf of California. The front was located in water of greater than 2500 m depth near the center of Pescadero Basin and was clearly linked to a region of strong cyclonic shear between inflowing and outflowing waters that extended to a depth of about 500 m below the surface front.

The surface frontal zone varied from 15 km to 10 km from south to north. The narrower northern front was stronger than the southern front, with temperature, salinity, density, and zonal velocity gradients of 0.073 °C km⁻¹, -0.044 km⁻¹, -0.056 kg m⁻³ km⁻¹, and -0.5 x 10⁻⁵ s⁻¹km⁻¹, respectively. The last quantity was an estimate of the deformation rate of the front. The observed slope of the northern front at 35 m was -0.0084±0.00015. Comparison of these quantities with scales for fronts of the World Ocean (Table 1.1, Fedorov, 1986) show that the width of the northern front was large, but other quantities (gradients, slope) were small. The

observed deformation rate was indicative of intensive local frontogenesis (Fedorov, 1986).

The data reported here did not resolve the along-front variability of subsurface features. Even though the 10 km sampling along the southern section was not able to resolve details of the subsurface frontal structure, it was clear that the character of the subsurface southern front differed from that to the north. The denser water in the mixed layer to the west did not appear to penetrate the pycnocline but was observed at the bottom of the mixed layer to the east of the surface front at 40 m depth. At 50 m, the CTD station on the western edge of the surface front showed a 10 m thick layer of low salinity water from west of the front. The zero isotach of the meridional flow did move westward from 108.5 °W to 108.7°W at 70 m depth, but this was much less than observed at the northern front (to 109.2°W). While some of these differences might have been due to much larger southward wind stress observed at the southern front, inhomogeneities existed on 35 km spatial and 28 hour time scales.

Frontal processes were resolved by high resolution sampling of the northern front. (If the survey is repeated, the speed of the vessel through the frontal zone should be halved to 1.9 m s⁻¹ so that the horizontal data resolution would be doubled. For example, this would have resulted in six measurements of the pressure of the 27.7 kg m⁻³ isopycnal vs. the three reported here.) Salinity and fluorescence indicated subduction of Gulf of California waters eastward along the base of the mixed layer under upwelled isopycnals. A 5 km wide northward flowing surface jet with peak northward surface velocities ~0.3 m s⁻¹ extending to 30 m depth was centered on the eastern edge of the front. Gradient Richardson numbers were minimum, ~1, in the surface jet, indicative of symmetric instability.

Small scale features, which persisted over 50-70 km on either side of the northern front, were disrupted at the front, in part due to the subduction of Gulf of California waters that occurred at the base of the mixed layer. The cross-front temperature flux was estimated from the small scale structure as 7.15-7.56 x 10⁻² °C m s⁻¹, while the salinity flux was constant, 0.865 x 10⁻³ m s⁻¹. The diffusivity of temperature (salinity) was 8200-8600 m² s⁻¹ (340 m² s⁻¹). These were relatively large

values compared to observations at open ocean fronts that make it clear these fluxes are important for the redistribution of thermohaline properties in Pescadero Basin.

The lenses of fresher water (intrusions?) seen beneath the salinity maximum at the base of the mixed layer to the west of the front imply more complex processes than described here. These features were 5-15 m thick, 5 km wide, and were observed at a depth of about 50 m along the 24.2 to 24.8 kg m⁻³ isopycnals (Fig. 5c). One of these was located on the western boundary of the frontal zone at 108°42'W and 60 m depth and is shown clearly in Fig. 7a. The temperature-salinity properties of this anomaly were similar to those of inflowing Pacific waters found immediately beneath the salinity minimum, as were the observed velocities (~ 0.1 m s⁻¹ eastward, ~ 0.1 m s⁻¹ northward). They were clearly evidence of complex mixing processes (submesoscale frontal eddies), recirculation of upper layer waters, or perhaps westward displacement of waters below subducted Gulf waters.

An unusual feature of the mixed layer at the northern front was the fact that the zero isotach for the north-south flow was 55 km to the west of the surface front and that the inflowing waters to the west of the front had characteristics of a mixture of Gulf and Subtropical Surface Waters. Two mechanisms might account for this feature. Westward (Ekman) currents at the front associated with northerly winds could push surface waters to the west, but winds were light and eastward velocities were observed in the upper layer. Symmetric instabilities due to positive surface buoyancy could also expand the front westward. It was noted that $Ri \sim 1$, observed near the surface jet at the northern front, was a signature of symmetric stability (Taylor and Ferrari, 2010). Thomas *et al.* (2013) note that symmetric instability could occur if the sum of the surface buoyancy flux (B_0) and the Ekman buoyancy flux (EBF), $(\rho_0^{-1} |\tau_w| |\partial \mathbf{u}_g / \partial z|_{z=0} | \cos \theta \sim -2.5 \times 10^{-8} \text{ m}^2 \text{ s}^{-3}$, were greater than zero. (These quantities were estimated from Table 1, Figure 6e,f and Figure 5h). This condition was met at the northern front, where $B_0 = 1.2 \times 10^{-7} \text{ m}^2 \text{ s}^{-3}$, but not at the southern front, where the wind stress was an order of magnitude greater. The buoyancy flux would cause isopycnals to move from the dense to light side of the front at a rate $B_0 / (H \frac{db}{dx})$ (where the buoyancy, b , is $b = -g\rho\rho_0^{-1}$, db/dx is the horizontal

buoyancy gradient of the front, and H the depth of the mixed layer), about 0.5 m per day. This was too slow to explain the observed westward extension of the front.

Profiles of upper ocean transports of heat ($H(z) = \sum_x \theta(x, z) c_p v_r(x, z) \Delta x$) and volume ($V(z) = \sum_x v_r(x, z) \Delta x$), for the northern section are shown in Fig. 9. Note that v_r is oriented along Pescadero Basin from 159.5°T toward 339.5°T (Collins et al., 1997) so that the transports represent flow into or out of the Gulf. The principal feature of the transport profiles is the marked change that occurred between 49 and 65 m, for example the heat transport increased from -0.96 TW at 49 m to -0.18 TW at 65 m. This indicated that upper ocean (mixed layer and the top of the pycnocline) dominated the total transports. Assuming that the 13 m observation also represents the upper 11 m, the total transport of heat (volume) in the upper 60 m was -72.3 TW (-0.7 Sv, 1 Sv = 10^6 m³ s⁻¹). This is representative of the maximum cooling rate for November, -60 TW, derived by Castro (1994) from climatology.

Note that the westward extension of the front from 108.6°W to 109.2°W resulted increase the area of upper layer inflow by 40% so that instead of transporting warm, saline near surface Gulf waters into the Pacific (as for the waters below 70 m), Gulf surface waters were mixed with TSW at the front and transported back into the Gulf. This slowed the rate at which the Gulf cooled, extended the length of time needed for exchange between the Gulf and Pacific Ocean, and increased the volume of water mixed within Pescadero Basin.

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List of Tables

Table 1. Surface Front Properties

Figure Captions

Figure 1. AVHRR image for November 11, 2004. The color bar at the side of the chart gives the temperature scale ($^{\circ}\text{C}$). Arrows represent near surface flow as measured by vessel mounted current profilers, 15-17 November 2004, for 0.1° longitude bins. Black arrows are from 300 kHz measurements and grey arrows from 75 kHz measurements. The trapezoid (dark-gray) indicates the frontal zone detected by continuous shipboard measurements of sea surface density and salinity.

Figure 2. Sampling locations, November 14-17, 2004. Section N consisted of profiling CTD (TRIAXUS) data and velocity measurements from a 300 kHz broadband ADCP (black segmented line). Section S consisted of CTD stations (indicated by triangles) and velocity data observations using a 75 kHz Ocean Surveyor ADCP. The trapezoid (dark-gray) indicates the frontal zone detected by continuous shipboard measurements of sea surface density and salinity.

Figure 3. Southern section water properties. The white line in figures (a)-(d) represents the base of the mixed layer. (a) Salinity; contour interval is 0.1. (b) Potential temperature, $^{\circ}\text{C}$; contour interval is 1°C . (c) Potential density anomaly, kg/m^3 ; contour interval is 0.3 kg m^{-3} . (d) Dissolved Oxygen, $\mu\text{mol kg}^{-1}$; for dissolved oxygen greater (less) than $20 \mu\text{mol kg}^{-1}$, the contour interval is 20 (5) $\mu\text{mol kg}^{-1}$. (e) East-west velocity, m s^{-1} (westward velocities are shaded blue) and (f) north-south velocity, m s^{-1} (southward velocities are shaded blue): the zero isotach for the velocity sections is indicated by the thick black contour, and the contour interval is 0.1 m s^{-1} . Note: The contour interval is not the same as the interval between different colors.

Figure 4. Temperature-Salinity relationship for TRIAXUS data along the northern CTD section. Longitude ($^{\circ}\text{W}$) is identified by the color bar to the left, and isopycnals (density anomalies have units of kg m^{-3}) are solid black lines which slope upward from left to right. Water masses are identified as Tropical Surface Water (TSW), Gulf of California Water (GCW), Pacific Water (PW) and Subsurface Subtropical Water (SsSTW). Black circles indicate mixed layer waters found between 109.2°W and 108.6°W .

Figure 5. Upper Ocean Measurements from TRIAXUS and a 300 kHz vessel mounted current profiler along the Northern line. For panels a-d, the white line

indicates the bottom of the mixed layer. (a) Potential temperature, °C: contour interval is 1°C. (b) Salinity: contour interval is 0.1. (c) Potential density anomaly, kg m⁻³: contour interval 0.5 kg m⁻³. (d) Fluorescence, µg l⁻¹: contour interval is 0.3 µg l⁻¹. (e) East-west velocity, m s⁻¹, (f) north-south velocity, m s⁻¹, (g) geostrophic velocity, m s⁻¹, relative to measured along section velocity at 100 m, and (h) ageostrophic velocity, m s⁻¹. The contour interval for velocity fields is 0.1 m s⁻¹ and the thick black contour is the zero isotach.

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Figures

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Table 1. Surface Front Properties.

	Southern Section	Northern Section
δx , Width, km	14.6	9.8
$\delta S/\delta x$, S/km	-0.030	-0.044
$\delta T/\delta x$, °C/km	0.011	0.073
$\delta \rho/\delta x$, kg/m ³ -km	-0.026	-0.056
$\delta u/\delta x$, s ⁻¹	-1.3 x 10 ⁻⁵	-0.5 x 10 ⁻⁵
R_ρ	-0.148	-0.695
τ_y , N/m ²	-0.055	-0.005
B_0 , m ² /s ³	1.8 x 10 ⁻⁷	1.2 x 10 ⁻⁷