Upwelling-enhanced seasonal stratification in a semiarid bay

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Abstract
The role of wind-driven upwelling in stratifying a semiarid bay in the Gulf of California is demonstrated with observations in Bahía Concepción, Baja California Sur, Mexico. The stratification in Bahía Concepción is related to the seasonal heat transfer from the atmosphere as well as to cold water intrusions forced by wind-driven upwelling. During winter, the water column is relatively well-mixed by atmospheric cooling and by northwesterly, downwelling-favorable, winds that typically exceed 10 m/s. During summer, the water column is gradually heated and becomes stratified because of the heat flux from the atmosphere. The wind field shifts from downwelling-favorable to upwelling-favorable at the beginning of summer, i.e., the winds become predominantly southeasterly. The reversal of wind direction triggers a major cold water intrusion at the beginning of the summer season that drops the temperature of the entire water column by 3–5 °C. The persistent upwelling-favorable winds during the summer provide a continuous cold water supply that helps maintain the stratification of the bay.

1. Introduction
Water column stratification of the density field in coastal environments can arise from a variety of causes. The major mechanisms controlling vertical stratification are buoyancy inputs, mixing produced by winds and tides, and the interaction of vertically sheared currents with a horizontal density gradient, i.e. tidal straining (Simpson et al., 1990). In coastal environments devoid of freshwater influences, the main stratifying agent to the water is atmospheric input of heat at the water’s surface, which produces typical seasonal patterns of vertical stratification. This study will show that coastal upwelling can provide an additional source of buoyancy, in this case negative buoyancy, into a coastal embayment from the adjacent ocean thus enhancing water column stratification.

Classically, coastal upwelling has been understood as a wind-forced process (e.g. Gill, 1982) with alongshore wind stress driving offshore surface Ekman transport and upwelling of deeper waters. Therefore, the forcing by winds and the reversal of their direction can be expected to play an important role in determining the water temperature and density stratification in coastal embayments. For instance, in the western coast of the Iberian Peninsula, shelf winds follow a seasonal pattern and water column stratification in the Ría de Vigo is enhanced by the cold water driven by upwelling winds (Gilcoto et al., 2007).

In this study, year-long observations of water column stratification were carried out in Bahía Concepción (Fig. 1), located in the east coast of the Baja California peninsula in Mexico. The main objective is to explore the relationship between water column stratification and wind direction, and in general to demonstrate the role of the adjacent ocean in enhancing the stratification of a bay or lagoon. In Section 2, a description of study site and data collection is given. In Section 3, the seasonal evolution of stratification, the estimation of advective contribution to heat content and the dominant modes of temperature variability are presented. In Section 4, relationships between winds and water surface level are addressed. Also, two mechanisms relevant to wind forcing are discussed: (a) the generation of longitudinal density gradients by wind-driven upwelling, thus contributing to density-driven flows; and (b) the mixing stratification competition. In Section 5, the main conclusions of this study are presented.

2. Study site and observations
Bahía Concepción is a ~40 km-long and ~5–10 km-wide bay oriented along a nearly NNW–SSE axis with an opening to the Gulf of California, the bay entrance, to the north (Johnson and Ledesma-Vásquez, 2001). The bay’s bathymetry is relatively simple. The entrance to Bahía Concepción consists of a central channel with a sill at about 15 m tucked against the eastern shore.
In the southern half, the bay geometry resembles a 30-m-deep bathtub. The seasonal wind pattern is predominantly NW during autumn and winter (cold period with well-mixed water column) and SE during spring and summer (warm period with stratified water column). During the summer season, the vertical temperature difference is well defined while salinity stratification is less obvious (Lopez-Cortes et al., 2003; Palomares-Garcia et al., 2006; Canar et al., 2008).

Moored records of water temperature were obtained from February to October 2005 along a cross-bay transect (Fig. 1) consisting of six thermistor (Tloggers) chains. At the central mooring sites (stations 3 and 4), temperature sensors were deployed at five equidistant depths ranging between near-surface and near-bottom (i.e., 5, 10, 15, 20, and 25 m). At the four other stations, water temperature was measured at three equidistant depths: 5, 10, and 15 m.

Bottom pressure was measured at 15 min intervals near the entrance and close to the head, at the south end of the bay using CTDs with Paroscientific pressure sensors (SeaBird SBE26). Winds were recorded near the pressure gauges using Aanderaa anemometers sampling at 1 Hz. The anemometers were deployed on land at a height of 10 m above mean sea level. In addition, regional winds were collected from QuikSCAT satellite open data with 12.5 km resolution (http://aicea.jpl.nasa.gov/DATA/QUIKSCAT/wind/). The station in the middle of the Gulf of California (27.375° N, 111.375° W) was chosen as a representative of remote wind conditions for the bay. Daily net heat flux at the water surface was obtained from the Global HYCOM (HYbrid Coordinate Ocean Model) model, which has ~7 km mid-latitude resolution (http://hycom.rsmas.miami.edu/dataserver/). HYCOM uses a penetrating solar radiation scheme (Kara et al., 2005) and can predict relatively accurate sea surface temperatures (Wallerf et al., 2008). Because no model grid points are available inside Bahía Concepción, the data from the grid point nearest to the bay entrance was selected as a representative of the entire bay.

The data analyses described here are based on 15-min averages of the original observations, except for the half-hourly QuikSCAT winds and the daily HYCOM heat flux. Short gaps are filled by linear interpolation and periods where longer gaps exist are excluded from the analysis. Wind stress is estimated from wind speed and direction following Large and Pond (1981) for a neutral atmosphere. Subtidal water level is obtained by low-pass filtering the bottom pressure with a Lanczos filter at half-power of 34 h. Local wind stresses were rotated along the axis of the south part of the bay while the remote wind stresses were rotated to the alongshore direction representative of the west coast of the Gulf of California.

3. Results

3.1. Seasonal evolution of stratification

Observations that spanned nearly a year are illustrated in Fig. 2. Water temperature at the central site (station 4) was selected as representative of the other sites (first panel). Generally, water temperature at all five depths increased from winter to summer, showing a seasonal warming (Fig. 2a). During the winter period (before day 110), water temperature was nearly uniform in depth indicating a well-mixed water column. Starting on day 110, the water temperature at the five depths diverged as the water column became stratified. From day 145 to 150, water temperature dropped throughout the water column. This drop was steeper at larger depth. After that drop, the water column restored its warming trend and water temperature continued to increase until past the autumn equinox, day 275, when vertical stratification was ~0.5 °C/m. During the stratified summer season (after day 150), the water temperature near the bottom (at depths of 20 and 25 m) increased approximately linearly (if tidal variations are neglected) by 5 °C in 125 days or 0.04 °C/day. In contrast, the temperature in the middle of the water column (depths of 10 and 15 m) fluctuated distinctly at higher amplitudes and frequencies than near the bottom. This indicated that the cooling event on days 141–155 resulted from a major cold water intrusion into the bay that dropped the entire water column temperature. After the major cooling event, intermittent cold water input also affected the heat content of the water column but not as markedly as the major cold water intrusion (this will be shown in Section 3.2).

The net heat flux increased from winter to summer showing seasonal heat transfer from the atmosphere to the ocean (Fig. 2b). From day 75 onward, the net heat flux became positive as the bay began to gain heat from the atmosphere. The general trend of net heat flux was responsible for the warming of the bay and contributed to the development of summer stratification in the water column. The major water cooling event on days 141–155 should not have been caused by atmospheric heat fluxes because such fluxes did not decrease during this period. Interestingly, there was a cooling from ~200 to ~60 W/m², on day 150, that only affected the water temperature at 5 m, i.e., it did not reach depths larger than 10 m. Furthermore, the fluctuations of the thermocline were not correlated to the net heat flux, which suggested that atmospheric heat fluxes during the stratified period only influenced the near-surface layer (upper 5 m) of the bay. Therefore, the water column cooling on days 141–155 must have been caused by advection of cold water into the bay from the adjacent Gulf of California in response to upwelling, as explored next.
The wind in the Gulf of California is characterized by distinct seasonal variations (Fig. 2c). Before day 140, during the winter and early spring, the wind was predominantly toward the SE while during the summer (days 140–275) it was toward the NW. After day 275, during the autumn, the wind shifted back toward SE. Winds toward the NW during the summer drove surface water away from the bay entrance and produced upwelling toward the coast. In general, summer winds produce upwelling in most of the western coast of the Gulf of California (Badan-Dangon et al., 1991). The cooling event on days 145–150 coincided with the beginning of northwestward winds in the gulf. Therefore, this major cooling event should have been caused by the upwelling-driven cold water intrusion from the adjacent ocean basin. The persistent northwestward winds throughout the stratified summer season maintained a cold water supply into the bay. During the winter and autumn, southeastward winds dominate and cause downwelling over the eastern coast of the Baja California peninsula. These persistent downwelling winds should effectively reverse the exchange pattern at the bay entrance and produce upwelling toward the coast. In general, summer winds produce upwelling in most of the western coast of the Gulf of California (Badan-Dangon et al., 1991). The setting of \( T_{ref} \) to zero rather than to some other value has practically no influence on the \( HC \) variability through \( \rho \), because the density difference in the water column \( \Delta \rho \) is small compared to the water density itself, i.e. \( \Delta \rho \approx \rho \). For computation of \( C_p \), the empirical formula given by Millero et al. (1973) has been applied. The estimation is carried out at the center site (station 4). In the calculation of heat content, high frequencies of the temperature data were removed by using a Lanczos filter at half-power of 3 days. The calculated heat content then was smoothed with a 5-point running mean (Fig. 3 top panel) in order to reduce fluctuations of the estimated heat flux.

The heat content \( HC \) in \( J/m^2 \) of the water column can be expressed as

\[
HC = \int_{-H}^{0} C_p \rho (T - T_{ref}) \, dz,
\]

where \( C_p \) is the specific heat of seawater, \( \rho \) is water density, \( T \) is water temperature, \( T_{ref} \) is a reference temperature, which is arbitrarily set to zero °C, \( H \) is the depth of the water column (taken as 30 m for station 4). The setting of \( T_{ref} \) to zero rather than to some other value has practically no influence on the \( HC \) variability through \( \rho \), because the density difference in the water column \( \Delta \rho \) is small compared to the water density itself, i.e. \( \Delta \rho \approx \rho \). For computation of \( C_p \), the empirical formula given by Millero et al. (1973) has been applied. The estimation is carried out at the center site (station 4). In the calculation of heat content, high frequencies of the temperature data were removed by using a Lanczos filter at half-power of 3 days. The calculated heat content then was smoothed with a 5-point running mean (Fig. 3 top panel) in order to reduce fluctuations of the estimated heat flux.

The heat content generally increases during the observation period and shows, as expected, that the water column is gaining heat from winter to summer. A noticeable decrease in heat content around day 150 corresponds to the major cooling event revealed from the temperature evolution (Fig. 2 top panel). Another similar cooling event occurred around day 240 but has a smaller reduction of heat content relative to the major cooling event. From day 170 to day 220, the heat content increased at a slower rate than that during the winter, indicating the additional supply of cold water.

The time derivative of the heat content represents the total net heat flux (Fig. 3 bottom panel). During winter, the estimated total heat flux shows a similar trend as that from HYCOM data.
showing that the heat content in the water column is mainly determined by atmospheric heat input. During the summer, however, large discrepancies between the two time series of heat flux can be found, corresponding to reductions in heat content. The drop in the estimated heat flux indicates loss of heat and is attributed to horizontal advection of cold water. The largest discrepancy appears on day 150 representing the major cold water intrusion. The other discrepancies that occurred during summer were smaller than the major cooling event, and show relatively weak intermittent cold water supply to the basin. This is consistent with the temperature evolution as shown in Fig. 2a.

3.3. Dominant modes of temperature variability

The spatial patterns of temperature at the observation transect are determined using empirical orthogonal functions (EOFs) (Fig. 4). The temporal mean of the transect shows a stratified water column where temperature decreases approximately linearly from surface to bottom with a top-to-bottom difference of \( \sim 6 \, ^\circ C \) or \( \sim 0.2 \, ^\circ C/m \) on average. This mean temperature distribution is practically uniform across the transect but the isotherms slope upward (from west to east) and then downward close to the eastern shore of the bay. This could be caused by increased mixing effects near the edges of the bay’s channel. The first two EOF modes explain 97% of the variance and the other modes are negligible.

The first EOF mode accounts for 89% of the total variance and has a similar spatial pattern to the mean, representing warming/cooling of the entire water column. The contribution of each EOF mode to the temperature variance is the product of the EOF mode and the corresponding principal component (PC). Because the first EOF mode is positive, positive PC1 represents warming while negative PC1 indicates cooling of the water column. The PC1 increases from \(-25\) to \(22\), showing a trend of warming of the water column from winter to summer. Before day 140, the PC1 is negative and gradually increases. This means that the water column is cooler than the mean in that period and is gradually heated between winter and summer. During the summer period the PC1 is mostly positive and represents warming of the water column. The evolution of PC1 is very similar to the heat content (Fig. 3a). The three cooling events identified from heat content are also found in PC1. According to the analysis just described, the first EOF mode represents effects of both surface heating and cold water intrusions from the adjacent gulf. The cold water intrusion is related to the along-shelf wind, which is obtained by projecting the remote winds to the axis of the Gulf of California (Fig. 4 lower panel). Positive values represent northwestward winds. Because the cold water intrusion is indirectly related to along-shelf winds, the correlation between the wind stress and PC1 is not high. However, the three cooling events inferred from the heat content and PC1 can be interpreted with the evolution of the wind stress. The cold water intrusion occurred around day 150 and is clearly related to the major shift in wind direction. During days 170–220, the southeastward winds are generally strong (around 10 m/s) and account for the cooling of the water column. Around day 140, there is a distinct pulse of southeastward winds that are responsible for the cooling event on day 140.

The second mode explains 8% of the total variance for the entire cross-section and is characterized by positive values near the surface and negative values at depth. This represents a warming at a surface layer and simultaneous cooling at a layer underneath. Both modes 1 and 2 include the signature of the cold water intrusions. When PC2 is positive, mode 2 tends to reinforce the mean pattern, i.e. to increase vertical stratification in temperature. In contrast when PC2 is negative, mode 2 tends to reduce the vertical temperature difference because the surface cools down and the near-bottom layer warms up. Therefore, a positive PC2 represents increased stratification and vice versa. The PC2 is well correlated with top-bottom temperature difference (the correlation coefficient, \(r\) is 0.87). Because stratification is partly driven by the remote wind, the general evolution of PC2 agrees with the trend of remote wind stress: when the wind is toward SE (positive wind stress), PC2 is negative, indicating that
northwesterly winds reduce stratification; when the wind is toward NW (negative wind stress), PC2 is positive, indicating that southeasterly winds enhance stratification.

4. Discussion

4.1. Relationships between winds and water surface elevation

Subtidal fluctuations of water level in a semiclosed basin can be caused by remote or local forcing (e.g. Wang, 1979; Garvine, 1985; Wong and Valle-Levinson, 2002). Remote forcing produces coastal water level fluctuations that drive unidirectional net flows throughout the entrance to a basin. Local winds produce sea-level slopes inside the basin and bidirectional exchange flow at the entrance to the basin. Because the orientation of the along-basin axis in Bahía Concepción is nearly parallel to that of the Gulf of California, along an approximately northwest–southeast direction, both remote and local winds act in concert and produce water level slopes along the bay. The subtidal water level differences between the head and the mouth stations (Fig. 5b) responded to the remote and local winds in the expected fashion: northwesterward winds caused sea level to rise toward the north and east in the Gulf of California. Inside the bay, northward winds caused the water level to increase at the mouth relative to the head, while southeastward winds produced the opposite response. The subtidal water level at the head was generally higher than that at the mouth during the winter (positive values, days 80–140, Fig. 5b), then dropped at the beginning of the summer (negative value) in response to the wind reversal. This further illustrates the shifting from downwelling to upwelling conditions at the beginning of the summer. The water surface elevation difference of ~5 cm during the major cooling period around day 150 was caused by the wind reversal. As southeasterly winds continued throughout the summer, the water level slope bounced back by day ~200 and fluctuated from then onward.

The local wind field in Bahía Concepción (Fig. 5c) was consistent with the remote wind outside the bay on the Gulf of California (Fig. 4). Positive values represent southeastward wind stress. It is clear that during the winter period local winds were mostly southeastward (positive) and the wind stress was larger than in the summer, exceeding 0.1 Pa at times. In the summer, local winds were mostly northwestward and relatively weak. The strong winds during the winter contributed to mix the water column and the weak winds during the summer helped to maintain the stratification. It is evident that the shifting of local wind pattern is not as influential as that of remote wind. This is partly because the local wind is highly influenced and steered by local topographical features. This is because Bahía Concepción is the result of land downthrow near a geological fault zone with steep escarpments (Johnson and Ledesma-Vásquez, 2001). Thus topography confines local winds along the Concepción fault zone.

In an elongate coastal embayment, the depth-averaged momentum balance between pressure gradient and wind stress can be simplified as

\[ 0 = -g \frac{\partial \eta}{\partial x} + \frac{\tau}{\rho_0 H}, \]

where \( g \) is gravitational acceleration, \( H \) is water depth, \( x \) represents longitudinal coordinate, \( \rho_0 \) is reference density, and \( \tau \) is the along-basin wind stress. A number of terms are omitted in
Eq. (2). Winant (2004) theoretically predicted that the wind-driven flow is forced by the local wind stress and the down-wind water surface slope. Therefore, the along-bay wind stress \( t \) can be related to the mouth–head water level difference (Fig. 6a). During the winter period, the data sets from day 20 to 80 were chosen for analysis. The squared coherence between the local wind stress and the subtidal water level difference is about 0.9 at almost all low frequencies. This agrees with Ponte (2009), who showed high correlation between local wind stress and water surface slope during winter. The local wind is nearly in phase with the water level slope (0°, Fig. 6c). This implies that southeastward winds produce higher water level at the head of the basin. During the summer period, the data sets from days 190 to 270 were chosen for analysis. The relation between local wind stress and water level difference became less significant and the phase slightly increased. This is due to the stratification of water column during the summer period. Janzen and Wong (2002) pointed out that stratification reduces the effective depth over which the wind stress must act. Moreover, the linear dynamics for wind driven circulation (e.g. Eq. (2)) might be impaired by stratification.

The cold water intrusion, in turn, generates horizontal density gradients that enhance the contribution of baroclinic pressure gradients (Fig. 6b). Under stratification, Earth’s rotation could be important. During the stratified summer season, the internal Rossby radius of deformation estimated using the stratification inferred from the top-bottom temperature difference is about 3.5 km, which is smaller than its width indicating that Coriolis forcing is effective. For wind-induced circulation, the water depth is larger than twice the Ekman depth, allowing Earth’s rotation to be influential (Winant, 2004; Sanay and Valle-Levinson, 2005).

The eddy viscosity under stratification has an order of \( 1 \times 10^{-3} \) m²/s in Bahía Concepción (Ponte, 2009), thus Ekman depth is about 5.5 m, much smaller than the average depth. In addition, the relative importance of advection terms, especially the lateral advection, increases with reduction of vertical mixing (Cheng and Valle-Levinson, 2009). Therefore, effects of nonlinear processes, Earth’s rotation, and baroclinic pressure should be addressed in future theoretical considerations for the stratified summer season.

The water level of a coastal bay is forced primarily by remote along-shelf winds (Janzen and Wong, 2002). During the winter season, the squared coherence between remote winds and water elevation at the mouth is relatively high at certain frequencies (Fig. 6). The phases are about 70° at the correlated frequencies. This implies the coastal water level corresponds to remote winds with a time lag. During the summer season, the correlation between remote winds and coastal water level became less obvious, suggesting that local dynamics could contribute to the fluctuations of water level in the bay.

4.2. Density-driven circulation

Bahía Concepción is a wind-dominated coastal embayment without freshwater runoff. Due to the arid climate, evaporation rates are high inside Bahía Concepción and cause a saltier water body (e.g. Mendoza-Salgado et al., 2006; Winant and Gutiérrez de Velasco, 2003). This could potentially drive an inverse estuarine circulation. The horizontal density gradient can be derived from the CTD measurement at the head and the mouth of bay at a

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**Fig. 5.** Time series of subtidal water level at the mouth of the bay (a); the subtidal water level difference between the head and the mouth of the bay (b); numbers mark different periods of measurement and the along-shelf wind stress taking from the QuikSCAT data. Positive values of wind stress represent northwesterly wind.
depth of 5 m (squares shown in Fig. 1). Values of estimated longitudinal density difference and density gradient are shown in Fig. 7a and b. Positive values indicate higher density at the head of the bay than at the mouth. The density at the head is generally greater than that at the mouth. The largest difference is approximately 0.45 kg/m$^3$. The magnitude of the horizontal density gradient $\partial \bar{\rho} / \partial x$ ranges from 0 to $14 \times 10^{-6}$ kg/m$^3$, i.e., two orders of magnitude lower than typical values in estuaries (Fig. 7b). Nevertheless, the density-driven flow might be important in this bay.

Officer (1976) provided analytical solutions of density-driven ($u_d$) and wind-driven ($u_w$) flows, on the basis of the simplified balance between pressure gradient and stress divergence:

$$ u_d = \frac{gH^2}{48A_v \rho_0} \frac{\partial \rho}{\partial x} \left(1 - 9 \frac{x}{H} - 8 \frac{z^2}{H^2} \right), $$

and

$$ u_w = \frac{\tau H}{4A_v \rho_0} \left(1 + 4 \frac{x}{H} + 3 \frac{z^2}{H^2} \right), $$

where $z$ represents the vertical coordinate, and $A_v$ is a constant vertical eddy viscosity. Moreover, the relative importance of density-driven flow over wind-driven flow can be evaluated using the Wedderburn number ($W$), which may be defined as the ratio between the magnitude of $u_d$ and $u_w$. Taking $z$ as 0 (i.e., the magnitude of $u_d$ and $u_w$) in Eqs. (3) and (4), $W$ may be represented as

$$ W = \frac{u_d}{u_w} = \frac{gH^2}{12} \frac{\partial \rho}{\partial x}. $$

Using the calculated along-bay density gradient and observed longitudinal wind stress (Fig. 5c), $W$ is estimated in Fig. 7c. The absolute values of wind stress and longitudinal density gradient are used to evaluate the relative importance of density- and wind-driven flows. On average, the magnitude of the density-driven circulation is about 75% of the wind-driven flow. At certain periods, for example, around days 220 and 260 when density gradient is large and wind stress is very weak, the density-driven flow could be 1.5 times as important as the wind-driven flow. Therefore, the longitudinal density gradient acts as an important mechanism driving circulations in this arid bay.

4.3. Mixing/stratification mechanisms during the summer season

The level of stratification in the water column is a result of the competition between stirring and stratifying processes. In coastal environments, the major stirring agents are winds and tides while the major stratification mechanisms are heating/cooling, baroclinic circulation, and tidal straining. Simpson et al. (1990) defined a scalar parameter $\phi$ (units J m$^{-3}$), the potential energy anomaly, to determine the energy required to mix the water column completely:

$$ \phi = \frac{1}{\rho} \int_{-H}^{0} (p - \rho) g z dz, $$

where the overbar represents depth average. The time rate of change of $\phi$ indicates evolution of stratification. Simpson et al. (1990) provided an equation for the evolution of stratification, including the major mechanisms for mixing and stratification:

$$ \frac{d\phi}{dt} = \frac{z \rho Q}{2F_p} - \frac{4}{3\pi} \frac{4}{\rho} \frac{\partial \bar{C}_d \rho}{\partial x} \left( \delta \bar{C}_d \rho \right) \frac{U_f^3}{H} - \frac{1}{2} g^2 H^4 \left( \frac{\partial \rho}{\partial x} \right)^2. $$

The first term on the right represents the increase in stratification due to surface heating at a rate $Q$, while the second and the third terms denote stirring by a tidal current of amplitude $U_t$ and a wind of speed 10 m above surface $U_w$, respectively. In these terms, $e$ (equals 0.0037) and $d$ (equals 0.023) are the corresponding non-dimensional efficiencies of mixing, and $C_d$ (equals 0.0025) and $C_h$ (equals 6.4 $\times$ 10$^{-5}$) are the effective drag
coefficients for bottom and surface stresses (also non-dimensional). Furthermore, \( \alpha \) is the thermal expansion coefficient and \( \rho_a \) is the density of air (1.2 kg/m\(^3\)). The last term on the right represents stratification induced by density-driven flow. Contribution from tidal straining is excluded since this study concentrates on subtidal processes and the measurements available have been filtered to eliminate tides.

In applying this \( \phi \) equation, winds are treated as a mixing agent. However, it is recognized that wind-driven flow can create or destroy stratification depending on its interaction with the horizontal density gradient (Scully et al., 2005). Wind-driven flows enhance stratification when the wind blows with the longitudinal density gradient and reduce stratification when the wind blows in the opposite direction to the longitudinal density gradient. In order to include the contribution of wind-driven flows (wind straining) in the \( \phi \) equation, a parameter is developed following the framework of Simpson et al. (1990):

\[
\frac{\partial \phi_{\text{we}}}{\partial t} = \frac{g}{H} \frac{\partial \rho}{\partial x} \int_{-H}^{0} (u_w - \bar{u}_w) \, dz, \tag{8}
\]

where \( \phi_{\text{we}} \) is the potential energy anomaly produced by wind-driven flow. Substituting Eq. (4) into Eq. (8), and noting that \( \bar{u}_w = 0 \), results in

\[
\frac{d \phi_{\text{we}}}{dt} = \frac{1}{48} \frac{g H^2 \bar{\rho}}{A \rho} \frac{\partial \rho}{\partial x}. \tag{9}
\]

It is clear that when \( \tau \) and \( \partial \rho / \partial x \) have opposite signs, the wind-driven flow decreases stratification (negative \( d \phi / dt \)). This is consistent with previous studies (e.g. Scully et al., 2005; Li et al., 2008).

The data sets for heat flux, winds, temperature, and current in the summer season are used to evaluate the relative importance of the five mixing/stratification processes (Fig. 7). All data sets are filtered with a Lanczos filter at half-power of 34 h in order to remove tidal influences. The time dependent \( A_v \) is estimated with the closure proposed by Pacanowski and Philander (1981):

\[
A_v = 0.01(1 + 5 Ri)^{-2} + 10^{-4}, \tag{10}
\]

where \( Ri \) is a gradient Richardson number.

Mixing is relatively weak during the summer season (Fig. 8a and b). The potential energy anomalies induced by wind stirring and tidal stirring are at least one order of magnitude smaller than the other mechanisms. The evolution of tidal mixing follows the tidal range and the weak tidal mixing results from the microtides. Surface heating is a major process enhancing stratification (Fig. 8c). Density-driven circulation contributes to increase stratification and has a similar magnitude to that of surface heating (Fig. 8d). Wind-driven flow is another important mixing/stratification mechanism (Fig. 8e). The potential energy anomaly induced by wind straining has the same order of magnitude as surface heating and density-driven flow. In particular, during strong wind events, wind straining is the dominant mechanism for producing stratification. Also, as a mixing agent, wind straining acts more effectively than tidal and wind stirring. This suggests that wind straining can play a dominant role in mixing/stratification for coastal basins.

![Fig. 7. Longitudinal density difference (a), density gradient (b), and Wedderburn number (c).](image-url)
through Project OCE-0726697. AVL also acknowledges the complementation of the stratification mechanism in wind-dominated coastal basins. This study is funded by the US National Science Foundation under Grant OCE-0726697.

5. Conclusions

Observations in Bahía Concepción demonstrate, for the first time in a bay of the Gulf of California, the role of wind-driven upwelling in stratifying a bay where freshwater input is negligible. Negative buoyancy supplied by coastal upwelling can contribute to seasonal stratification in coastal embayments, in addition to the expected surface heat transfer from the atmosphere. Onset of summer upwelling winds triggered a cold water intrusion that dropped the water temperature in the entire water column. Persistent upwelling-favorable winds maintained the supply of cold water into the bay, which increased the temperature stratification as the water surface warmed up. Therefore, the seasonal evolution of stratification is induced in part by increased atmospheric heat input and in part by cold water intrusions drawn by upwelling. Also, the wind-driven upwelling maintains a longitudinal temperature difference which produces a baroclinic exchange flow. Spatial patterns of temperature revealed by EOF analysis indicate section-wide warming/cooling as depicted by the first two modes. The first two principal components contain signature of seasonal heat transfer from the atmosphere and of cold water intrusions. Strong evaporation leads to a density difference between the bay and adjacent coastal water resulting in an inverse gravitational circulation that has the same order of magnitude as the wind-driven circulation. Also, it is found that wind straining could be the dominant mixing/stratification mechanism in wind-dominated coastal basins.

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References