Research papers

The Costa Rica Coastal Current, eddies and wind forcing in the Gulf of Tehuantepec, Southern Mexican Pacific

Cristóbal Reyes-Hernández a,*, Miguel Ángel Ahumada-Sempoal a, Reginaldo Durazo b

a Universidad del Mar, Instituto de Recursos. Ciudad Universitaria Campus Puerto Ángel, Pochutla, Oaxaca, C.P. 70902, México
b Universidad Autónoma de Baja California, Facultad de Ciencias Marinas, km 107 Carretera Tijuana-Ensenada, Zona Playitas, Ensenada, BC, C.P.: 22860, México

A R T I C L E  I N F O

Article history:
Received 9 September 2014
Received in revised form
17 November 2015
Accepted 23 December 2015
Available online 24 December 2015

Keywords:
Hydrography
Tropical perturbation
Dynamic topography
Tehuantepec

A B S T R A C T

The hydrographic structure and circulation of the Southern Mexican Pacific, from August 31 to September 24 2004, when tropical atmospheric activity was at its peak, was analyzed based on AVISO absolute dynamic topography and an array of 106 CTD profiles, within an area of about 500 km × 500 km between Punta Maldonado and Puerto Chiapas. The surveyed area was occupied by mesoscale anticyclonic and cyclonic eddies that determined the path of water with temperature and salinity characteristic of the Costa Rica Coastal Current. The origin of each eddy was investigated with respect to QuikSCAT wind conditions. The sequence of AVISO images and wind data showed that the largest anticyclonic eddies originated outside the Gulf of Tehuantepec through mechanisms distinct from local wind forcing, although two northerly wind events in the Gulf of Tehuantepec possibly had an influence on the smallest anticyclonic and cyclonic eddies. The relative position of each eddy allowed the flow of relatively low temperature and salinity water (the Costa Rica Coastal Current) into and throughout the Gulf of Tehuantepec, converging at about Puerto Angel with relatively high temperature and salinity water moving from the west.

© 2015 Elsevier Ltd. All rights reserved.

1. Introduction

The Southern Mexican Pacific surrounding the Gulf of Tehuantepec (GT) is located within the Northeastern Tropical Pacific (Fig. 1a) between the southeastern and northeastern boundaries of the large hemispheric anticyclonic eddies that characterize the surface circulation of oceanic scale. Early interpretations of ship drifts by Wyrtki (1965, 1967) led him to propose the Costa Rica Coastal Current (CRCC) as the main current seasonally occupying the south coast of Mexico, directed northwest with a speed exceeding 0.25 ms⁻¹, a width of between 300 and 500 km, and a depth of up to 600 m. Fed by equatorial countercurrents that turn northward at the Costa Rica Dome (CRD), the main water masses potentially related to the CRCC, according to Fiedler and Talley (2006), are: Tropical Surface Water (TSW, T > 25 °C, S < 34), Equatorial Surface Water (ESW, T < 25 °C, S > 34), Subtropical Surface Water (STSW, T~25 °C, S > 35), Subtropical Under Water (STUW, 22 °C > T > 13 °C, 35.5 > S > 34.3), and Antarctic Intermediate Water (AAIW, T~5 °C, S < 34.5). Quarterly thermocline depth maps produced by Fiedler and Talley (2006), their Fig. 9 and dynamic height maps produced by Kessler (2006), his Fig. 7 are consistent with the general picture of the CRCC, but also show some discrepancies regarding both the northward reach of the current and its seasonality. On the other hand, observations made by Trasviña and Barton (2008) and Barton et al. (2009) have challenged the perception of the CRCC as a persistent clearly defined coastal flow.

According to Wyrtki (1967), the CRCC flows along the Mexican coast up to Cabo Corrientes in June–July, but is absent from the coast of Mexico from January to March as it flows westward after detaching from the coast of Costa Rica. Kessler (2006) set the presence of the CRCC as being between the CRD (9 °N, 90 °W) and the Tehuantepec Bowl (13 °N, 105 °W); therefore, west of Puerto Angel and north of 17 °N, the West Mexican Current is independent of the CRCC. The author related the seasonality of the CRCC to the expansion and contraction of the CRD in response to variations in the wind stress curl in the Tropical Pacific, the displacement of the Intertropical Convergence Zone (ITCZ), and variations in the North Equatorial Counter Current (NECC). He also found the NECC to be strongest in November, months after the ITZC reaches its most northern position and when wind stress curl and equatorial upwelling intensify, which consequently intensify the CRCC as well (i.e., months after the CRCC intensification time proposed by Wyrtki). From November to January, as the ITZC...
moves south, the equatorial isotherms are progressively depressed, weakening the NECC and therefore also the CRCC. Because of this the NECC is absent east of 110 °W at the end of the spring (June) and the CRCC is absent from the Gulf of Tehuantepec. Consistent with temperature, salinity and geostrophic flow characteristics typical of the CRCC (cf. Wyrtki (1967)) and in seasonal consistency with Kessler (2006), Brenes et al. (2008) observed a poleward flow between the CRD and Central America during Sep–Oct 1993 and Feb–Mar 1994. Observations within the Gulf of Tehuantepec made in both summer (Jun–Aug 2000; Trasviña and Barton (2008)) and winter (Jan–Feb 1988 and Feb 1996; Barton et al. (2009)) do not totally agree with the concept of the CRCC as a clearly identifiable poleward flow, bound to the coast. Instead of the CRCC, Trasviña and Barton (2008) found an anticyclonic eddy
in response to relatively moderate northerly winds perpendicular to the coast on the western side of the Gulf of Tehuantepec. Therefore, the authors concluded that despite the low frequency and energy of north wind events during summer (and the absence of tropical atmospheric perturbations), it is possible to observe anticyclonic eddies similar to those of winter. They also proposed that the concept of the CRCC as a persistent coastal flow is just the result of the averaged available velocity data used by Wyrkki (1965). Although the intent of Trasviña and Barton (2008) was to characterize the CRCC during its most intense phase (Wyrkki, 1965), their observations actually occurred as the CRCC progressed from its weakest phase in June to a relatively stronger phase in November. It is also very suggestive that their satellite-derived geostrophic velocities showed a persistent meandering current along the periphery of the eddies (June and July) that gradually changed to a rather northwestward flow throughout the mouth of the gulf (August), as north winds relaxed. An absence of hydrographic data meant that it was not possible for them to compare the water characteristics of that flow with the expected CRCC water characteristics.

As the CRCC weakens in winter, Barton et al. (2009) also found northerly wind-driven anticyclonic circulation on the western side of the Gulf of Tehuantepec, as well as eastward and westward coastal currents at the west and east sides of the wind axis, respectively, in a similar fashion to that described by Roden (1961) and Alvarez et al. (1989). It is remarkable that the observed buoyancy-driven coastal westward current was a persistent feature during the two surveys, exhibiting temperature, salinity and geostrophic characteristics consistent with the concept of the CRCC, just as Brener et al. (2008) identified it.

Gap winds commonly provoke mesoscale eddy structures in the gulf of Tehuantepec, Papagayo and Panama (Liang et al., 2009; Alexander et al., 2012) that represent deviations from the geostrophic flow. During the winter, a cold and relatively dry northerly wind regularly blows offshore, perpendicular to the coast of Salina Cruz at the Gulf of Tehuantepec. Most of the gap winds at the Gulf of Tehuantepec show a circular path. Clarke (1988) proposed that such an inertial wind path can form a surface ocean anticyclonic eddy. To the west of the wind jet axis, wind-driven circulation is anticyclonic, the thermocline is depressed, and sea surface temperature (SST) rises. Along the axis, SST decreases due to mixing, while to the east of the jet, the thermocline rises in response to wind-driven cyclonic circulation. Clarke (1988) also showed analytically that the shape of the thermocline is the result of convergence and divergence at the ocean surface to the west and east of the wind jet, respectively.

Detailed atmospheric forcing and oceanic response descriptions of gap wind events can be found in Hurd (1929), Brandhorst (1958), Roden (1961), Blackburn (1962), Stumpf (1975), Stumpf and Legeckis (1977), Clarke (1988), McCready et al. (1989), Barton et al. (1993), Trasviña et al. (1995), Trasviña and Barton (1997), Willet et al. (2006), and Barton et al. (2009). During summer, the atmosphere over Oaxaca lies among the southeastern, southwestern and northern boundaries of the North Pacific high, the Bermuda high and the ITCC low, respectively. Therefore, the north wind events are less numerous, the trade winds are weak (Tomczak and Godfrey, 1994) and are often modified by the displacement of tropical disturbances, reversals in wind direction due to the position of the ITCC (Mexican Monsoon), the Intra-Americas Sea Low Level Jet (Amador, 2008) and positive wind stress curl (e.g. Romero-Centeno et al. (2003)).

Altimetry measurements and numerical modeling carried out by Zamudio et al. (2006) indicated the existence of anticyclonic eddies in the summer that did not respond to northerly wind events, but rather to trapped equatorial waves transmitted to the Gulf of Tehuantepec. Also based on numerical modeling, Zamudio et al. (2008) found coastal trapped waves in the early spring of 2003 to be not of equatorial origin, but instead due to a northerly wind blowing along the coast of the Northern Bight of Panama. These coastal trapped waves propagated poleward and produced similar sea surface height fluctuations to those registered by a coastal tidal gauge located in Manzanillo, Mexico. The study of poleward coastal trapped waves along the Mexican coast was initiated by Christensen et al. (1983) and Enfield and Allen (1983), who demonstrated their origin in tropical storms. These authors also found that from Salina Cruz to Manzanillo, low frequency sea level variations were coherent with wind forcing at Costa Rica.

Numerical experiments examining the generation and properties of mesoscale eddies in the Northeastern Tropical Pacific have indicated that low frequency winds and boundary forcing are the main eddy forcing mechanisms for the whole region. However, the contribution of high frequency winds is more important than that of remote equatorial Kelvin waves within the Gulf of Tehuantepec (Liang et al., 2012). Recently, Flores-Vidal et al. (2014) discovered indications of inertial coastal trapped waves at the Gulf of Tehuantepec that originated from forcing by gap winds at the Gulf of Panama.

Despite observational improvements and an increased understanding of the origin of water masses and forcing mechanisms in the study area, there remains a lack of a complete and consistent conceptual model for the existence of the CRCC within the Gulf of Tehuantepec and its relation with coastal trapped waves, even less so during the season of tropical atmospheric perturbations.

The aim of this work was to compare remotely sensed dynamic topography with in situ dynamic topography and water characteristics in order to establish the path of the CRCC within the Gulf of Tehuantepec during September 2004. Hydrographic data were collected as the CRCC progressed to its annual maximum intensity (following Kessler (2006)). The survey sampled pre-existing as well as emerging mesoscale anticyclonic and cyclonic eddies. The origin and development of each eddy and the corresponding wind conditions were analyzed.

The paper is organized as follows: Section 2 presents the sampled area; Section 3 relates the wind conditions to the transit of the atmospheric perturbations occurring during the survey; Section 4 describes the origin and evolution of the AVISO-identified eddies, along with the related average wind conditions; Section 5 presents the observed hydrographic data and the calculated geostrophic velocity sections; and Section 6 discusses the results.

2. Sampled area

Observations were carried out in and around the Gulf of Tehuantepec between August 31 and September 24, 2004 (Fig. 1). The survey was originally designed with ten transects perpendicularly to the coast (1–X), covering a total area of about 500 km × 500 km, each transect with 12–17 CTD stations. However, adverse weather conditions prevented the survey of transect V, and thus this particular transect is missing in Fig. 1b. The temperature and salinity profiles were constructed using a factory-calibrated CTD SeaBird-19 profiler down to 1100 m depth when possible. The CTD data were pre-processed using the routines provided by the manufacturer and subsequently smoothed and interpolated to every meter; therefore, the term salinity refers to Practical Salinity. In order to avoid bad weather conditions, the vessel surveyed the western transects, VI–I, from Aug 31 to Sep 10, and the eastern transects, VII–X, from Sep 14 to Sep 24. The sampling order is indicated in Fig. 1b by the arrows. Dynamic heights and geostrophic velocities were calculated as geopotential anomalies (m² s⁻²) and from the horizontal pressure gradients, respectively, assuming a reference level of 800 m.
3. Atmospheric tropical perturbations

Between Aug 30 and Sep 24, the mean sea level pressure configuration among the Equatorial Low, North Pacific High and Bermuda High consisted of two troughs, over the Pacific and the Gulf of Mexico respectively, separated by a ridge over continental Mexico (Fig. 2a). Over the Isthmus of Tehuantepec, at 16°–18°N, 94°W, this configuration created a northwestward sea level pressure gradient. The study area experienced the effects of two tropical atmospheric perturbations on the Pacific (Howard and Javier) and of two on the Atlantic (Frances and Ivan, Fig. 1a). The positions and progression of these perturbations relative to the mean distribution of the Bermuda High and Equatorial Low drove zonal wind at the southern end of the surveyed area. Within the Gulf of Tehuantepec, northerly and southerly wind events, the former usually identified as the forcing mechanism for oceanic mesoscale eddies, also responded to the transit of the perturbations. For example, QuikSCAT wind near Salina Cruz from Sep 9 to 20, shifted from northerly to southerly, as first Tropical Storm Javier and later Hurricane Ivan were positioned close to the Tehuantepec Isthmus on the Pacific and Atlantic sides, respectively (Fig. 2b and c).

4. AVISO altimetry and QuikSCAT wind data

During the survey period, the study area was occupied by mesoscale anticyclonic and cyclonic eddies. Two anticyclonic and two cyclonic eddies developed prior to the survey period, and two anticyclonic and one cyclonic eddy developed during the survey period. In this section, the origin and evolution of each eddy are described using selected daily AVISO absolute dynamic topography images for dates prior to and during the survey period, that is, from May 26 to Sep 23 (Fig. 3), together with the corresponding 3-day average QuikSCAT wind conditions (Fig. 4). These data show that the anticyclonic and cyclonic structures did not originate from northerly wind conditions in the Gulf of Tehuantepec, with the probable exception of a single case.

The absolute dynamic topography image for May 26 shows, at the southeastern boundary of the Gulf of Tehuantepec, a sea surface elevation corresponding to a small anticyclonic eddy (Ab) barely visible at the coast of Puerto Chiapas, as well as an anticyclonic eddy (Ae) centered around 12.74°N, 91.95°W (Fig. 3a). On this and the preceding two days (i.e., from May 24 to 26), 3-day average QuikSCAT winds were ≤7 m s⁻¹ and easterly off the gulf, and <3 m s⁻¹ and southerly within the gulf (Fig. 4a). Each of the eddies detached from a westward-migrating ‘parent’ anticyclonic eddy (P), located on May 26 at 10.81°N, 95.32°W. Eddy Ae separated from the northern edge of P around May 3 (not shown), and eddy Ab from the eastern edge of P around May 18 (not shown). After May 26, Ab moved northwestern, following the coast of the gulf to reach Puerto Angel on June 10 (not shown), where it remained until around Jul 1 (Fig. 3b); it then separated from the coast during a northerly wind event with speed ~10 m s⁻¹ (Fig. 4b). Once offshore, Ab blended with Ae, which was previously displaced westward away from the coast and was apparently unaffected by the northerly wind event in the gulf, and with a third anticyclonic eddy also detached from P after May 26 (not shown). While around the Gulf of Tehuantepec the wind on Aug 8 was easterly-southeasterly ~7 m s⁻¹ and within-gulf wind was southerly <3 m s⁻¹ (Fig. 4c), the three anticyclonic eddies formed the single anticyclonic eddy A1 (Fig. 3c). Also on Aug 8, A1 reached about the size found at the beginning of the survey on Aug 31 (not shown). As discussed below, the anticyclonic eddy A2 apparently resulted from a coastal trapped wave originated at the peninsula on the west side of the Gulf of Nicoya (10°N) around Jul 1 (see Fig. 3b), when ~12 m s⁻¹ northeasterly and southwesterly winds blew at the north and south of the peninsula, respectively (see Fig. 4b). Under southeasterly winds, A2 traveled NW along the coast, reaching the southeastern coast of Puerto Chiaapas (15°N) around Aug 8 (Figs. 3c and 4c). On Aug 23, A2 separated from the eastern Gulf of Tehuantepec coast (Fig. 3d), under anticyclonic wind circulation with magnitude <7 m s⁻¹ within the gulf (Fig. 4d).

Although easterly winds dominated most of the surveyed area (except within the gulf, where winds were northerly, Fig. 4e), by Aug 28, A2 reached about the size found later at the beginning of the survey (Fig. 3e). By Aug 31, on the starting day of the survey, the centers of A1 and A2, located at 13.8°N, 97.8°W and 13.8°N, 93.2°W, had radii of about 175 and 155 km, respectively, and...
maximum geostrophic speeds of about 0.5 m s$^{-1}$ (not shown). In between A1 and A2, a meandering current intruded towards the Gulf of Tehuantepec from the south, with maximum intensification occurring around Sep 12 (not shown). For description purposes and later comparison with the hydrographic features, it is considered that two cyclones, centered at 15°N, 94.8°W...
(C1 hereafter) and 10.7 °N, 95.2 °W (C2 hereafter), occupied the north and south ends of the meander, respectively (refer to Fig. 3e). North of 14 °N, the intruding current formed two cyclonic eddies that, as time progressed, moved further north. One eddy remained around the coast of Puerto Angel (C1) and the other (C3 hereafter), with radius ~50 km, was centered at 15 °N, 95 °W. A third anticyclonic eddy (A3 hereafter), which became discernible around Sep 3 centered at 13.0 °N, 90.8 °W, had a radius of about 55 km and maximum geostrophic speed ~0.5 m s⁻¹ (Fig. 3f). A3 separated from the eastern edge of A2 while the latter occupied

Fig. 4. 3-day average QuikSCAT wind velocity field (m s⁻¹) from May 26 to Sep 23, 2004, coincident with the AVISO absolute dynamic topography shown in Fig. 3. Vectors in oceanographic convention. Contours show the Ekman pumping velocity We × 10⁶ m s⁻¹ at intervals of 40 units for |We| < 100 and 200 units for |We| > 100. Positive and negative We represent upward and downward motion, respectively.
the eastern side of the gulf, with C3 displacing eastward towards the
toast of the central gulf and an easterly wind ~10 m s⁻¹
blowing over the study area (Fig. 4f). On Sep 18 a fourth anticyclonic eddy (A4 hereafter) barely appeared between C3 and the
west coast of Salina Cruz (Fig. 3g), when westerly winds ~12 m s⁻¹ blew over the surveyed area (Fig. 4g) and after a
northerly wind event lasting from Sep 10 to 15 (not shown). At the end of the survey on Sep 23, weak easterly and southerly winds (~7 m s⁻¹) existed throughout most of the study area (Fig 4h);
under these conditions the centers of A1, A2 and A3 were located at 13.77 °N, 99.69 °W; 13.26 °N, 95.00 °W; and 12.52 °N, 92.15 °W, respectively (Fig. 3h), and the centers of C1, C2 and C3 at 15.55 °N, 97.23 °W, 12.37 °N, 97.23 °W; and 15.18 °N, 94.53 °W, respectively.

The previous descriptions reveal that the largest eddies—A1 and A3—originated outside the Gulf of Tehuantepec and that the
northerly wind event in the Gulf of Tehuantepec that occurred around Jul 1, contributed only to separate Ab from the coast.
The anticyclonic eddy A2 traveled along the coast from the southern limit of the Gulf of Papagayo after a sea level height crest extended
from about 6 °N under vigorous offshore and shoreward wind. Only anticyclonic eddy A4, located to the west of the usual wind jet axis position, was formed after a northerly wind event in the
Gulf of Tehuantepec. During the survey, the anticyclones migrated mainly in a westward direction: A1 and A2 at about 0.4 km h⁻¹, and A3 at around 0.3 km h⁻¹. C1 and C3 migrated northwestward and northeastward at about 0.4 and 0.05 km h⁻¹, respectively, and C2 moved westward at about 0.3 km h⁻¹. A1, A3, C1 and C3 increased their radii from ~100, 45, 45 and 45 km on Aug 31 to ~225, 87, 75 and 75 km on Sep 23, respectively; A2 reduced its radius from ~129 km to 114 km during the same time interval.

5. Hydrographic data and geostrophic velocity

Most of the temperature profiles (not shown) exhibited the
typical tropical Eastern Pacific three layer structure, comprising a
mixed surface layer no deeper than 70 m, a strong thermocline
(~0.26 °C m⁻¹), and a deep layer with gentle decaying tempera-
ture, starting at roughly less than 150 m. The gross temperature change between the mixed layer and the deep layer was about
16 °C. The TS diagram shown in Fig. 5 reveals the presence of three
water masses: TSW, STUW and AAIW (Fiedler and Talley, 2006),
with the first interpreted as reflecting the surface and inter-
mediate water temperature and salinity characteristics associated
with the CRCC. As expected, near surface water density was mainly
dependent on salinity (TSW is subject to precipitation and evap-
oration, resulting in the wide dispersion of temperature–salinity points in the upper part of the TS diagram), while deeper water density was mainly dependent on temperature. STUW and AAIW characteristics can confidently be related to Subsurface Counter-
currents (SSCCs, Kessler (2006) and Fiedler and Talley (2006)).
The blue and brown colors in the TS diagram reveal a clear distinction between the western high salinity water (transects I–VI)
and eastern low salinity water (transects VII–X) at temperatures higher than 20 °C. As will be seen from the maps and vertical sections presented below, near surface low salinity water was advected from the southern limit of the study area.

Based on the strength of the thermocline, the MLD was defined as the depth at which the vertical temperature gradient was less than or equal to 0.25 °C m⁻¹. According to this criterion, the MLD (not shown) varied between 20 and 60 m and its horizontal dis-
tribution exhibited two depressions, separated by a ridge less than
25 m in depth, almost oriented in the north–south direction
around 96.5 °W and resembling the absolute dynamic topography from AVISO. The maximum depth at each depression was 60 m,
with the minimum depth of about 20 m appearing at two isolated highs at the north and south ends of the ridge.

The dynamic topography map at 5 m depth, estimated from in situ temperature and salinity profiles (Fig. 6a), exhibits most of the anticyclonic and cyclonic eddies identified in the AVISO abso-
dute dynamic topography images (Fig. 3). Eddy A3 was located
outside the surveyed region and therefore is not visible in this
map. A1 appears incomplete, centered around 13.5 °N, 99.5 °W,
with radius about 170 km, close to its Sep 18 position given in
AVISO (Fig. 3g) and coinciding fairly well with the location of the
deepest MLD. A2 had a radius ~150 km, located at 13.2 °N and
94.5 °W, close to the Sep 23 AVISO image position (Fig. 3h) and
near the position of the eastern MLD depression. Between A1 and A2, the two cyclonic eddies C1 and C2 are discernible, centered
about 15.1 °N, 95.7 °W and 11.7 °N, 96.4 °W, respectively. The estimated radii of C1 and C2 were about 150 km; C3, with radius less than 100 km, appeared near the coast of Puerto Chiapas (14.2 °N, 93.5 °W), close to its Sep 23 position in the AVISO image (Fig. 3h). The anticyclonic and cyclonic structures are also evident at 250
and 600 m depth (Fig. 6d and g), but their centers appear shifted
slightly southwestward with depth, such that C1 and C2 increased the separation between A1 and A2. At 600 m depth, the dynamic
height map also reveals the small anticyclonic eddy (A4) centered
at 15.4 °N, 96.5 °W and south of Puerto Angel.

Consistent with the dynamic height contours, the temperature and salinity contours at 5 m depth reveal relatively warm and fresh water (T > 29.5 °C, S < 33.3) spreading westward from Puerto
Chiapas along the peripheries of C3 and C1 (Fig. 6b and c). At Puerto Angel, this water continued southward to mix with warmer water located between Punta Maldonado and Puerto Angel. The temperature and salinity contours at 250 m depth (Fig. 6e and f) show that relatively warm and saline water (T > 11.60 °C, S > 34.760) penetrated the Gulf of Tehuantepec from the south-
east, following the western and northern edges of A2 and then the southern and eastern edges of C3, up to the coast of Puerto Chiapas. Near the southeastern coast of Puerto Angel, also at 250 m, a
tongue of water with the same temperature and salinity charac-
teristics extended southward from the northeast of the gulf,
suggesting the southwest flow of that relatively warm and saline water from Puerto Chiapas to Puerto Angel that continued
southward along the eastern edge of A1. This meandering flow surrounded a relatively cold and fresh water mass (T < 11.60, S < 34.760) extending northward, parallel to the warm water along

![Fig. 5. TS diagram showing the water masses occupying the first 1000 m depth of the Southern Mexican Pacific: Tropical Surface Water (TSW), Subtropical Under-Water (STUW), and Antarctic Intermediate Water (AAIW). Brown dots: eastern region; blue dots: western region. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)](image-url)
the western edge of A2. In addition, by spreading northwest, this cold and fresh water broke the continuity of the warm water southward flow at about 14.5 °N, 97 °W. A similar meandering circulation pattern occurred at 600 m depth (Fig. 6h and i); relatively warm and saline water (T = 7.1 °C, S = 34.570) circulated along the eastern edge of A2, the southern edge of A4 and the eastern edge of A1, isolating relatively cold and fresh water (T < 7.1 °C, S < 34.570) extending from the south.

In the following hydrographic and geostrophic velocity vertical sections, those components perpendicular and parallel to the coast are analyzed next. Because the geostrophic velocity was computed from the dynamic height gradients, each vertical section of this variable corresponds to an intermediate position between temperature and salinity sections. Given the similarity found between sections, only transects VII, IV and II, and lines 13 and 9 are described. The geostrophic velocity components parallel and perpendicular to the coast are defined positive eastward and northward, respectively.

Section VII intersected the center of anticyclonic eddy A2 and the western edge of cyclonic eddy C3 (Fig. 6). Consequently, the shape of isotherms and isohalines was concave from the surface to more than 600 m depth between 200 and 480 km from the coast.
(Fig. 7a and b), and correspondingly the coast-parallel geostrophic velocity component was eastward and westward (Fig. 7c). A2 was symmetric, reaching more than 700 m depth and maximum surface speeds above 20 cm s$^{-1}$. On the other hand, the relatively small temperature and salinity dome about 100 km from the coast and between 30 and 300 m depth, corresponded to C3. The filament of weak westward geostrophic speed, between 50 and more than 800 m depth (Fig. 7c), indicates the western portion of this cyclonic eddy (Fig. 6g). As described on the temperature and salinity maps at 5 m depth (Fig. 6b and c), the shallow core of relatively warm and fresh water above the thermocline ($T > 29^\circ$ C, $S < 33.3$) situated near the coast, corresponded to surface water spreading from the east along the northern edge of C3 (not visible). Below the thermocline, STUW ($10.5 < T < 15.5^\circ$ C and $S > 34.70$) occupied a layer between 70 and 350 m depth. This is the same relatively cold and fresh water mass entering the Gulf of Tehuantepec from the south along the western edge of A2, parallel to relatively warm and saline water to the east, as mentioned in the 250 m depth, temperature and salinity contour maps description above. Below 350 m and down to at least 800 m depth, AAIW with $6 < T < 10.5^\circ$ C and $34.5 < S < 34.7$ occupied the lower column.

Transect IV intersected the western edge of C1, the eastern edge of A1, and the western edge of C2 (Fig. 8). Therefore, both...
Isotherms and isohalines exhibited soft convexity and concavity from the coast to 400 km offshore, and from the surface to about 350 m depth (Fig. 8a and b). At the surface, relatively warm and fresh water ($T > 29.5^\circ C, S < 33.8$) extended throughout the section, except between 90 and 150 km from the coast, where relatively cool and fresh water ($T < 29^\circ C, S < 33.5$) emerged. Consistent with the relative positions of C1, A1 and C2, the temperature and salinity maps at 5 m depth (Fig. 6b and c) show that water with $T > 29.5^\circ C$ and $S < 33.5$ moved westward along the coast of Puerto Angel and then to the south, to surround the cool and fresh water mentioned above. Also, again corresponding to the positions of these eddies, the coast-parallel geostrophic velocity component (Fig. 8c) consisted, in the top 100 m and in the offshore direction, of westward and eastward velocities, respectively, and below 100 m depth and beyond 220 km from the coast, of westward velocity. This last westward speed corresponded to the northern edge of C2. Maximum eastward surface velocities $\sim 10 \text{ cm s}^{-1}$ decayed to less than $1 \text{ cm s}^{-1}$ at 460 m depth, while maximum westward velocities, about 2 and 1 cm s$^{-1}$, occupied layers between 100 and 400 m depth, at the northern and southern ends of the section. STUW ($10.5 < T < 15.5^\circ C, S > 34.70$) was present at between 80 and 380 m depth near the shore, whereas AAIW
A1, A4 and C1 (Fig. 9). Consistent with the size of A1, isotherms and isohalines were gently concave in shape across the section (Fig. 9a and b). Above the thermocline, however, warm and fresh water \((T > 30.5^\circ \text{C}, S < 33.5)\) occupied the upper 45 m and first 50 km from the coast, corresponding to the eastward geostrophic velocity that formed part of the small anticyclonic eddy attached to the northern edge of A1 (Fig. 9c). This water spread from Puerto Escondido to Puerto Angel, where it turned southward, as suggested by the shape of the 33.4 isohaline on the 5 m depth map (Fig. 6c). Eddy A1 leaned southwest with depth, so that the vertical structure of the coast-parallel geostrophic velocity component consisted of eastward and westward velocity bands related to A4, as well as to the western edge of C1 that intruded progressively below the eastward velocity with depth. Eastward and westward velocities were at their maxima near the surface \((\sim 5 \text{ cm s}^{-1})\) and decayed to zero at \(\sim 550\) m depth, with their vertical axis bending northward at a depth coincident with that of the thermocline. Below the thermocline, STUW occupied a layer between 110 and 310 m depth, while AAIW occupied a layer from 310 to about 800 m depth. Both water masses moved mainly westward as part of the meandering current described above.
The vertical section corresponding to line 13 spanned from transect I to transect IX, a distance of almost 500 km. Consistent with the positions of the northern edge of A1 and the southern edge of C1, isotherms and isohalines were convex in shape, with the crests located between 300 and 400 km from transect I (Fig. 10a and b). Therefore, southward and northward geostrophic velocities occupied the western and eastern sides of the section, respectively, to form the meandering current described above (Fig. 10c). Above the thermocline, relatively warm and saline water \((T > 29.5\, ^\circ C, S > 33.6)\) moved southward, while relatively cool and fresh water \((T < 29.5\, ^\circ C, S < 33.5)\) moved northward. Below the thermocline, STUW and AAIW occupied the whole section from 80 to 300 m depth and from 300 to more than 800 m, respectively.

As line 9 intersected C1 and the western edge of C3, isotherms and isohalines above the thermocline were thus convex over the eastern half of the section, while below the thermocline and to the west of the section they were concave in shape (Fig. 11a and b), suggesting the presence of a small anticyclonic circulation at the northern edge of A1 (see description of transect II above). Consequently, C1 appeared embedded in the upper 100 m of a larger anticyclonic circulation. In the top 50 m, relatively warm and low salinity water \((T > 29.5\, ^\circ C, S < 36.6)\) moved southward and relatively cool and fresh water \((T < 29.5\, ^\circ C, S < 36.6)\) moved northward over the western and eastern sides of the section, respectively (Fig. 11c). As indicated by the isotherms and isohalines in the
5 m depth map (Fig. 6b and c), these two waters corresponded to
the eastward and westward surface flows that converged south of
Puerto Angel to turn southward. Between those two waters, rela-
tively saline water ($S > 33.6$) emerged at the surface and flowed
southward, at the usual location of the north wind jet axis within
the Gulf of Tehuantepec ($\sim 320$ km to the east of transect I). Below
the thermocline, the salinity maximum identifying STUW occu-
pied a layer between 70 and 300 m, while AAIW occupied that
from 300 to more than 800 m depth. As components of the
meandering current, both water masses moved northward and
southward along the western and eastern sides of the section,
respectively. A portion of STUW, upon reaching $15^\circ$ N, was able to
flow eastward along the northern periphery of A2 and then flow
toward the Gulf of Tehuantepec along the eastern edge of C3 (see
Fig. 6e and f). The maximum northward and southward velocities
were $> 5$ cm s$^{-1}$, decaying to $\sim 1$ cm s$^{-1}$ between 500 m and
700 m depth.

6. Discussion

AVISO absolute dynamic topography and dynamic topography
derived from the CTD cast array both exhibited most of the me-
oscale structures of the anticyclonic eddies A1, A2 and A4, as well
as those of the cyclonic eddies C1, C2 and C3. A4 and C3 were the smallest in scale and were located at the western and eastern sides of the coast of Salina Cruz, respectively. A3, while present in the AVISO images, was located outside the CTD sampling area.

The sequence of AVISO images suggests that most of the identified anticyclonic and cyclonic eddies (A1, A2, C1, C2 and A3) originated outside the Gulf of Tehuantepec—the first four before the start of the oceanographic survey. In contrast, A3, C3 and A4 formed during the survey. A1 and A3 detached from pre-existing anticyclonic eddies, and A2 from the propagation of a coastal trapped wave. C3 detached from the meandering current intruding toward the gulf.

As revealed by AVISO and QuikSCAT data for Sep 3, the development of A3 and the eastward migration of C3 occurred under easterly wind conditions, just before the brief northerly wind event of Sep 4–6. This northerly wind event resembled the inertial wind path analyzed by Clarke (1988), albeit with a rather smaller radius of curvature (~180 km). Eddy A1 did not evidence changes in the path of the wind jet other than a relatively rapid southward displacement. The northerly wind conditions corresponded to the southward expansion of the atmospheric pressure ridge over continental Mexico and the presence of Hurricane Frances.

A4, located west of Salina Cruz, initiated after the northerly event occurring from Sep 10 to 15 and was recognizable in the AVISO images by Sep 18. Therefore, A4 had the appropriate space and time locations related to the role of northerly gap wind events in the formation of mesoscale eddies in the Gulf of Tehuantepec. However, the wind did not resemble an inertial wind path but instead veered from northwesterly to northeasterly as time progressed (not shown). These wind conditions were due to the intensification of the northwestern sea level atmospheric pressure gradient caused by Hurricane Ivan.

Due to the perturbation to the depth of the thermocline that zonal wind can produce, A2 appears to have originated as a coastal trapped wave at the peninsula of the Gulf of Nicoya.

By using QuikSCAT wind velocity vectors, the effect of wind on thermocline depth can be obtained by assessing the Ekman pumping velocity \( W_e \) as a function of the wind stress curl and the \( \beta \) effect:

\[
W_e = \frac{1}{\rho} \frac{1}{f} \frac{\partial \tau_Z}{\partial x} - \frac{\partial \tau_x}{\partial y} - V\beta
\]

where \( f, V = \int_0^D v dz, \beta = \frac{df}{d\varphi}, \rho = 1022.0 \text{ kg m}^{-3}, \tau_x \) and \( \tau_y \) are the Coriolis parameter, the meridional Ekman transport, the meridional gradient of the Coriolis parameter, the mean water density, and zonal and meridional wind stress components at the surface, respectively. In this exercise, the lower limit of the integral for the meridional Ekman transport corresponded to the estimated frictional layer depth \( D \) (meters). The frictional layer depth itself was estimated using the expression given by Ekman (1905), but modified with wind stress parameterized in terms of the drag coefficient (cf. Pond and Pickard (1978)):

\[
D = \frac{4.3W}{\sqrt{\sin \varphi}}
\]

here, \( W \) represents the wind speed in m s\(^{-1}\) and \( \varphi \) the latitude. The frictional layer depth results from the solution to the Ekman balance equations and is conceptually different from the MLD, which is related to the history of the momentum and heat exchange in the water column; \( D \) will be shallower than the MLD if the available turbulent kinetic energy is not enough to break stratification.

The contours of \( W_e \left( \times 10^6 \text{ m s}^{-1}\right) \), positive (upward) and negative (downward), paralleled the wind stress direction to the left and right of the down-wind direction, respectively (Fig. 4). Clearly, westerly and easterly winds caused the largest thermocline displacements throughout the whole region. Near the Gulf of Nicoya, westerly and easterly winds periodically rise and depress the thermocline (increasing or decreasing relative vorticity), as observed on Jun 29–Jul 1 when A2 became evident (Figs. 4b and 3b). Within the Gulf of Tehuantepec, northerly wind events occurring on Jun 29–Jul 1, prior to the survey, as well as on Sep 1–3 and Sep 10–15, raised and depressed the thermocline to the east and west of the jet, but with varying effectiveness according to the magnitude of the velocity and the velocity field distribution. Whereas the moderate Jun 29–Jul 1 event corresponded to the separation of eddy Ab from the coast (Figs. 4b and 3b), the weak event recorded on Sep 1–3 had no associated eddy formation (Figs. 4f and 3f). In contrast, the strong event of Sep 10–15 (not shown) agreed qualitatively with the appearance of eddy A4 (Fig. 3g), despite the fact that after Sep 15, wind conditions were westerly (Fig. 4f).

The three anticyclonic eddies (A1, A2 and A3) and two cyclonic eddies (C1 and C2) migrated mainly westward at speeds of ~0.3 km h\(^{-1}\). This value agrees satisfactorily with the celerity (C\(_e\)) of a Rossby wave, 0.21 km h\(^{-1}\) (cf. Tomczak and Godfrey (1994)), if a thermocline depth of about 70 m, two layers of density of 1021 and 1026 kg m\(^{-3}\), and latitudes of 14 °N and 15 °N are assumed. On the other hand, the deformation radius, \( R_0 = \frac{f}{\beta} \sim 43 \text{ km} \), suggests that A1, A2 and A3 were close to being in geostrophic balance. Using the radius \( R \) of each of the largest eddies, the latitude of 14 °N and \( f = 3.5 \times 10^{-5} \text{ s}^{-1} \), the Rossby numbers \( R_0 = \frac{V}{R} \sim 10^2 \) indicate that they were in fact subinertial.

At the end of the survey, the position of A1 suggests it reached the location of the Tehuantepec Bowl; however, A1 was a deeper feature than this dynamic structure, whose estimated depth is about 200 m (Kessler, 2006). Kessler (2006) pointed out that in winter the Tehuantepec Bowl can block the passage of surface water with CRCC characteristics further west of the Gulf of Tehuantepec, although it does allow the passage of subthermocline poleward flow. In summer, however, when the Tehuantepec Bowl is weak and located offshore, an “additional leakage to the north” of the CRCC can exist. The temperature and salinity maps, as well as the dynamic height contours up to about 450 m depth, are consistent with such a northward coastal flow of the CRCC up to Puerto Angel for September 2004 (Figs. 6a and d). Water with temperature and salinity associated with the CRCC (TSW and STUW) traveled northwestward following two paths (Fig. 12). On the first path, water moved westward along the coast, around the northern edge of C3. From there, it split in two branches: one continuing along the coast and the other turning southward...
between A4 and C3, westward between A4 and C1, and, after reaching the west of Puerto Angel, turning southward along the periphery of A1. On the second path, water meandered along the eastern periphery of C2 and the western edge of A2. Part of this water moved eastward between C3 and C1 and part converged with water from the first path moving southward between A4 and C3, to move then westward along the southern edge of A4, or southward along the western edge of C2 and the eastern edge of A1. In contrast, near-surface water with relatively high temperature, and high salinity water from the west (but also in correspondence to CRCC characteristics), moved along the northern edge of A1 to converge with water moving from the east, in front of Puerto Angel. Between Punta Maldonado and Puerto Escondido, part of the eastward-moving water possibly turned back along a barely discernible coastal cyclone.

These results also suggest that during the survey, the near-surface flow was cut off by a northerly flow perpendicularly to the coast of Salina Cruz. Additionally, the sequence of AVISO absolute topography and geostrophic velocity images (Fig. 3) suggests that water of the CCCR flowing westward near the coast of Central America should be blocked by the anticyclones (A2 and A3) on the eastern side of the gulf, a seemingly novel result. The interruption by wind of the surface coastal flow, originally mentioned by Reyes-Hernández and Murad (2005), is consistent with the findings of Flores-Vidal et al. (2011), who described a westward near-surface current interrupted by moderate to strong northerly wind bursts in the Gulf of Tehuantepec; however, they failed to describe the subsurface circulation. Here it was observed that below the thermocline, the characteristics of the anticyclonic and cyclonic eddies prevailed, despite the diversity of wind conditions present along the survey.

Acknowledgments

Hydrographic data were collected through ship time awarded to UMAR by Secretaría de Marina, Armada de México. QuiLoS/CAT winds were provided by NASA, Jet Propulsion Laboratory, California Institute of Technology, http://podaac.jpl.nasa.gov/. Altimetry maps were provided by AVISO http://www.aviso.oceanobs. com/. Special thanks go to the BO Altair Chief Commander and its crew, to Lt. Paul Murad, and to UMAR technicians and students: Samuel Ramos Carrillo, Sergio Mendoza Vázquez, Ramón Sánchez Vázquez, Elvira Rodríguez Marcos, Adriana Barragán Aparicio, and Jesús Germán Romero. Thanks also go to three anonymous reviewers that helped to improve substantially this work and to Dr. Miguel Lavin Peregina and Dr. Victor Godínez for their useful comments regarding an earlier version of the manuscript. RD was partially at SIO-UCSD-CASPO hosted by D. Rudnick, as the recipient of the Earth’s Rotation on Ocean Currents. Reprinted from Arkiv für Matematik astronomi och fysik. Publ. by K. Svenska Vetenskapsakademien (R. Swedish Academy of Sciences). Band 2: No. 111.


