Seasonal variability of temperature, salinity, velocity, vorticity and sea level in the central Gulf of California, as inferred from historical data

By P. RIPA and S. G. MARINONE

Oceanología, CICESE, México

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SUMMARY

Historical data from 17 cruises into the Guaymas basin, central part of the Gulf of California, are globally analysed. Temperature, across-gulf integrated velocity, sea level, and salinity have a significant seasonal cycle. The last variable has an important semi-annual component, presenting four extremes per year. The others are mainly annual, reaching their maximum values in summer and their minima in winter.

By interaction with the atmosphere, the central gulf is gaining heat and losing fresh water during the whole year, on a monthly-mean time scale (the only exception is an unimportant gain of fresh water in August). This implies net heat and salinity fluxes into the open ocean, through the gulf’s mouth.

Temperature and salinity vertical gradients at the surface are well correlated with atmospheric forcing, through turbulent diffusion of heat and salt. Average velocity correlates with the along-gulf wind stress with zero lag, as if frictionally dragged in a surface layer, with a short relaxation time. Seasonal sea-level changes are mainly caused by the thermal expansion of the water column.

Horizontal processes (advection and diffusion) are as important as, or more important than, vertical diffusion for both heat and salt balances in the upper layers.

Vorticity, the across-gulf derivative of the velocity field, does not show a seasonal cycle. Its contribution to the instantaneous geostrophic velocity and surface elevation near each coast is larger than that due to the across-gulf average velocity.

1. INTRODUCTION

The Gulf of California is a beautiful natural blue box for the observation and modelling of ocean physics. Almost half a century has elapsed since the first modern oceanographic cruise into its waters (Sverdrup 1941) and about two dozen more have been made since then, collecting data at different stations along its megametre extent (see for instance Roden 1964; Alvarez-Borrego 1983, and references therein). This information is by no means exhaustive in either time, depth or transverse and longitudinal directions, but it is, nevertheless, quite complex, revealing the wealth of processes taking place in the gulf.

In this paper, we use the simplest possible longitudinal resolution, namely, a single section in the central part, with the intention of extracting robust patterns of temporal evolution. Two important time scales are investigated: the seasonal and the interannual (Baumgartner and Christensen 1985; Robles and Marinone 1987; Bray 1988; Marinone 1988), the latter mainly restricted to the study of the effects of El Niño phenomena. Here, we concentrate on the seasonal time scale, leaving for another paper, which we hope to finish some day, the analysis of the interannual and sub-seasonal anomalies.

So far, statements on seasonality, or lack of it, have been qualitative in most cases, based on visual inspection of pieces of information from different months, usually from diverse years (e.g. Rosas-Cota 1977; Badan-Dangon et al. 1985; Robles and Marinone 1987; Marinone and Ripa 1988; Bray 1988). This procedure might be sufficient for variables with a strong seasonal signal (like temperature, near-surface transport, and sea level) but it can be misleading for other variables, like salinity and vorticity, given the poverty of the temporal sampling. Bray agrees with us on the seasonality of the former variables, but, with respect to salinity and vorticity, she reaches conclusions opposite to those of this work. We believe ours to be more trustworthy, for they are based on a more rigorous statistical analysis.
Salinity has an important semi-annual component, making the estimation of its seasonal cycle more difficult, and furthermore shows a relatively stronger El Niño response. Vorticity, on the other hand, is rather important, reflecting the presence of basin-wide gyres (Badan-Dangon et al. 1985; J. M. Figueruela, personal communication, 1987), and has proved to be highly variable, probably even on the sub-seasonal time scale. Consequently, in this work we have made an effort to obtain quantitative results concerning the seasonality of each variable; statistical criteria were found to be very useful, but not always sufficient.

The following section and the appendices are dedicated to the description of the data base and the methods used for its analysis. Temperature, salinity, velocity, vorticity and sea level are studied in sections 3, 4, 5, 6 and 7 respectively. A global description of the variables and the conclusions of this work are presented in section 8.

2. Data base and Analysis

This work is based on the analysis of 86 hydrographic casts from 17 transects going roughly from Santa Rosalia to Guaymas in the central part of the Gulf of California, taken between March 1939 and October 1983, see Fig. 1. A table with dates, maximum depths, number of casts and ship names is presented in Marinone and Ripa (1988) (the cruise numbering is different from that used here). The same data base is described, and analysed in terms of temperature–salinity diagrams, by Robles and Marinone (1987).

The space–time resolution is quite poor. The temporal sampling is illustrated in Fig. 2: it is quite heterogeneous on a yearly basis, and the annual cycle is not resolved too well (particularly for the trimesters of July-September and November-January, with only one cruise in each). Four of the transects—indicated by arrows in Fig. 2—were made at times when El Niño effects were detected at the Guaymas basin (Robles and Marinone 1987); exclusion of these cruises, as in some of the analysis here, renders the temporal sampling even poorer. On the other hand, 59 of the 86 casts reach a depth of 500 m and 38 of them go down to at least 700 m. Table 1 shows how many cruises reached certain depths and had a certain minimum number of casts; at least two stations are needed in order to compute geostrophic flow, and three for its across-gulf derivative, the vorticity.

Different mathematical operations, described below, were performed on the data bank, in order to enhance the statistical robustness of its information and to be able to treat the different cruises on a similar footing.

In the first place, we reduced the horizontal structure of the temperature $T$ and salinity $S$ fields, for each cruise (denoted by time $t$) and as a function of depth $z$, to the polynomials

$$
\begin{align*}
T(x, z, t) &\approx T(z, t) + xT_x(z, t) + \frac{1}{2}x^2T_{xx}(z, t) \\
S(x, z, t) &\approx S(z, t) + xS_x(z, t) + \frac{1}{2}x^2S_{xx}(z, t)
\end{align*}
$$

(1)

where $x$ is the across-gulf coordinate, with origin at the mid-point. The coefficients were calculated by least-squares fitting when the number of casts reaching depth $z$ for cruise $t$ was four or more. If the number was three or two, they were fitted exactly by a parabola or a straight line, respectively. With only two points, at a certain $(z, t)$, the coefficients $T_{xx}$ and $S_{xx}$ are absent; this possibility had to be taken into account in the subsequent analysis.

The coefficients $T_x$, $S_x$, $T_{xx}$ and $S_{xx}$ were not analysed as such, but rather used in the thermal wind relation

$$
f \frac{\partial v}{\partial z} = g[\alpha(z)\partial T/\partial x + \beta(z)\partial S/\partial x]
$$

(2)
to calculate the along-gulf current (positive towards the head) as

$$v(x, z, t) = V(z, t) + x\Omega(z, t).$$

(3)

(The expression between curly brackets in (2) is equal to the horizontal density gradient divided by the mean density.)

Figure 1. Stations occupied in the different cruises in the Guaymas basin. The bathymetry, in metres, is shown by broken lines.
Figure 2. Dates on which the data were collected. Here and in the rest of text and figures, the numbers label the different cruises and arrows point to those made during an El Niño event. The histogram indicates the number of sections made in each month; shaded squares are excluded when El Niño cruises are omitted.

<table>
<thead>
<tr>
<th>Maximum depth</th>
<th>Casts</th>
<th>Any</th>
<th>≥500 m</th>
<th>≥700 m</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>≥1</td>
<td>17</td>
<td>16</td>
<td>12</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(13)</td>
<td>(13)</td>
<td>(10)</td>
</tr>
<tr>
<td></td>
<td>≥2</td>
<td>14</td>
<td>11</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>(12)</td>
<td>(10)</td>
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</tr>
<tr>
<td></td>
<td>≥3</td>
<td>12</td>
<td>8</td>
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<td></td>
<td>(10)</td>
<td>(7)</td>
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</tbody>
</table>

Velocity or vorticity calculations require at least 2 or 3 casts, respectively. Figures in parentheses are obtained excluding data taken during El Niño events.
Henceforth we define 'temperature', 'salinity' and 'velocity' as the fields $T(z, t)$, $S(z, t)$ and $V(z, t)$, which are very close to the across-gulf averages, and 'vorticity' as $\Omega(z, t)$. The latter is presumably the main contribution to the relative vorticity, since scaling arguments suggest that $|\partial u/\partial x| > |\partial u/\partial y|$; in any case, hereafter 'vorticity' refers to $\partial v/\partial x$ since $u$ cannot be evaluated with these data.

The next step was to analyse these fields into principal components (or empirical orthogonal modes), i.e. make

$$\Gamma(z, t) = \langle \Gamma(z) \rangle + \sum_{\alpha} A_{\alpha}(z) B_{\alpha}(t)$$

(4)

where $\Gamma$ denotes $T$, $S$, $V$ or $\Omega$, the angle brackets indicate time average and the modes are chosen orthogonal, in the sense

$$\int_{-h}^{0} A_{\alpha}(z) A_{\beta}(z) dz = \lambda_{\alpha} \delta_{\alpha\beta}, \quad \langle B_{\alpha} B_{\beta} \rangle = \delta_{\alpha\beta}$$

(5)

and ordered by decreasing variance

$$\lambda_{1} \geq \lambda_{2} \geq \ldots \geq 0.$$  

(6)

The total depth $h$ was chosen as 700 m or 500 m, in order to have results for a sufficient number of cruises, for their temporal analysis, described next. We make no assumptions on either thermohaline variability or flow below that depth. Typically data from some cruises had to be excluded from the principal component analysis either because the casts were too shallow (see Table 1) or were made during El Niño events. Then, a posteriori, we used the structure functions $A_{1}(z)$, $A_{2}(z)$ and $A_{3}(z)$, along with the information from those cruises, say, at times $t_i$, to estimate the coefficients $B_{1}(t_i)$, $B_{2}(t_i)$ and $B_{3}(t_i)$, in a least-squares sense.

Each principal component provides an independent mode of coherent variability (i.e. simultaneous at all depths). It is difficult, however, to perceive its physical meaning, unless the corresponding $B_{\alpha}(t)$ takes peculiar values (like a clear seasonal signature independent of the cruise's year; extremes during El Niño events; etc.), mainly owing to the irregularity of the temporal sampling. A more direct analysis was performed independently, with the purpose of extracting the seasonal signal, by least-squares fits of the form

$$\Gamma(t) \sim \Gamma_0 + \Gamma_1 \cos(\omega t) + \Gamma_2 \sin(\omega t) + \Gamma_3 \cos(2\omega t) + \Gamma_4 \sin(2\omega t)$$

(7)

where $2\pi/\omega$ equals one year. Here $\Gamma$ denotes $T$, $S$, $V$ or $\Omega$ at a certain depth, any flux through the air–sea interface, or even any of the $B_{\alpha}$ products of the previous analysis. In some cases, the semi-annual component had to be included in order to get a significant fit; its importance is measured by the relative variance

$$\gamma = (\Gamma_3^2 + \Gamma_4^2)/(\Gamma_1^2 + \Gamma_2^2 + \Gamma_3^2 + \Gamma_4^2).$$

(8)

The cosine and sine functions are not orthogonal in the set of dates used for the fit, because the temporal sampling is non-homogeneous. As a consequence, the residuals can be very small—which is what the least-squares algorithm is required to achieve—at the expense of too large coefficients. Care should be exerted before accepting some fit just because the explained variance is small in some absolute sense. The relevance of each fit was therefore assessed by comparing the variance explained with that corresponding to random data: minimum explained variances for different confidence levels (CL) were determined by Monte Carlo simulations with the same dates, which usually depend upon the depth and type of variable (see appendix A).
3. Temperature

The mean temperature profile \(\langle T \rangle\) is shown in Fig. 3(a): it decreases monotonically from 21 °C at 25 m to 6.5 °C at 650 m (more precisely, for the 0–50 m and 600–700 m layers, respectively). Also shown are the curves of \(\langle T \rangle \pm A_1\), namely, mean plus the contribution of the first mode at the times when \(B_1\) equals ±1; when \(B_1 = 0\) the temperature field is that of \(\langle T \rangle\); when \(B_1 = \pm 1\), \(T\) is half way between the solid line and one of the thin curves (neglecting higher modes). \(A_1(z)\) is everywhere positive: positive (negative) values of \(B_1\) represent a warming (cooling) of the whole water column. The actual values of \(B_1(t)\) are presented as crosses in Fig. 3(b): a seasonal fit of the form (7) explains 93% of its variance, which is significant at the 99% confidence level. The semi-annual component is relatively unimportant (\(\gamma = 0.01\)), and a fit excluding this component is much more significant in the statistical sense. Empirical modes calculated down to 500 m, with 13 cruises instead of 10, give essentially the same results for the vertical and temporal structure of the first mode, as well as for its seasonal fit. The \textit{a posteriori}

![Diagram](image)

Figure 3. (a) Mean temperature profile (solid line) and the contribution of the first empirical mode (93% of the variance), excluding El Niño cruises. The thin lines show the mean plus and minus one standard deviation. Equivalently, they represent the temperature profile at those times when the temporal structure function (b) equals ±1. (b) Temporal structure function, \(B_1\). The amplitudes of the first mode for the cruises used in the analysis are shown as crosses. Some cruises were not included in the principal component analysis, but their amplitudes were estimated \textit{a posteriori}; the circled dots indicate the estimated value of the temporal amplitude.
estimated values of $B_1$ for those cruises not included in the principal component analysis, being either too shallow or corresponding to El Niño years, are shown with circled dots: they are scattered around the structure drawn by the other points, reinforcing the physical significance of the first mode. Figure 3 shows clearly the seasonal warming and cooling of the upper layers.

The first, second and third modes explain 96.7%, 1.8% and 1.1% of the temperature variance in the upper 700 m. The values of $B_2$ (not shown) do not present a seasonal pattern, indicating some non-seasonal variability, not necessarily interannual: transects 13 and 17 were made only 7 days apart, yet they show important differences both in the thermohaline structure (Robles and Marinone 1987) and the inferred geostrophic circulation pattern (Marinone and Ripa 1988). Even though it represents such a small part of the total variance, the third mode, on the contrary, does have a pattern of seasonal variation (significant at the 95% CL), which could mean that the warming and cooling does not occur exactly in phase throughout the water column. The second and third modes will not be considered further, because of their small variance.

These principal components correspond to non-El-Niño years (10 cruises); for an analysis with all 12 transects, the variance explained by the first mode is reduced to 89.9%, and a seasonal fit explains 78% of its temporal structure, which is significant at the 97% CL. Of the seven cruises not included in the previous analysis (circles in Fig. 3(b)), those numbered 5, 9, 16 and 12 were made during El Niño events; the first three present temperatures greater than the seasonal cycle in accordance with the intrusion of warm waters of equatorial origin during this event (Robles and Marinone 1987). The largest anomaly is that of cruise 5, which was made at the trough of the seasonal cycle; cruise 12, on the other hand, was performed at the peak of the seasonal cycle and is, indeed, colder than the interpolated seasonal cycle.

The seasonal fits of temperature at each depth are shown in Fig. 4, and their statistical significance is presented in Table 2. There is a slight delay of the hottest day of the year with depth, from 0 to 100 m, and a larger lag for the coldest one. The surface signal goes from a maximum of 33 °C in late August to a minimum of 16 °C in late January; thus the warming phase is about two months longer than the cooling one. Below 250 m the

![Figure 4. Seasonal signal of temperature at the surface and selected layers. The statistical significance of the fit is presented in Table 2.](image-url)
amplitude is about one degree Celsius or less, and indeed the fit is no longer significant at the 90% CL.

Temperature variations are governed by the heat balance equation

\[ \rho c_p \frac{\partial T}{\partial t} = -\frac{\partial F}{\partial z} - \frac{\partial G}{\partial y} \]  \hspace{1cm} (9)

where \( F \) and \( G \) represent the upward and headward heat flux components in the gulf. The horizontal part includes both advective and diffusive effects and cannot be calculated with data from the Guaymas section alone. At the air–water interface \(-F\) equals \( Q \), the net heat flux into the ocean, which was calculated using climatological data from Guaymas (on the coast of mainland Mexico) corresponding to the period 1941 through 1970, and bulk formulae. \( Q \), shown by the solid curve in Fig. 5, is always positive; the ocean gains heat from the atmosphere all year round, varying from 10 W m\(^{-2}\) in early December to 220 W m\(^{-2}\) in July. This net flux is the sum of those due to short-wave incoming radiation (which varies in the range of 130 to 310 W m\(^{-2}\)), latent heat (\(-85\) to \(-40\) W m\(^{-2}\)), long-wave outgoing radiation (\(-70\) to \(-45\) W m\(^{-2}\)) and sensible heat flux (0 to 20 W m\(^{-2}\)). The net heat flux is also positive in the northern part of the gulf (Organista 1987; Bray 1988).

The atmosphere–ocean heat flux was related to the vertical gradient of temperature, evaluated at the surface, through the turbulence formula

\[ Q \approx K \rho c_p \left. \frac{\partial T}{\partial z} \right|_o \]  \hspace{1cm} (10)

where \( K \) is a vertical eddy diffusion coefficient. (The temperature gradient was estimated by finite difference between the value of \( T \) at the surface and that corresponding to the 0–50 m layer.) A dispersion diagram of \( \rho c_p (\partial T/\partial z)_o \) for each cruise \( \nu \), the climatological value of \( Q \) which corresponds to its date is shown by the circles in Fig. 6: there is indeed a positive correlation. The solid curve with the months' initials was drawn using the seasonal representation type (7) for both \( Q \) and \( \rho c_p (\partial T/\partial z)_o \). With this smooth para-
Figure 5. Solid curve: incoming heat flux calculated with climatological mean of atmospheric variables and sea surface temperature. Dashed line: estimated by the turbulence formula, with the best estimate of the diffusion coefficient. Dotted curve: downward heat flux at 50 m depth. The three curves represent seasonal fits.

Figure 6. Dispersion plot of climatological incoming heat flux and the vertical temperature gradient evaluated at the surface. The circles represent individual cruises and the continuous line was drawn using the seasonal smooth representation for both variables. The slope of the straight line, with the error bar, gives the best estimate of the diffusion coefficient.

metrization, the dispersion is smaller: a least-squares fit to Eq. (10) gives

$$ K = (3.2 \pm 0.1) \times 10^{-4} \text{ m}^2 \text{ s}^{-1} $$

(11)

which is well within the typical oceanic range. This is a very encouraging result, particularly given the poor temporal sampling of our data. This value of $K$ is indicated in Fig. 6 by a straight line through the origin, with an error bar for its slope. The larger
dispersion of the circles from that line might be caused by having used a climatological expression for $Q$ instead of its actual value at the time of each cruise (which was not available).

The right-hand side of (10), with the optimum value for $K$ given by (11), is presented by a dashed line in Fig. 5. Its correlation with $Q$ is quite evident: the regression explains 94% of the variability, which is significant at the 98% CL (see appendix B for $N = 3$). It is interesting that the fit is even better if one lets the temperature gradient have a lag with the heat flux of the order of one month. The correlation is worse during winter, when $Q$ passes through a non-vanishing minimum, but the upper 50 m are well mixed (see Fig. 4) giving a zero value for $(\partial T/\partial z)_0$.

We then use the parametrization

$$F = \begin{cases} -Q & \text{for } z = 0 \\ -K\rho c_p \partial T/\partial z & \text{for } z < 0. \end{cases} \quad (12)$$

The vertical heat flux at 50 m (dotted line in Fig. 5) has a smaller amplitude than that at the surface and is retarded about a month from it. The difference between the fluxes at the surface and at 50 m is the vertical contribution to the heat storage in the upper 50 m, indicated by a dashed line in Fig. 7. The solid line in Fig. 7 represents the heat storage for 0–50 m and, therefore, the distance between the two curves is equal to the horizontal convergence, $-\partial G/\partial y$. The latter is very similar to the vertical contribution, both in amplitude and phase. In other words, the heating of the upper 50 m, which varies from a minimum of $-6$ W m$^{-3}$ in November to a maximum of 4 W m$^{-3}$ in July, is equally due to vertical diffusion and horizontal advection–diffusion.

![Figure 7. Solid line: heat storage in the upper 50 m. Dashed line: difference in the vertical heat flux at the top and bottom of the layer. The distance between the curves (shaded) is the contribution of horizontal processes, like advection and diffusion.](image)

The picture is not so simple in the 50–100 m layer (Fig. 8) because the heat storage seems to have two maxima and two minima per year. We are not very confident of the details of this figure, but at least we can see once again that vertical and horizontal divergences of heat flux are equally important.
The importance of the vertical component of heat flux alone can be understood by studying the solution of a 'one-dimensional' model, i.e. one without horizontal advection or diffusion: Let $T'(z, t)$ be the annual component of the temperature fluctuation. Solving

$$\frac{\partial T'(z, t)}{\partial t} = K \frac{\partial^2 T'(z, t)}{\partial z^2}$$

subject to

$$K \rho c_p \frac{\partial T'(0, t)}{\partial z} = Q_o \cos(\omega t)$$

results in

$$T'(z, t) = T_o \exp(z/H) \cos(\omega t - \frac{1}{2} \pi + z/H)$$

with

$$H = \sqrt{(2K/\omega)} \quad \text{and} \quad K \rho c_p T_o = Q_o H/\sqrt{2}.$$ 

Thus the surface temperature fluctuation has an amplitude $T_o$ and is one quarter of a period (three months) delayed from the surface flux. The total heat fluctuation is equal to that of a homogeneous layer with depth $H$ and temperature

$$\langle T'(t) \rangle = H^{-1} \int_{-\infty}^{0} T'(z, t) dz = (T_o/\sqrt{2}) \cos(\omega t - \frac{1}{2} \pi)$$

i.e. with an amplitude $T_o/\sqrt{2}$ and delayed one half period (six months) from the surface heating. A value of $K = 3.2 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ gives $H = 57 \text{ m}$; moreover, with $Q_o = 105 \text{ W m}^{-2}$ we have $T_o = 3.2 \text{ degC}$ and $T_o/\sqrt{2} = 2.3 \text{ degC}$. The observed seasonal cycle has amplitudes, for the surface and the 0–50 m layer, which are larger than those predicted by the one-dimensional model by a factor of 2.7 and 3.0, respectively (see Fig. 4). Moreover, the phase lag with the heat flux at the air–sea interface is only one to two months, instead of the 1½ to 3 months predicted by the model. The heat storage in the upper 50 m has an amplitude of 5 W m$^{-3}$ (see Fig. 7), which is 2.7 times larger than $Q_o/H (= 1.85 \text{ W m}^{-3})$, and the phase is again incorrect. The semi-annual component may be partially responsible for the discrepancies between model and observation, particularly for the heat storage signal, but the main lesson of this exercise is that horizontal advection and diffusion are as important as, or more important than, the vertical component in the seasonal heat balance.
4. Salinity

The mean salinity profile \(\langle S \rangle\) is shown in Fig. 9(a): it decreases monotonically as does temperature in Fig. 3(a)—from 35.25 psu at 25 m to 34.55 psu at 650 m. (Individual casts are not necessarily monotonic.) Density is, of course, stably stratified, because \(\alpha \Delta T/\beta \Delta S = 7\). Extracting a statistically and physically significant seasonal signal for salinity proved to be harder than for temperature: an account of the work follows.

Salinity fluctuations in time and depth were analysed by principal component analysis down to 700 m (12 cruises) and 500 m (16 cruises). The three gravest modes explain 66%, 17% and 10% of the total variance in both cases. Moreover, the structures of the first mode in both depth (Fig. 9(a)) and time (Fig. 9(b)) are essentially identical in the overlapping part of the water column and for all the cruises. Unlike temperature (Fig. 3), \(A_1(z)\) has a node at about 150 m, i.e. salinity variations below and above this level are in opposition.

![Figure 9](image.png)

**Figure 9.** As Fig. 3, for salinity, including all cruises that reach 700 m; analysis down to just 500 m gives essentially the same results. The first mode explains 66% of the variance. Solid (dashed) line in (b) shows the annual (annual plus semi-annual) fit to the temporal amplitudes for all 17 cruises.

The main difference observed between the two analyses is in the seasonal fit to \(B_1(t)\), if it is done using only the values that result from the empirical modes decomposition: for the 700 m points, squares in Fig. 9(b), Eq. (7) explains 81% of the variance, and the Monte Carlo calculation indicates this to be significant at the 95% CL. However, this good fit (not shown) is obtained at the expense of a large negative excursion in July—August—September, which is a trimester devoid of casts deep enough for that analysis; a similar problem is found in the analysis of the vorticity data, in section 6. The tongue in the fitted function—right-hand side of (7)—is found to be excessive on both physical and statistical grounds. In the first place, it predicts salinities which are too low for the Guaymas basin and the \textit{a posteriori} estimated value of \(B_1\) for cruise 12, August 1957, is much saltier than this interpolated plunge (the mispredicted minimum is \(B_1 \approx -5.5\) for a normal August). This cruise was made during an El Niño event, which brings waters of tropical origin to the gulf, that are fresher than normal; in addition, in 1957 it rained much less than usual, which should further increase the salinity (Robles and Marinone 1987). In the second place, the variance of the fitted function \((\Gamma_1^2 + \Gamma_2^2 + \Gamma_3^2 + \Gamma_4^2)\) is much larger (by a factor of seven) than that of the original data points—left-hand side of (7). The heart of the problem is that the fitting functions are not orthogonal in the set.
of dates used, and then the least-squares algorithm does what it is required to do—minimize the residues—at the expense of coefficients which are excessively large. (A similar problem can occur, for instance, if one attempts to calculate two semi-diurnal tidal constituents with a record which is too short to discriminate between the two frequencies.)

Therefore, we had to include all other a posteriori estimated values of $B_1$, circles in Fig. 9(b), before doing the seasonal fit. Alternatively, we can use the $B_1$ values from the 500 m analysis (only cruise 16 has to be estimated a posteriori in this case). The two (augmented) sets of 17 points $B_1(t_1)$ are practically identical and so is, therefore, their seasonal fit. This is shown by the dashed curve in Fig. 9(b): the variance explained equals 44%, which is significant at the 88% CL. The semi-annual component is important, $\gamma = 0.26$; and $(\Gamma_1^2 + \Gamma_2^2 + \Gamma_3^2 + \Gamma_4^2)$ equals 1.3 times the variance of the fitted points (i.e. there are no spurious interpolations). The solid curve is obtained excluding the semi-annual components from the fit: the explained variance is then 26%, which is significant at the 87% CL, and the ratio of variances decreases to 0.4. Cruises 5, 9 and 16 present large fresh anomalies, which is fine, because they correspond to data taken during El Niño events. But then, cruises 8 and 15, from ‘normal’ years, also show large anomalies, of both signs. Finally, cruise 12 (the one that prevents the large tongue in August) is still saltier than the seasonal fit including semi-annual constituents (dashed curve). Although we believe this to be a reasonable estimate of the seasonal signal, clearly we need more data from the trimester July–August–September of ‘normal’ years, in order to reinforce the significance of the fit, as well as that of others in this section.

The seasonal signals of salinity at each depth are shown in Fig. 10 and their statistical significance is again presented in Table 2. The curves correspond to analysis of all cruises; exclusion of El Niño data seems to improve the fit (columns at the right in Table 2), but at the expense of too low salinities in August for the top hundred metres, as for the first empirical mode. Values at the very surface are not very significant. In the top fifty metres, the minimum salinity (34.9 psu) occurs in August and the maximum (35.5 psu) in December; there is a secondary minimum in March and a secondary maximum in May, evidence of an important semi-annual component. Notice that this structure is very similar to that of $B_1(t)$ in Fig. 9(b). Below the surface, a similar pattern is observed, with

![Figure 10. As Fig. 4, for salinity, including all cruises that reach each depth.](image-url)
Figure 11. Climatological monthly mean of evaporation and precipitation at Guaymas. The solid curve is the seasonal fit to the evaporation excess.

a smaller amplitude and roughly the same phase, down to 150 m where a reversal of phase occurs (maximum salinity in August), in agreement with the zero crossing of $A_1(z)$. Fits are not significant around the reversal depth (see Table 2).

Salinity variations are governed by the salt balance equation

$$\rho \beta \frac{\partial S}{\partial t} = -\frac{\partial F}{\partial z} - \frac{\partial G}{\partial y}$$  \hspace{1cm} (18)

where $F$ and $G$ now represent the upward and headward salt flux components in the gulf. At the air–water interface $-F$ equals $\rho \beta S(E - P)$, where $E$ and $P$ are the evaporation and precipitation rates. The climatological monthly means of $E$ and $P$, calculated from the same source as $Q$, are shown in Fig. 11. Evaporation shows a very strong semi-annual pattern, with maxima of the order of 1 m y$^{-1}$, in April and November, and minima of roughly half that value, in February and August. The precipitation is practically nil from April to June, has a broad and large peak of 0.75 m y$^{-1}$ in August and a weaker one in December. The resulting $E-P$ signal, fitted to a seasonal expression of type (7), is presented by the solid line in Fig. 11: the net $E-P$ in one year equals 58 cm and the fluctuation is strongly semi-annual ($\gamma = 0.65$). We related this flux to the vertical salinity gradient at the surface through the turbulence formula

$$E - P = (K/(S)) \frac{\partial S}{\partial z}$$  \hspace{1cm} (19)

where $K$ is again a vertical eddy diffusion coefficient, and the average surface salinity, $\langle S \rangle$, is 35.28 psu. (The salinity gradient at the surface was estimated by a straight line fit to the value at $z = 0$ and those corresponding to the 0–50 m and 50–100 m layers.) A dispersion diagram of $\langle S \rangle^{-1} \frac{\partial S}{\partial z}$ for each cruise $v$, the climatological value of $E-P$ which corresponds to its date is shown by circles in Fig. 12: there is positive correlation for all non-El-Niño cruises; those circles outside the first quadrant (numbers 5, 9, 12 and 16) correspond precisely to transections made during El Niño events. The solid curve with the months’ initials was drawn using the seasonal representation type (7) for both $E-P$
and \(\langle S \rangle^{-1}\partial S/\partial z\). With this smooth parametrization, the dispersion is smaller: a least-squares fit to Eq. (19) gives

\[
K = (2.2 \pm 0.4) \times 10^{-4} \text{ m}^2 \text{ s}^{-1}
\]

which is again in the typical oceanic range. Equation (19) explains 68% of the variability (see Fig. 13), which is significant at the 97% CL (see appendix B for \(N = 5\)).

Values from Eqs. (11) and (20) are a bit incompatible, i.e. their uncertainties are underestimated. Increasing them accordingly and calculating a weighted average we get \(K = (3.1 \pm 0.3) \times 10^{-4} \text{ m}^2 \text{ s}^{-1}\); a 10% error in the estimation of an eddy diffusion coefficient is more than we expected.
We then use the parametrization

$$F = \begin{cases} -\rho \beta (S - P) & \text{for } z = 0 \\ -K\rho \beta \partial S / \partial z & \text{for } z < 0. \end{cases} \tag{21}$$

Figure 14 shows the salt storage and the convergence of the vertical salt flux for the upper 50 m. The difference between the curves gives the horizontal convergence, $-\partial G / \partial y$, that closes the salt balance. It seems to be of the same order as, or even more important than, the vertical one. (This figure was drawn using all cruises. The vertical flux divergence is little sensitive to the inclusion or not of El Niño cruises; that is not the case for the salt storage curve, which has a much larger amplitude if the August point, cruise 12, is not included in the fit.)

![Figure 14. As Fig. 7, for the salt storage.](image)

As with temperature, we can study the importance of the vertical component of salt flux alone with the analytical solution of a similar 'one-dimensional' model. The annual component of the salinity fluctuation $S'(z, t)$ satisfies

$$\frac{\partial S'(z, t)}{\partial t} = K \frac{\partial^2 S'(z, t)}{\partial z^2} \tag{22}$$

with the top boundary condition

$$K \frac{\partial S'(0, t)}{\partial z} = \langle S \rangle (E - P)_o \cos (\omega t + \delta); \tag{23}$$

the evolution of the semi-annual component is governed by the same equations with different values for the amplitude $(E - P)_o$ and phase $\delta$ of the fresh water flux. The solution is as in Eqs. (15) and (17), with $T$ replaced by $S$, where $H$ is given by (16) and $K S_o = \langle S \rangle (E - P)_o H/\sqrt{2}$. Appropriate values for the annual component are $(E - P)_o = 8.94 \times 10^{-9} \text{ m s}^{-1} (= 28 \text{ cm y})$ and $H = 56 \text{ m}$. For the semi-annual one, on the other hand, $(E - P)_o = 1.21 \times 10^{-8} \text{ m s}^{-1} (= 38 \text{ cm y})$ and $H = 39 \text{ m}$. This results in $S_o = 0.04 \text{ psu}$ and $S_o/\sqrt{2} = 0.03 \text{ psu}$, for each component. The observed seasonal cycles have amplitudes of 0.17 psu and 0.14 psu, for the 0-50 m layer, which are six or five times larger than those predicted by the one-dimensional model (see Fig. 10). The salt storage in the upper 50 m has amplitudes of $2.6 \times 10^{-8} \text{ kg m}^{-3} \text{ s}^{-1}$ and $4.2 \times 10^{-8} \text{ kg m}^{-3} \text{ s}^{-1}$ (see Fig. 14), which are five to six times larger than $\rho \beta (S)(E - P)_o / H (=4 \times 10^{-9} \text{ kg m}^{-3} \text{ s}^{-1}$ and $8 \times 10^{-9} \text{ kg m}^{-3} \text{ s}^{-1}$): equivalently, horizontal advection and diffusion are as important as—or more important than—in the vertical in the seasonal salt balance.
5. VELOCITY

A flow field of the form \( v(x, z, t) = V(z, t) + x \Omega(z, t) \) was calculated using the thermal wind relation (2) and either an assumption of a level \( z_0 \) of no motion (Fomin 1984)

\[
V(z_0, t) = \Omega(z_0, t) = 0
\]  
(24)

or assigning values to both \( V(z_0, t) \) and \( \Omega(z_0, t) \) (i.e. two numbers per cruise) with the least-kinetic-energy criterion:

\[
\text{minimum} \int \int \{V(z, t) + x \Omega(z, t)\}^2 dx dz.
\]  
(25)

The values of \( V(z_0, t) \) and \( \Omega(z_0, t) \) are obtained in the last case by inverting a 2\times2 matrix for each cruise. The range of \( x \) in (25) is a function of \( z \). The integration depths might be different for \( V \) and \( \Omega \) allowing for maximum use of the data (see Table 1).

The geostrophic flow for a particular cruise may have a more complicated horizontal structure than the one used here (e.g. see Fig. 2 in Marinone and Ripa (1988)). The very simple linear \( x \)-dependence was chosen with the purpose of performing a significant intercruise comparison.

There is no a priori reason for either the classical (24) or the modern (25) method to give a better result. In fact, it was found useful to have different estimates of the flow field—although they are not completely independent—as will be seen in the discussion of the vorticity field (section 6).

Figure 15 shows the seasonal signal of the surface velocity \( V(0, t) \) corresponding to four different calculations. The criteria used in choosing the reference flows and the significance of the fits are presented in the top part of Table 3. One way of gauging the surface flow is by recognizing that it is a measure of the sea surface slope, namely

\[
\eta_x(t) = g^{-1}fV(0, t).
\]  
(26)

Thus the ordinate in Fig. 15 is labelled for \( V \) on the left and for \( \eta_xL \) on the right, with \( L = 70 \text{ km} \), which is roughly the half-width of the basin. One metre of sea surface
elevation (depression) at Guaymas relative to Santa Rosalia, $\eta_2L = \pm 50$ cm, corresponds to an average velocity of $\pm 1.0$ m s$^{-1}$.

<table>
<thead>
<tr>
<th>Criteria</th>
<th>Cruises</th>
<th>Var.</th>
<th>CL</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Type</td>
<td>No.</td>
<td>%</td>
</tr>
<tr>
<td>a: Min. energy</td>
<td>Non-El-Niño</td>
<td>13</td>
<td>39</td>
</tr>
<tr>
<td>b: Min. energy</td>
<td>All</td>
<td>17</td>
<td>46</td>
</tr>
<tr>
<td>c: R.L. = 500 m</td>
<td>All</td>
<td>14</td>
<td>57</td>
</tr>
<tr>
<td>d: R.L. = 700 m</td>
<td>All</td>
<td>11</td>
<td>47</td>
</tr>
<tr>
<td>a: Min. energy</td>
<td>Non-El-Niño</td>
<td>13</td>
<td>68</td>
</tr>
<tr>
<td>b: Min. energy</td>
<td>All</td>
<td>17</td>
<td>13</td>
</tr>
<tr>
<td>c: R.L. = 500 m</td>
<td>All</td>
<td>12</td>
<td>7</td>
</tr>
<tr>
<td>d: R.L. = 700 m</td>
<td>All</td>
<td>8</td>
<td>97</td>
</tr>
</tbody>
</table>

Fits significant at the 90% CL are underlined. Labels a, b, c and d correspond to those of Fig. 15. (R.L. means reference level.)

The four estimates of surface velocity give the same signature pattern: flow towards the head in April through September and towards the mouth in November through February. The classical method with 700 m reference level (curve d) is the only one to show two maxima and minima per year (an important semi-annual component) but this fit is not significant at the 90% CL. Neither is that of curve a, which corresponds to the minimum energy criterion excluding data taken during El Niño events. The two statistically significant estimates, minimum energy with all cruises (curve b) and classical with 500 m reference level (curve c) differ in their amplitudes, as follows.

The maximum surface velocity, reached in August, is $13$ cm s$^{-1}$ for case b and $18$ cm s$^{-1}$ for case c: thus, least energy requires a minimum flow of $-5$ cm s$^{-1}$ at 500 m, and in the same month. On the other hand, the minimum surface velocity, realized about New Year’s Day, is $-18$ cm s$^{-1}$ for case b and $-28$ cm s$^{-1}$ for case c: least energy needs a maximum flow of $10$ cm s$^{-1}$ at 500 m, at this time. We have no clear way for choosing between these two possibilities at this point; when combined with sea level, section 7, case b turns out to be the best.

The vertical structure of the flow is illustrated in Fig. 16(a), drawn with the mean and first ‘empirical’ mode calculated with the 14 cruises that reach 500 m (see Table 1); the velocities at this level were estimated using the ‘minimum energy’ criterion. This results in a flow reversal at 200 m for the fluctuating part (the mean flow is not significantly different from zero). The first mode explains 81% of the vertical/temporal velocity variability, and a seasonal fit to $B_1(t)$ explains 53% of its variance, which is significant at the 88% CL; the semi-annual component is not very important ($\gamma = 0.14$). Not surprisingly, $(V(0)) + A_1(0)B_1(t)$ resembles very well the surface velocity for case b, see Fig. 15 and Table 3.

Very similar results were obtained with the data from the 11 cruises that reach 700 m. For instance, $A_1(z)$ has a similar reversal, at 250 m. The $a posteriori$ estimated value of $B_1$ for cruise 12, which is too shallow to be included in this principal component analysis, falls right on the curve of the seasonal fit of $B_1(t)$. We feel that this extra calculation yields more confidence in the robustness of the results of this section.
The momentum flux at the air–sea interface is given by the wind stress, which was calculated from daily mean wind data collected at the Mexican Weather Service’s Guaymas Observatory, and corresponding to whole years from 1977 to 1980. Figure 17 shows monthly means of
\[
\tau = \rho^{-1} \rho_a C_d V_w (U_w^2 + V_w^2)^{1/2}
\]  
(27)
where \(\rho_a\) is the air density, \(C_d\) is a wind-speed-dependent drag coefficient, and \((U_w, V_w)\) are the transverse and headward wind components. The curve shows the seasonal fit, which explains 74% of the variance, significant at the 95% CL. Of course, we cannot relate \(\tau\) to the vertical shear of \(V\) at the surface—as we did with similar gradients of \(T\) and \(S\) and the heat and salt fluxes—because that would need use of the Ekman flow, not the geostrophic one. However, the visual similarity between Figs. 15 and 17 is too tempting to resist making a regression of the form
\[
\sigma V(0, t) \approx \tau(t)
\]  
(28)
with $\sigma$ an unknown coefficient for which we do not have a physical model. A least-squares fit yields

$$\sigma = (0.9 \pm 0.3) \times 10^{-4} \text{ m s}^{-1}$$

for case b, explaining 52% of the variability, significant at the 95% CL (or the 86% CL if the semi-annual components are neglected; see appendix B). Case c gives similar results: $\sigma = (0.3 \pm 0.2) \times 10^{-4} \text{ m s}^{-1}$, explaining 37% of the variability, significant at the 90% or 80% CL, respectively.

Of course, we do not have a physical model from ‘first principles’ to justify (28). One way to interpret it could be to think of the wind stress $\tau$ distributed over a mixed layer of depth $H$ in balance with a friction force of the form $-\epsilon V$, where $\epsilon = \sigma/H$, and, in turn, $V$ being in geostrophic balance in the normal direction. For $H = 50$ to 100 m, the damping time $\epsilon^{-1}$ equals one or two weeks, which is short enough to force the along-gulf current to appear in phase with the wind stress on annual and semi-annual time scales. We insist that more work needs to be done before (28) is interpreted as causality: it is too easy to discover coherence between two signals which are essentially mono-chromatic and with the same frequency.

Horizontal divergence of the heat and salt fluxes were found to be important, through a deficit in their respective budgets. Since we have extracted a reasonably good estimate of the seasonal velocity signal, it is interesting to ask whether or not it can give advection contributions of the right magnitude, when combined with appropriate along-gulf gradients of temperature or salinity. The excursion of particles driven into Guayas basin by the seasonal velocity signal indicates how far we can go in order to estimate these derivatives. Consider, for the sake of simplicity, a hypothetical velocity signal uniform in the along-gulf direction, say $V(y, t) = \langle V \rangle + V_0 \cos(\omega t + \delta)$. Values for case b are $\langle V \rangle = -1.25 \text{ cm s}^{-1}$ and $V_0 \approx 15 \text{ cm s}^{-1}$ (see Fig. 15). Thus in one year particles experience a net displacement of approximately 400 km and a standard fluctuation, $V_0/\sqrt{2\omega}$, of 540 km. We have data from sections south of Guaymas, made at the same time as nine of the cruises used here, namely 5, 16 and 17 at a distance of 50 km; 2, 3 and 6 at a distance of 123 km; and 7, 10 and 12 at distance of 220 km. Mean values and standard deviations for the 0–50 m averages of the temperature and salinity gradients are presented in Table 4. The Guayas basin is always about 0.06 psu saltier than the waters 100 km to the south; the gradient oscillating between a maximum of twice that value, for September, and zero, for April. Temperature differences can be of either sign and have an amplitude of about half a degree in 100 km; the fluctuation seems to occur in phase with that of salinity. These gradients are combined with 0–50 m average velocities of case
b to calculate the heat and salt advection terms, whose statistics are also presented in Table 4. Comparing these figures with the deficit between the storage and vertical flux divergence, Figs. 7 and 14 respectively, we see that the estimated advectons have the right order of magnitude to close the balance. More data are needed, particularly of non-El-Niño years, before a more detailed comparison can be attempted.

### Table 4. Statistics of the Along-Gulf Gradients of Temperature and Salinity, and Advections of Heat and Salt, for the Top 50 m.

<table>
<thead>
<tr>
<th></th>
<th>$aT/\partial y$ (degC m$^{-1}$)</th>
<th>$\rho c_p V \partial T/\partial y$ (W m$^{-3}$)</th>
<th>$\partial S/\partial y$ (psu m$^{-1}$)</th>
<th>$\rho \beta V \partial S/\partial y$ (kg m$^{-3}$ s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average</td>
<td>$-1.5 \times 10^{-6}$</td>
<td>$-0.8$</td>
<td>$6.0 \times 10^{-7}$</td>
<td>$0.1 \times 10^{-8}$</td>
</tr>
<tr>
<td>Standard deviation</td>
<td>$5.3 \times 10^{-6}$</td>
<td>$1.2$</td>
<td>$5.5 \times 10^{-7}$</td>
<td>$5.8 \times 10^{-8}$</td>
</tr>
</tbody>
</table>

6. **Vorticity**

The evaluation of $\Omega(z, t)$ was explained in the previous section. With the classical criterion (24) it is calculated independently of $V$, whereas with the least-energy one (25) the reference values of both variables are tied. In either case the depth ranges of $V$ and $\Omega$ might be different, for a given cruise, depending on the number of deep casts (see Table 1).

We calculated vorticity with the same four criteria that we used for velocity (see Table 3). The surface vorticity is a measure of the sea-surface curvature:

$$\eta_{ax}(t) = g^{-1}f\Omega(0, t).$$

(30)

The mean value is in all cases negative (anticyclonic) and small, compared with the standard deviation. For instance, for case b it is

$$\langle \Omega(0, t) \rangle \approx -0.02f \quad \text{or} \quad \langle \Omega(0, t) \rangle L \approx -9 \text{ cm s}^{-1}$$

(31)

which corresponds to a depression of sea level at both coasts, relative to the centre of the gulf, of

$$\frac{1}{2}L^2\langle \eta_{ax}(t) \rangle \approx -3 \text{ cm.}$$

(32)

As a comparison, the mean velocity $\langle V(0, t) \rangle$ is somewhere between 0 and $-1 \text{ cm s}^{-1}$, which implies an elevation in Santa Rosalia (Baja California peninsula) and a depression in Guaymas (mainland Mexico) of 0.5 cm. The standard deviation, s.d., of the surface vorticity is given by

$$\text{s.d.}[\Omega(0, t)] = 0.067f \quad \text{or} \quad \text{s.d.}[\Omega(0, t)] L \approx 32 \text{ cm s}^{-1}$$

(33)

with any of the four calculations, representing a standard deviation on both coasts of

$$\text{s.d.}[\frac{1}{2}L^2[\eta_{ax}(t)]] \approx 7.7 \text{ cm.}$$

(34)

With velocity, s.d.$[V(0, t)] = 11 \text{ cm s}^{-1}$, which translates into elevations of opposite signs on the two coasts of s.d.$[L[\eta_{ax}(t)]] = 5.3 \text{ cm.}$ Consequently, the effect of vorticity on coastal sea level is of the same order as, and larger than, that due to the mean flow. Notice that $\Omega$ exceeds $V$ even more, by a factor of two, in the coastal velocity signal
\( (V \pm L\Omega) \) than in the sea-level one \( (V \pm L\Omega - \frac{1}{2}L^2\Omega) \): the flow on either coast might be quite different from the across-gulf mean presented in Fig. 15.

The vertical structure of the vorticity field is illustrated in Fig. 18(a), which depicts the contribution of the mean and first mode, for a principal component analysis of those results of case b that go down to 700 m. As with velocity, Fig. 16(a), a reversal of the variance field is observed at about 200 m.

Of course, we attempted to obtain the seasonal signal of the vorticity field by fits of the type (7). The lower part of Table 3 shows the variance explained with the four realizations of the surface vorticity, as well as the bounds for 90% and 95% CL, determined by the Monte Carlo simulations. Two of the cases, a and d, look surprisingly good and the other two, b and c, look surprisingly bad, neatly illustrating our absolute failure in extracting the seasonal variability of vorticity. The two 'good' estimates are so at the expense of lack of data in the July–September trimester, predicting an unrealistic negative plunge \( (\Omega \approx -0.5f, 0.5L^2\eta_{ax} \approx -55 \text{ cm}) \) in August. The two 'bad' cases have a positive vorticity datum in August, which all but obliterates the possibility of a good fit.

![Figure 18](image)

Figure 18. As Figs. 3, 9 and 16 for the vorticity, i.e. the across-gulf derivative of the velocity. It was not possible to extract a significant seasonal signal for this variable.
7. **Sea Level**

Figure 19 shows the monthly mean of sea level from the Guaymas tide gauge for the period 1952–1984, with some gaps. The error bars depict the standard deviation of the non-seasonal deviations from the monthly mean, which are approximately 7 cm. A seasonal fit, solid curve, explains 98.7% of the variance, which is highly statistically significant (the 99% CL requires a minimum 71% of the variance for just the annual component, or 84% for the annual plus semi-annual). The maximum, 22 cm, is reached in August, and the minimum, −16 cm, in February; the asymmetry of the curve is a manifestation of a slight semi-annual component ($\gamma = 0.024$).

![Sea Level Graph](image)

**Figure 19.** Monthly-mean sea level at Guaymas, circles, and standard deviation of the anomalies, error bars, for a 33-year interval. The solid curve represents the seasonal fit.

Sea-surface changes may in part be due to **barotropic** signals, like those produced by continental shelf waves and the long-period tides, which are unrelated to the thermohaline field. The **baroclinic** contribution, on the other hand, represents variations in the height of the water column due to temperature- and salinity-induced changes of density. Thus

$$\eta_b(t) = \int_{-h}^{0} \{\alpha(z) T(z, t) - \beta(z) S(z, t)\} \, dz$$

(35)

gives a uniform change across the gulf, whereas the slope and curvature of thermohaline origin are given by (26) and (34).

The seasonal fit of sea level at Guaymas is reproduced in Fig. 20 with a thick solid line; the standard deviation from the annual mean is 14 cm. The dashed line represents the seasonal fit of $\eta_b$; by itself, this part can explain 90% of the variability in the observed signal. Adding to $\eta_b$ the contribution due to the across-gulf slope $\eta_s L$ (related to the surface velocity through Eq. (26)) increases the explained variance up to 98%—thin curve in Fig. 20. This result adds confidence in the robustness of the seasonal signal of
velocity estimated in section 5. Notice that the sea-level signal due to the vorticity field, \( 2L^2 \eta_{xx} \), for which no seasonal signal could be extracted, has a very large variance (approximately one third of that of Guaymas' sea level).

8. SUMMARY AND CONCLUSIONS

The Guaymas basin is always gaining heat and losing fresh water through the atmosphere–ocean interface at an annual rate of \( \langle Q \rangle = 113 \text{ W m}^{-2} \) and \( \langle E - P \rangle = 58 \text{ cm y}^{-1} \) respectively. Heat and salt diffuse down; the process is controlled by an eddy coefficient \( K = 3.1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1} \). Thus, the average vertical gradients of temperature and salinity at the surface are (Figs. 3 and 9)

\[
\frac{\partial \langle T \rangle}{\partial z} = \langle Q \rangle / (\rho c_p K) = 0.089 \text{ degC m}^{-1}
\]

and

\[
\frac{\partial \langle S \rangle}{\partial z} = \langle S \rangle (E - P) / K = 0.0021 \text{ psu m}^{-1}.
\]

In other forms, the temperature and salinity contributions to the static stability at the surface are

\[
N_T^2 = g \alpha \frac{\partial \langle T \rangle}{\partial z} = 2.4 \times 10^{-4} \text{ s}^{-2}
\]

and

\[
N_S^2 = -g \beta \frac{\partial \langle S \rangle}{\partial z} = -1.5 \times 10^{-5} \text{ s}^{-2}
\]

which give a net stability of \( N^2 = 2.3 \times 10^{-4} \text{ s}^{-2} \), i.e. a Brunt–Väisälä period of seven minutes. Below the surface, temperature and salinity have, on average, positive second derivatives with depth. Therefore, heat and salt are extracted from Guaymas basin by
horizontal processes. A similar result was found in the northern part of the gulf by Organista (1987) and Bray (1988).

No statistically significant temporal mean of the geostrophic circulation was determined, neither for the velocity (Fig. 16) nor the vorticity (Fig. 18) fields.

The 'normal' seasonal fluctuation is as follows. Heat flux into the ocean, temperature, wind stress, across-gulf mean velocity and sea level have mainly an annual cycle, reaching their minima sometime in December–January–February and their maxima in July–August–September. These extrema are 10 and 230 W m$^{-2}$ for heat flux (Fig. 5); 16 and 32 °C for temperature at the surface; the rest of the water column having a thermal cycle with the same phase but with amplitude decreasing with depth (Figs. 3 and 4); −0.01 and 0.005 N m$^{-2}$ for wind stress (Fig. 17); −18 and 13 cm s$^{-1}$ for surface velocity, water below 200 m flowing in opposition to that above (Fig. 16); and −16 and 22 cm for sea level (Fig. 19). The warming phase is longer—and therefore weaker—than the cooling one (peak heat storage rates for the upper 50 m are 3 and −6 W m$^{-3}$ respectively); vertical diffusion and horizontal advection/diffusion are equally responsible for these changes in temperature (Fig. 7). Thermohaline changes in the density of the upper layers account for about 90% of the sea-level variability at Guaymas; if the geostrophic tilt of the surface associated with the velocity field is added to this effect, then the variance explained increases to 98% (Fig. 20).

The fresh water flux into the ocean and the salinity field behave in a different way from the variables described above, because they have an important semi-annual component. Thus the evaporation-minus-precipitation rate goes up to 3-1 mm d$^{-1}$ in April and down to essentially zero in August, but there is a secondary maximum of 2-2 mm d$^{-1}$ in November and a secondary minimum of 1-3 mm d$^{-1}$ in February (Fig. 11). Salinity variations can also have four extrema per year (Figs. 9 and 10). For instance, the value for the upper 50 m reaches its minimum (34.95 psu) in August, and its maximum (35.48 psu) in December, but there is a secondary minimum of 35.16 psu in March and a secondary maximum of 35.22 psu in May. Salinity changes are due to both vertical diffusion and horizontal advection/diffusion; the latter being probably the more important. Typical extreme values of salt storage in the upper 50 m are −4 and 6 g m$^{-3}$ d$^{-1}$. Salinity variations below about 200 m are in opposition to those above this level.

The estimate of the annual components of salinity is quite stable to changes in the data set. The semi-annual components, however, cannot be calculated if the only cruise in the trimester July–August (which was done during an El Niño event) is excluded. The anomalies from the seasonal signal (particularly those of El Niño cruises) are large, and that is why Bray (1988) concluded—incorrectly according to this work—a lack of salinity seasonality.

In addition to the across-gulf average velocity, whose seasonal variation is mentioned above, there is a vorticity field, which resisted every attempt to extract a seasonal signal from it. With respect to this variable, we are also at variance with Bray (1988) who sees a cyclonic surface circulation in summer and an anticyclonic and weak flow in spring and fall. Vorticity also shows a reversal at about 200 m, suggesting a two-layer pattern for current and salinity variations (instead of the one-layer structure observed for temperature). The contributions of the vorticity to the net surface velocity near each coast have opposite signs and a standard deviation of 32 cm s$^{-1}$, which is three times larger than that due to the across-gulf average flow. Geostrophic balance associates vorticity with the curvature of surface elevation, implying a sea-level signal near each coast with a standard deviation of 7-7 cm. This value is larger than the contribution of the sea-surface slope due to the average velocity, 5-3 cm, and smaller than that due to thermohaline changes in the volume of the water column, 15 cm.
We are confident of the physical reality of the results discussed here—the degree of confidence not being the same for all of them. They are sufficiently robust to endure parameter changes, like the depth of the principal component analysis, or the inclusion or not of particular cruises. It was easier to derive a quantitative seasonal signal for temperature and velocity than for salinity. The latter has relatively more important semi-annual and interannual components; data collected during August of non-El Niño years are needed in order to improve the estimates of the seasonal signal. We hope to be able to use data from more cruises and other sections of the Gulf of California to advance knowledge of its physics.

ACKNOWLEDGEMENTS

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APPENDICES

A. Statistical significance of seasonal fits

In order to calculate confidence levels (CL) for a fit of the form given in Eq. (7) to $M$ observations made at times $(t_j, \quad j = 1, \ldots, M)$ we generated 100 sets of $M$ random numbers in each one, and performed a similar fit to each set using the same dates $(t_j)$. The explained variances were ordered monotonically. We then repeated the process many times and averaged the results until we obtained a stable estimate of the maximum explained variance experienced by 1%, 2%, ..., etc of the random cases.

B. Proportionality of two seasonal signals

The other statistical test used in this work is the following: Consider the regression $a(t) \sim k b(t)$, with $k$ an unknown coefficient, where

$$a(t) = a_0 + a_1 \cos \omega t + \ldots, \quad b(t) = b_0 + b_1 \cos \omega t + \ldots.$$  

Examples are those of Eqs. (10), (19) and (28). Assume that $k$ is determined by minimizing

$$\int \{a(t) - k b(t)\}^2 dt;$$

that the percentage of explained variance is 100$\mu$; and that $k$ has the ‘right’ sign, say, positive. For a certain CL equal to 100$(1 - \lambda)$% we need to determine $\mu_0$ such that $\lambda$ equals the probability that, at random, $\mu > \mu_0$ and $k$ has a given sign. (If the sign of $k$ were irrelevant, which is not the case here, $2\lambda$ must be replaced by $\lambda$ in the formulae below.) Now, both $a(t)$ and $b(t)$ are determined by $N$ coefficients, with $N = 5$ ($N = 3$) if the semi-annual component is (is not) included. If $a_0$ and $a_j/\sqrt{2}$ ($j = 1, \ldots, N - 1$) are considered to be similar independent random variables, with zero mean, then $\lambda$ is the relative area of a circle of radius $\theta_0$ (where $\mu_0 = \cos^2 \theta_0$) on the unit hypersphere in $N$-dimensional space. In other words, if $\theta$ is the ‘angle’ between $a$ and $b$, $\lambda$ is the probability that $\theta < \theta_0$ for uniformly random orientations. For the cases of interest here, we obtain

$$1 - \lambda = \frac{1}{2}(1 + \sqrt{\mu_0}) \quad N = 3,$$

$$1 - \lambda = \frac{1}{4}(3 - \mu_0) \sqrt{\mu_0} \quad N = 5.$$
Thus, for the 90%, 95% and 99% CL, the minimum explained variance equals 37%, 53% and 78%, respectively, if the semi-annual components are included ($N = 5$), or 64%, 81% and 96%, respectively, if they are not ($N = 3$).

$C$. Parameters

\[
\begin{align*}
    f &= (6.78 \pm 0.07) \times 10^{-5} \text{s}^{-1} \\
    g &= 9.8 \text{ m s}^{-2} \\
    \alpha &= 2.8 \times 10^{-4} \text{(degC)}^{-1} \quad \text{at the surface} \\
    &= 1.4 \times 10^{-4} \text{(degC)}^{-1} \quad \text{at 650 m} \\
    \beta &= 7.4 \times 10^{-4} \text{psu}^{-1} \quad \text{at the surface} \\
    &= 7.6 \times 10^{-4} \text{psu}^{-1} \quad \text{at 650 m} \\
    c_p &= 3990 \text{J(kg degC)}^{-1} \quad \text{at the surface} \\
    &= 3980 \text{J(kg degC)}^{-1} \quad \text{at 650 m} \\
    \rho &= 1024 \text{ kg m}^{-3} \quad \text{at the surface} \\
    &= 1030 \text{ kg m}^{-3} \quad \text{at 650 m}
\end{align*}
\]

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