Poleward flows in the southern California Current System: Glider observations and numerical simulation

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Three years of continuous Spray glider observations in the southern California Current System (CCS) are combined with a numerical simulation to describe the mean and variability of poleward flows in the southern CCS. Gliders provide upper ocean observations with good across-shore and temporal resolution along two across-shore survey lines while the numerical simulation provides a dynamically consistent estimate of the ocean state. Persistent poleward flows are observed in three areas: within 100 km of the coast at Point Conception, within the Southern California Bight (SCB), and offshore of the SCB and the Santa Rosa Ridge (SRR). Poleward transport by the flows within the SCB and offshore of the SRR exceeds the poleward transport off Point Conception, suggesting that the poleward flows are not continuous over the 225 km between observation lines. The numerical simulation shows offshore transport between the survey lines that is consistent with some of the poleward flow turning offshore before reaching Point Conception. The poleward current offshore of the SRR is unique in that it is strongest at depths greater than 350 m and it is observed to migrate westward away from the coast. This westward propagation is tied to westward propagating density anomalies originating in the SCB during the spring-summer upwelling season when wind stress curl is most strongly positive. The across-shore wave number, frequency, and phase speed of the westward propagation and the lack of across-shore transport of salinity along isopycnals are consistent with first-mode baroclinic Rossby dynamics.


1. Introduction

[2] The California Current System (CCS), an eastern boundary current system, is composed of variable equatorward and poleward currents. The equatorward flowing California Current carries cold, fresh waters of northerly origin while poleward flows transport warm, salty waters of southerly origin [Wooster and Jones, 1970; Hickey, 1979; Lynn and Simpson, 1987; Huyer et al., 1989]. The poleward flowing California Undercurrent (CU) within the Southern California Bight (SCB) [Lynn and Simpson, 1990] and near Point Conception, as well as surface poleward flow near the coast, have been observed for decades by the California Cooperative Oceanic Fisheries Investigations (CalCOFI) program (http://www.calcofi.org) and other investigations [Lynn and Simpson, 1987; Bray et al., 1999]. Recent velocity observations [Davis et al., 2008; Gay and Chereskin, 2009] have revealed an additional subsurface poleward current offshore of the SCB to be a persistent feature in the CCS. This paper uses new observations collected by underwater gliders in the southern portion of the CCS and a new, regional, numerical state estimate to characterize the mean and variability of the poleward flows. Our observations show that the poleward current offshore of the SCB propagates westward in response to density anomalies propagating westward from the SCB and that the across-shore wave number and frequency of this westward propagation are consistent with first-mode baroclinic Rossby wave dynamics.

[3] Historically, flow within the southern CCS has been diagnosed from thermal wind and decades of repeat hydrography, largely through the CalCOFI program. In most cases, the 500 db surface has been used as a level of no motion since CalCOFI measurements extend to 500 m. The resulting picture of the geostrophic flow [Sverdrup and Fleming, 1941; Hickey, 1979; Lynn and Simpson, 1987] shows the equatorward flowing California Current near the surface and somewhat offshore, the poleward flowing CU near the coast with highest velocity at depths of 100–300 m, and seasonally reversing surface flow near the coast. South of Point Conception, the CU appears inshore of the Santa Rosa Ridge (SRR, Figure 1) and flows between the various islands. Gaps in the SRR provide pathways for the CU to exit the SCB [Lynn and Simpson, 1990].
[4] Subsurface poleward flow offshore of the SRR has been inferred occasionally from hydrography. Sverdrup and Fleming [1941] found northward flow at a depth of 200 m near the offshore side of the SRR during three cruises from March to July 1937. They identified the poleward flowing waters as having higher temperature and salinity than waters within the California Current, an indication of southerly origin. Lynn and Simpson [1990] used thermal wind referenced to 1000 db from a single survey in July 1985 to infer poleward flowing water of southern origin in the same region; they attributed the flow to an eddy formed by CU waters discharged through a gap in the SRR.

[5] Velocity measurements within the southern CCS have recently revealed a poleward current offshore of the SRR to be a persistent feature. Geostrophic velocities referenced to vertically averaged currents from glider measurements on CalCOFI Lines 80.0 and 93.3 [Davis et al., 2008] showed mean subsurface poleward flow within 100 km of the coast (the CU) as well as 200–250 km offshore on Line 80.0 and Line 93.3 from 2005 to early 2007. Gay and Chereskin [2009] used 10 years of shipboard acoustic Doppler current profiler (ADCP) measurements from the quarterly CalCOFI cruises to show that the offshore poleward current is a persistent feature along Lines 86.7, 90.0, and 93.3, which fall within the SCB, but found only a single core (the CU) in the 10 year mean near Point Conception. The second poleward core at Line 80.0 seen in the glider observations was likely an artifact of the shortness of the data record available [Davis et al., 2008], and glider observations of longer duration presented here do not show significant mean poleward flow at that location. The current offshore of the SRR has significant poleward flow at depths of 500 m and its speed diminishes with decreasing depth. Consequently, geostrophic calculations that assume zero flow at 500 m produce a surface intensified equatorward flow at the same location [Davis et al., 2008].

[6] The variability of the CU and shallow poleward flow within the SCB have been well documented [Chelton, 1984; Lynn and Simpson, 1987; Bray et al., 1999; Gay and Chereskin, 2009], so our analysis focuses on the variability of the poleward current offshore of the SRR. Gay and Chereskin [2009] quantified the seasonal variability in transport of this offshore current and found it to be strongest in the fall. They found relatively little variability in the position of the current, but this is likely an effect of their mapping procedure which used long decorrelation scales and averaging in the alongshore direction [Gay and Chereskin, 2009]. The equatorward flowing California Current in the same region is known to meander [Lynn and Simpson, 1987], and our observations show that the poleward current offshore of the SRR propagates westward in a manner that is largely consistent with Rossby wave dynamics.

[7] Rossby waves have been previously observed and modeled within the CCS. White et al. [1990] used satellite observations of sea surface height to characterize annual Rossby waves generated along the California coast. Strub and James [2000] hypothesized that westward propagating Rossby waves control offshore movement of a seasonal equatorward jet off central and northern California. Lynn and Bograd [2002] found that El Niño related dynamic height anomalies along Line 90.0 propagated westward at a phase speed consistent with a westward propagating Rossby wave. In a quasi-geostrophic numerical model, Auda et al. [1991] found that wind forcing generated first-mode baroclinic Rossby waves between 25° and 33°N. Di Lorenzo [2003] used the Regional Ocean Modeling System (ROMS) to demonstrate that density anomalies generated by alongshore wind stress and wind stress curl propagated westward from the SCB only when the \( \beta \) effect was included. This dependence on the \( \beta \) effect indicated that the westward propagation was the result of Rossby wave dynamics.

[8] The remainder of this paper is arranged as follows: section 2.1 describes glider observations in the southern CCS; section 2.2 describes the numerical simulation; section 3.1 discusses the observed mean, transport, and variability of alongshore flow in the southern CCS; section 3.2 discusses the mean structure of poleward jets; section 3.3 characterizes the westward propagation of the current offshore of the SRR and compares the observations with Rossby wave dynamics; and section 4 summarizes the results. Two appendices discuss the accuracy of glider-based vertically averaged current measurements and describe velocity estimation using glider-mounted acoustic Doppler profilers.

2. Methods

2.1. Glider Observations

[9] Spray gliders [Sherman et al., 2001; Davis et al., 2002; Rudnick et al., 2004] are buoyancy-driven autonomous underwater vehicles that profile from the surface to a programmed depth while moving along a sawtooth flight path. Glider surveys in the southern portion of the CCS began in 2005 along established CalCOFI survey lines. Since October 2006, CalCOFI Lines 80.0 and 90.0 have been surveyed nearly continually by Spray gliders (Figure 2). Line 80.0 extends...
from Point Conception southwestward for approximately 350 km and passes through the upwelling center off Point Conception. Line 90.0 extends roughly 525 km southwestward from Dana Point, California and passes through the SCB (Figure 1). We use data collected between October 2006 and November 2009 in this analysis.

A typical deployment in the southern CCS begins just offshore of the kelp zone where the glider is placed in the water from a small boat. Gliders surveying Line 80.0 are deployed from Santa Barbara, California, and Line 90.0 gliders are deployed between San Diego and Dana Point, California. The gliders navigate to the inshore end of the appropriate CalCOFI survey line, then begin following the line to near its offshore end (typically to CalCOFI stations 80.100 and 90.100) where they reverse course and return to the inshore end of the line. Gliders take about three weeks to complete a transect in one direction, and battery life is sufficient for a glider to complete at least four full transects before being recovered and replaced by a refueled glider. The gliders are programmed to dive to 500 m when in deep water so that they cover approximately 3 km horizontally during a dive that lasts about 3 h. This analysis uses 57 transects from 13 deployments along Line 80.0 and 42 transects from 11 deployments along Line 90.0. Strong eddy activity at Line 80.0 often forces the gliders significantly off of their intended paths (Figure 1); gliders on Line 90.0 are only rarely displaced more than a few kilometers from the established survey line. Gliders deployed in the southern CCS are each equipped with a Sea-Bird 41CP conductivity-temperature-depth (CTD) sensor, a custom Sontek Argonaut 750 kHz acoustic Doppler profiler (ADP), and a Seapoint chlorophyll a fluorometer, all of which collect measurements on the ascending portion of the dive only [Davis et al., 2008].

GPS fixes at the beginning and end of each dive are combined with records of pitch, heading, and depth during the entirety of each dive to measure vertically averaged currents, and these measurements are used to reference both geostrophic [Davis et al., 2008; Todd et al., 2009] and Doppler-derived velocity profiles (e.g., Figure 3). Vertically averaged currents from gliders have been shown to agree well with independent measurements from moored ADCPs [Davis et al., 2002; Todd et al., 2009], and Appendix A considers sources of error in vertically averaged current measurements.

We use glider observations directly in our analysis whenever possible, but some applications (e.g., calculations of mean sections) require data on a uniform grid. For these applications, we objectively map the observations at each depth to a uniform grid using a Gaussian covariance matrix with a 30 km decorrelation scale and noise-to-signal ratio of 0.05. The length scale is close to the integral length scale of observed vertically averaged currents.

We also use objective mapping to calculate geostrophic velocities. We estimate the across-shore (along track) gradient of the density field by specifying appropriate covariance matrices. We again specify a 30 km length scale in the mapping, which eliminates density gradients due to internal waves. Alongshore (across-track) geostrophic

Figure 2. Across-shore and temporal sampling pattern along CalCOFI Lines (a) 80.0 and (b) 90.0. Each point denotes the location of a glider profile. Black points indicate profiles with valid ADP-derived current estimates; profiles without ADP-derived current estimates are in grey. The dashed black line in Figure 2b shows the location of the Santa Rosa Ridge.

Figure 3. Alongshore velocities from an inshore-to-offshore transect along Line 90.0 from 18 June 2008 to 10 July 2008. (a) Vertically averaged alongshore velocities for each dive used to reference ADP velocities (blue) and objectively mapped velocities used to reference geostrophic velocities (red). (b) ADP-derived velocities with no temporal smoothing. (c) The same ADP-derived velocities after filtering with a 30 h Gaussian in the time domain. (d) Geostrophic velocities. Tick marks on the uppermost horizontal axis indicate the locations of individual profiles. Positive (negative) velocities are poleward (equatorward). Dark grey shading (Figures 3b–3d) represents the bathymetry along Line 90.0.
velocities are calculated from the estimated density gradient by integrating the thermal wind relation and referencing to mapped vertically averaged current observations (e.g., Figure 3a).

The glider-mounted ADP provides vertical profiles of horizontal shear much like a lowered ADCP (LADCP) attached to a CTD rosette and lowered from a ship. Our processing is similar to that of the LADCP processing described by Visbeck [2002] but with the glider-based measurement of vertically averaged velocity used to reference the velocity profiles. Appendix B provides the details of estimating horizontal velocity fields from the ADP. Sections of ADP-derived velocity (Figure 3b shows an example transect) have significant high-frequency variability from tidal, inertial, and other effects. Since this analysis focuses on lower-frequency processes, we apply a 30 h Gaussian filter in the time domain [Pope, 2000, Table 13.2] to the ADP velocity profiles when using individual transects of alongshore velocity in sections 3.2 and 3.3. These filtered velocities (Figure 3c) agree well with alongshore geostrophic velocities (Figure 3d). Since both the ADP-derived velocities and the geostrophic velocities are referenced to the measured vertically averaged currents, it is the agreement in vertical structure (shear) between filtered ADP-derived currents and geostrophic currents that is most reassuring. Davis [2011] calculates velocities from Spray glider-mounted ADPs by integrating vertical shears and finds similar agreement between smoothed ADP-derived velocities and geostrophic velocities. The slow horizontal speed of the gliders (about 0.25 m s$^{-1}$) results in a 30 km filter being nearly equivalent to a 30 h filter, so velocity estimates are not filtered in the time domain before being mapped onto a uniform grid. In this analysis, we use ADP-derived velocity estimates whenever possible, and only substitute geostrophic velocities during periods when ADP-derived current estimates are unavailable due to instrument failures (Figure 2).

The glider-based velocity estimates used in this analysis are referenced to measured vertically averaged currents and do not assume a level of no motion. This distinction sets them apart from the relative geostrophic velocity estimates used for most previous work in the CCS. Our results, with those of Davis et al. [2008] and Gay and Chereskin [2009], emphasize the need for properly referenced velocity measurements in the CCS.

### 2.2. Numerical Simulation

The numerical simulation used here is a regional version of the MITgcm [Marshall et al., 1997a, 1997b] that is based on the incompressible Navier Stokes equations. The model domain is [130°W 114°W] × [27.2°N 40°N] (Figure 1) with $\frac{1}{16}$ (about 6 km) horizontal resolution and 72 vertical levels. The model domain includes the west coast of North America from near Punta Eugenia in Baja California Sur, Mexico to near Cape Mendocino in California. The model is run for the period from 1 January 2007 to 30 July 2009 with a time step of 1200 s. Table 1 gives the values of viscosity, bottom drag, and diffusivities used in the model. First guess initial conditions and lateral boundary conditions are from an MIT-ECCO product [Forget, 2010]. First guess of the atmospheric state is from the North American Mesoscale (NAM) model produced by the National Centers for Environmental Prediction (NCEP).

We employ the 4DVAR state estimation technique [Wunsch and Heimbach, 2007] to bring the model solution into consistency with observations. The method uses the adjoint of the MITgcm [Heimbach et al., 2005] to solve for the model controls (i.e., initial conditions and atmospheric state) that minimize the weighted misfit between the simulation and observations. The observations to be fit include: satellite sea surface temperature, satellite sea surface height (SSH), Argo CTD profiles, CalCOFI CTD profiles, Spray glider CTD profiles, moored CTD data, and other ship-based CTD measurements from the region. Glider profiles from Lines 80.0 and 90.0 (section 2.1) are supplemented by additional glider profiles from CalCOFI Line 66.7 off Monterey Bay. Velocity measurements are not used in the optimization. From a reference run with first guess controls, the model controls are iteratively adjusted to fit the observations in a least squares sense. The controls are constrained to remain within estimated errors of the first guesses, and smoothness constraints are also enforced. Our approach to optimizing the simulation has been to progressively increase the amount of structure in the controls, thereby ensuring large-scale agreement with observations before attempting to match smaller-scale features. To this end, the adjoint model is run with elevated horizontal viscosity (Table 1). For more on the adjoint method, see Wunsch [2006] and Wunsch and Heimbach [2007].

The simulation analyzed is a 31 month forward run of the model. The solution is strictly governed by the modeled dynamics and the optimized initial conditions and forcing fields. At this point in the optimization (the 64th data-assimilating iteration), only surface forcing and initial conditions have been solved for; boundary conditions are directly from the $1^\circ \times 1^\circ$ MIT-ECCO product. In contrast to sequential data assimilation techniques where unphysical discontinuities in the model state are a byproduct of the optimization, our simulation is dynamically consistent (adhering rigidly to model physics) while also being qualitatively consistent with the observed ocean. Though not a converged state estimate, the present solution exhibits the structure and temporal variability of the observed poleward flows (section 3.1), making it useful for analysis. The most significant remaining issue in the simulation is that, though the current structure is correct, the speeds, and therefore transports, tend to be 2–3 times smaller than observed. We suspect that this discrepancy is due to a combination of the smoothness constraints on the surface forcing fields as well as elevated viscosity in the adjoint run and our use of unadjusted, low-resolution ($1^\circ \times 1^\circ$) lateral boundary conditions. Our choice in developing the simulation has been to err on the side of smoothness and obtaining correct structure, and the slower
from more than a decade of shipboard observations and (a and b) Mean alongshore currents from all survey lines, since this component of the flow is largest in the remainder of our analysis focuses on only the alongshore component, defined to be perpendicular to the bathymetry along the survey lines. The mean in Figures 4a and 4b is over the period October 2006 to November 2009 while the mean in Figures 4c and 4d is over the period January 2007 to July 2009.

3. Results and Discussion
3.1. Alongshore Currents and Transport
[19] Mean vertically averaged currents from glider observations along Lines 80.0 and 90.0 (Figure 1) are largely orthogonal to the two survey lines. In areas where the mean vertically averaged currents are equatorward, there tends to be an onshore component to the flow. Where poleward flows are strongest (e.g., near Point Conception, within the SCB on Line 90.0, and near 120°W on Line 90.0), depth averaged flow has little across-shore component. The remainder of our analysis focuses on only the alongshore component of the flow, defined to be perpendicular to the survey lines, since this component of the flow is largest in the mean and can be analyzed using geostrophic velocity estimates during times when ADP-derived velocity estimates are unavailable (Figure 2).

3.1.1. Mean Alongshore Flow
[20] The mean alongshore flow calculated over all available glider data on Lines 80.0 and 90.0 (Figures 4a and 4b) shows equatorward and poleward flow along both survey lines. Equatorward flow is strongest in the upper 150 m of the water column along both survey lines and is generally found greater than 50 km offshore at Line 80.0 and greater than 175 km offshore at Line 90.0. This equatorward flow is consistent with the well known California Current [Hickey, 1979; Lynn and Simpson, 1987]. Poleward flow at Line 80.0 is concentrated within 100 km of shore in the CU with peak speeds near 0.08 m s\(^{-1}\) at depths between 50 and 300 m. At Line 90.0, mean flow inshore of the SRR (the dashed line 175 km from shore in Figures 4b and 4d) is poleward at all depths with some surface intensification and peak speeds of 0.05–0.07 m s\(^{-1}\). This inshore poleward flow has been referred to alternately as the CU and the Inshore Countercurrent, with the latter referring to flow that manifests at the surface [Lynn and Simpson, 1987], but the mean flow presented here shows no distinction between subsurface and surface poleward flow that would warrant two names. Offshore of the SCB, mean poleward flow is found from the SRR to the offshore end of the survey line at depths below 300 m. The mean alongshore flow derived from the glider observations agrees well with the mean of Gay and Chereskin [2009] from more than a decade of shipboard observations, which suggests that the glider surveys now provide sufficient data to calculate stable mean fields.

3.1.2. Net Volume Transport
[21] The numerical simulation reproduces the key features of the mean alongshore flow at Lines 80.0 and 90.0 (Figures 4c and 4d). Off Point Conception, the model generates a narrow CU within 100 km of the coast with mean poleward velocities exceeding 0.02 m s\(^{-1}\) between 100 and 400 m depth. Poleward velocities in the CU core are low by a factor of 2 to 3 in the model, but poleward flow extends beyond 500 m as in the observations. Along Line 90.0, the model produces poleward flow throughout the water column inshore of the SRR with highest velocities in the upper 150 m in agreement with observations. Offshore of the SRR, the model produces poleward flow in the region of the observed poleward current, but again with speeds reduced by a factor of 2 to 3. Equatorward flow is again surface intensified and split into at least two branches as in the observations. The agreement between the observed and modeled velocity fields is particularly encouraging since the numerical simulation does not incorporate any direct velocity observations. To produce realistic currents, the model is adequately reproducing the observed density and sea surface height fields that drive currents.

3. Results and Discussion
poleward transport through Line 90.0 is 1.0 Sv. Simulated mean transports are of the correct sign along both survey lines, but transports through Line 90.0 have smaller magnitude than observed due to a combination of the weaker mean currents in the numerical simulation (Figure 4) and the difference in temporal coverage between the observations and simulation (Table 2 gives observed transports during the time span of the simulation). Transport through the section between CalCOFI stations 80.100 and 90.100, which connects the offshore endpoints of the two glider survey lines, is 1.9 Sv directed offshore in the model, and balances the horizontal transport into the region through Lines 80.0 and 90.0. Net vertical transport at 500 m is not a significant contribution to the net volume transport.

[24] These estimates of alongshore transport through the survey lines demonstrate the importance of using properly referenced velocity measurements in the CCS. Previous work by Roemmich [1989] and Bograd et al. [2001] using geostrophic velocities relative to a level of no motion at 500 m in a box bounded by CalCOFI Lines 76.7 and 93.3 (about 75 km north and south, respectively, of our survey lines) found net equatorward transport through both the northern and southern boundaries. A similar calculation of geostrophic transports relative to 500 m using the glider observations also gives net equatorward transport through Line 90.0. The assumption of negligible flow at 500 m results in estimates of upper ocean transport that are qualitatively different from transport estimates obtained using properly referenced velocity estimates.

### 3.1.3. Variability in Poleward Flows

[25] The mean alongshore flow discussed in section 3.1.1 averages the meandering of the California Current and the poleward currents with the result being broader, slower flow than the synoptic current [Chereskin and Trunnell, 1996; Bray et al., 1999]. Time series of alongshore currents averaged over the upper 500 m (or full depth in shallower water) from each transect along Lines 80.0 and 90.0 (Figure 5) show that the strength and position of poleward and equatorward currents vary significantly over the available record. At Line 80.0 (Figure 5), the CU usually manifests as poleward (positive) alongshore flow within 100 km of the coast with peak vertically averaged velocities exceeding 0.20 m s$^{-1}$. At Line 90.0 (Figure 5b), poleward flow dominates the vertical averages within the SCB with peak vertically averaged velocities greater than 0.15 m s$^{-1}$. Offshore of the SRR on Line 90.0, the position of the strongest poleward flow is more variable and peak speeds regularly exceed 0.15 m s$^{-1}$. Poleward flow offshore of the SRR tends to migrate westward repeatedly along the survey line; we focus on this westward propagation in section 3.3.

[26] The numerical simulation also shows considerable temporal variability in the poleward flows (Figure 6). The color scale is smaller in Figure 6 than in the previous plot to show the structure of the weaker flow in the model. Westward propagating signals are more apparent in the simulated alongshore velocity fields at Line 80.0, but the CU remains near the coast most of the time while other features migrate offshore. At Line 90.0, poleward flow dominates within the SCB as in the observations, while the poleward current offshore of the SRR migrates westward regularly.

### 3.2. Mean Poleward Jets

[27] The poleward currents off Point Conception, within the SCB, and offshore of the SRR are narrow, meandering features with peak speeds substantially larger than indicated by the mean velocity sections of Figure 4. The narrowness and relative swiftness of these features justifies referring to

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**Figure 5.** Hovmöller plots of vertically averaged alongshore current in the upper 500 m at (a) Line 80.0 and (b) Line 90.0. Observations were objectively mapped using a Gaussian covariance with decorrelation scales of 30 km and 60 days in the across-shore direction and time, respectively. The noise-to-signal ratio was set to 0.1. Regions with mean square errors exceeding 20% are masked. Contours are drawn every 0.02 m s$^{-1}$ with the zero contour given in bold. Positive velocities are poleward. The dashed line at 175 km along Line 90.0 (Figure 5b) denotes the location of the Santa Rosa Ridge. The bold green line in Figure 5b indicates a westward propagation speed of 0.90 km d$^{-1}$ as calculated in section 3.3.
them as ‘jets’. To characterize the currents and the associated transports, we identify the poleward jets in each transect along Lines 80.0 and 90.0 and calculate mean properties of each current in a jet-following coordinate system. This allows us to define the width and depth of each jet and to show that the current offshore of the SRR is notably deeper than the others and likely not continuous with the CU at Pt. Conception.

[26] We identify subsurface poleward jets as local maxima in the alongshore velocity averaged between 150 and 500 m (\(v\)). For each maximum, we define the edges of the jet to be the first point in the onshore and offshore directions for which \(v\) equals half the peak value. A particular local maximum is excluded if it is exceeded by another local maximum within its identified edges, the value of \(v\) at the peak is less than 0.04 m s\(^{-1}\), or the edges are less than 5 km apart. Figure 7 shows a typical transect on Line 90.0 and the identified jets. Along Line 80.0, the largest jet with a peak within 100 km is included in averaging; along Line 90.0, the largest jet inshore of the SRR (less than 175 km from shore) and the largest jet between the SRR and 400 km from shore are included in separate averages. We create a jet-following horizontal coordinate, \(\tilde{x}\), by setting \(\tilde{x} = 0\) (\(\tilde{x} = 1\)) at the inshore (offshore) edge of each jet. Observed properties of the jets are interpolated onto a uniform grid in \(\tilde{x}\). We then average jet observations between transects in the jet-following coordinates.

[29] The position and width of each jet determines the jet-following coordinate system, \(\tilde{x}\), so it is appropriate to discuss those properties of the jets before considering the results of averaging jets together. Table 3 gives the mean position and width of each of the three poleward jets. The jet off Point Conception is located closest to the coast. The mean position of the jet within the SCB is near the center of the bight, reflecting its tendency to meander within the bight. The jet offshore of the SRR on Line 90.0 is found about 100 km offshore of the SRR on average. We find the widths of the jets to be somewhat smaller than inferred from 10 years of shipboard ADCP data from CalCOFI [Gay and Chereskin, 2009]. The discrepancy may be the result of Gay and Chereskin [2009] using long decorrelation scales and averaging multiple transects in processing shipboard ADCP data. In both analyses, the jet offshore of the SRR is wider than the nearshore jets at Point Conception and within the SCB.

[30] As expected, averaging the observations in a jet-following coordinate system results in peak velocities of the resulting mean jets (Figures 8a–8c) that are substantially larger than when the averaging does not account for meandering jets. At Line 80.0, poleward velocities exceed 0.11 m s\(^{-1}\) at depths from 100 to 300 m near the center of the CU. This mean speed agrees well with the mean speed for the CU off northern California inferred from Lagrangian

**Table 3.** Mean Position, Width, and Volume Transport for the Poleward Jets Off Pt. Conception on Line 80.0. Within the SCB on Line 90.0, and Offshore of the SRR on Line 90.0.

<table>
<thead>
<tr>
<th>Region</th>
<th>Point Conception</th>
<th>Offshore SCB</th>
<th>Within SCB</th>
</tr>
</thead>
<tbody>
<tr>
<td>Position (km)</td>
<td>43 ± 4</td>
<td>282 ± 12</td>
<td>76 ± 6</td>
</tr>
<tr>
<td>Width (km)</td>
<td>36 ± 3</td>
<td>55 ± 8</td>
<td>31 ± 3</td>
</tr>
<tr>
<td>Volume transport (Sv)</td>
<td>2.0 ± 0.2</td>
<td>2.9 ± 0.5</td>
<td>1.4 ± 0.3</td>
</tr>
</tbody>
</table>

*Values are reported as the mean plus or minus standard error for each quantity with units noted. Estimates of standard errors are based on the integral timescale for each quantity.
drifters [Garfield et al., 1999] and from shipboard ADCP surveys over the shelf break from southern California to Vancouver Island [Pierce et al., 2000]. Poleward velocities exceed 0.05 m s$^{-1}$ throughout the upper 500 m in the core of the jet. Within the SCB on Line 90.0, poleward flow peaks around 0.08 m s$^{-1}$ near the surface and exceeds 0.05 m s$^{-1}$ between the surface and 420 m depth. Though the jet was identified based on its subsurface properties, the mean jet (Figure 8c) extends to the surface; there appears to be no distinction between surface and subsurface poleward flow in the SCB. The poleward jet offshore of the SRR at Line 90.0 is concentrated deeper than the other jets. Poleward flow exceeding 0.08 m s$^{-1}$ is located deeper than 350 m and extends below the 500 m profiling depth of the gliders. Above this core, vertical shear is weak with isolines of poleward velocity nearly vertical. The depth of the velocity core offshore of the SRR is notably deeper than reported by Gay and Chereskin [2009] whose data were limited to the upper 400 m because of the range of the shipboard ADCP.

[31] Salinity and density sections through the mean jets (Figures 8d–8f) confirm that the poleward flows transport warm and salty water northward below the thermocline. For each of the three cores of poleward flow, salinity increases along isopycnals in the onshore direction. Although isopycnals slope downward toward the coast, isohalines slope upward toward the coast within the core of the jets. This increase in spiciness [Munk, 1981] toward the coast indicates a change in water masses across the jets with warm and salty waters of southern origin [Lynn and Simpson, 1987] on the inshore side of the poleward jets. At the depth of the core of the jet offshore of the SRR, density and salinity are notably greater than in the jets within the SCB and off Point Conception.

[32] Mean poleward transport by the two jets on Line 90.0 exceeds the poleward transport by the CU at Point Conception by more than a factor of two (Table 3). The CU at Point Conception has mean poleward transport of 2.0 Sv, an estimate that compares well with direct observations of the CU off northern Baja California near 31°N by Wooster and Jones [1970] and with the estimate of Gay and Chereskin [2009] from shipboard ADCP measurements on CalCOFI Lines 76.7, 80.0, and 83.3. Poleward transport by the jets along Line 90.0 is notably greater than reported by Gay and Chereskin [2009], and may be attributable to the deeper profiling depth of the gliders compared to the shipboard ADCP. The mean offshore transport between CalCOFI stations 80.100 and 90.100 of 1.9 Sv in the model is consistent with much of the poleward flow by the jets on Line 90.0 turning offshore before reaching Point Conception. The deeper depth and greater density and salinity of the jet offshore of the SRR and the evidence for offshore turning of the flow suggest that the poleward current offshore of the SRR is generally not continuous with the CU at Point Conception.

3.3. Westward Propagation

[33] The variability in vertically averaged alongshore currents (Figure 5) suggests that the poleward undercurrent offshore of the SRR on Line 90.0 propagates westward while the CU at Point Conception and the poleward flow within the SCB meander back and forth in more limited regions. In this context, “westward” means toward the offshore end of Line 90.0, about 26° south of due west, since our observations are only along the line. We use autocorrelations of observed variables to verify this result and infer the dynamics behind the westward propagation of the poleward current offshore of the SRR.

3.3.1. Autocorrelations

[34] We calculate autocorrelations by averaging products of detrended pairs of measurements with similar spatial and temporal separations into bins that are 10 km by 30 days, a size that ensures a sufficient number of data in each bin, then normalizing by the variance. Observations are detrended by removing a mean and across-shore trend using least squares; no spatial or temporal harmonics are removed. Positive across-shore separations are to the west since we measure distance from the coast.

[35] The autocorrelation of alongshore velocity at 300 m depth and offshore of the SRR (greater than 200 km offshore) on Line 90.0 shows a clear signal of westward propagation (Figure 9a). We show only the autocorrelation of alongshore velocity at 300 m since autocorrelations are similar throughout the upper 500 m and the core of the poleward flow is found near 300 m. Pairs of observations in which the later observation is further offshore (to the west) tend to be well correlated over greater separations than pairs for which the later observation is inshore. The banding pattern of the autocorrelation resembles the autocorrelation of a westward propagating sinusoid, which would itself be a two-dimensional sinusoid with a maximum at zero separation and the same wave number and frequency as the original sinusoid. This similarity suggests that we can extract the across-shore wave number, frequency, and across-shore phase speed of the westward propagating signal directly from the autocorrelation. We fit a sinusoid to the autocorrelation using least squares and weighted by the squared number of pairs of observations in each bin.
with best fit to the observed autocovariance has an across-shore wave number of $-4.1 \times 10^{-3}$ km$^{-1}$ (244 km wavelength) and frequency of $3.7 \times 10^{-3}$ d$^{-1}$ (273 day period). A negative across-shore wave number corresponds to offshore propagation. The across-shore phase speed is then 0.90 km d$^{-1}$. A line representing westward propagation at this speed is plotted on the Hovmöller plot of vertically averaged alongshore velocity (Figure 5b) for comparison. Inshore of the SRR on Line 90.0 and at Line 80.0 off Point Conception, autocorrelations of observed alongshore velocity and other variables (not shown) show little evidence of westward propagation, so we limit further investigation of westward propagation to the region offshore of the SRR along Line 90.0.

The autocorrelation of simulated alongshore velocity between 200 and 550 km from shore on Line 90.0 (Figure 9b) is similar to the autocorrelation of observed velocities (Figure 9a). The westward propagating signal is again clearly evident but with longer correlation scales in the direction of propagation. The best-fit sinusoid has an across-shore wave number of $-3.0 \times 10^{-3}$ km$^{-1}$ (334 km wavelength) and a frequency of $3.1 \times 10^{-3}$ d$^{-1}$ (326 day period) with associated across-shore phase speed of 1.03 km d$^{-1}$; the simulation is producing westward propagation with slightly greater wavelength and period than observed. Although details of the simulated velocity field differ from the observed velocities, the agreement between the autocorrelations of observed and simulated velocities indicates that the model is generating the same type of variability as observed.

At the low frequencies considered here, subsurface flow in the CCS is primarily geostrophic, so, given a relatively constant barotropic flow (e.g., a reasonably steady-in-time sea level gradient), we expect the variability in observed poleward currents to be coupled to the variability in the density gradients via the thermal wind relation. For various isopycnals below the thermocline, the autocorrelations of across-shore isopycnal slope (not shown) have similar structure to the autocorrelation of alongshore velocity (Figure 9a). The across-shore wave number and frequency of the best-fit sinusoid to the autocorrelation of across-shore slope of the 26.5 kg m$^{-3}$ isopycnal are $-3.6 \times 10^{-3}$ km$^{-1}$ (279 km wavelength) and $3.7 \times 10^{-3}$ d$^{-1}$ (270 day period). This isopycnal is located at a depth of about 225 m in the region of the poleward current (Figure 8e).

The sinusoidal autocorrelations of alongshore velocity and isopycnal slope suggest the presence of a wave-like phenomenon. Analysis of properties along isopycnals provides further support for this hypothesis. The autocorrelation of depth of the 26.5 kg m$^{-3}$ isopycnal (Figure 10a) yet again shows evidence for westward propagation, but the autocorrelation of salinity on that isopycnal (i.e., spice) shows no evidence of westward propagation (Figure 10b). While isopycnal depth anomalies propagate westward, there is no associated across-shore transport of salinity along the isopycnal. If the observed westward propagation was caused by an advective process, we would expect salinity anomalies to be transported westward; since this does not occur, our observations are consistent with a wave-like phenomenon.

### 3.3.2. Rossby Wave Dynamics

The observations show that variability in the alongshore currents offshore of the SRR is due to a wave-like phenomenon with across-shore wavelength of a few hundred kilometers and period nearing 1 year that propagates westward, suggesting that Rossby wave dynamics may be important. To determine whether the observed westward propagation is consistent with Rossby wave dynamics, we compare the observed wave number and frequency with the theoretical Rossby wave dispersion relation. Since we are only considering westward propagation offshore of the SCB and SRR where the water depth generally exceeds 2000 m (Figure 1) and we will show in section 3.3.4 that the main forcing region is within the SCB, we consider only free Rossby waves dynamics in the absence of topographic effects.

In a coordinate system with the $x$ axis rotated by an angle $\alpha$ counterclockwise from east, the flat bottom, free, baroclinic Rossby wave dispersion relation is

$$
\omega = -\frac{\beta (k \cos \alpha - l \sin \alpha)}{k^2 + \nu + \frac{l^2}{\beta}}
$$

where $\omega$ is the angular frequency, $k$ is the across-shore wave number, $l$ is the alongshore wave number, $\nu$ is the local Coriolis parameter, and $\beta$ is the local rate of change of the
Coriolis parameter, and $c_n$ is the speed of the $n$th gravity wave mode [Gill, 1982, equation 6.11.18]. For Line 90.0, $\alpha = 26^\circ$. To estimate $c_n$ in the region, we combine glider observations with World Ocean Atlas 2005 climatologies of temperature and salinity [Antonov et al., 2006; Locarnini et al., 2006] to produce a full depth profile of the buoyancy frequency. We make a similar calculation of $c_n$ in the model by using the simulated stratification.

Since we observe only the across-shore component of the wave number, we must make some simplifications to (1) for comparison to the observations. For Rossby waves traveling directly along the survey line, and roughly perpendicular to the coast, we have $l = 0$, and the dispersion relation is

$$\omega = -\frac{\beta k \cos \alpha}{k^2 + \frac{\gamma}{\rho_W}}. \tag{2}$$

The dispersion relation (1) also provides an upper limit for the frequency $\omega$ at any given across-shore wave number $k$. For $\alpha = 0$, that limit is a wave propagating due west with $l = 0$, but in our rotated coordinate system the alongshore wave number that maximizes the frequency varies with the across-shore wave number. The case for $l = 0$ (2) and the maximum frequency case are shown for the first three baroclinic modes in Figure 11.

![Figure 11. Rossby wave dispersion relations (lines) and across-shore wave number ($k$) and frequency ($\omega$) of observed alongshore velocity at 300 m (circle), observed slope of the 26.5 kg m$^{-2}$ isopycnal (triangle), and simulated alongshore velocity at 300 m (diamond). The black lines are for the observed stratification; the grey lines are for the simulated stratification. The thin lines are for waves propagating along the observation line (2), and the bold lines show the maximum frequency for each across-shore wave number. The first (solid), second (dashed), and third (dotted) baroclinic modes are shown for the dispersion relation with observed stratification; only the first baroclinic mode is shown for the modeled stratification.](image)

[42] The wave number and frequency of the westward propagation of both the observed and simulated alongshore velocity and the observed isopycnal slope agree well with the theoretical Rossby wave dispersion relation (Figure 11). The observed frequencies lie between the dispersion curve for a first-mode baroclinic wave moving along the observation line (2) and the curve for a first-mode wave with maximum frequency, and they are far removed from the curves corresponding to the higher baroclinic modes. The across-shore wave number and frequency of westward propagation in the numerical simulation, though somewhat smaller than observed, are similarly consistent with first-mode baroclinic Rossby wave dynamics (Figure 11). The dispersion curves for the model are slightly different from those for the observed fields due to minor differences in stratification in the model.

### 3.3.3. Alongshore Wave Number and Direction of Propagation

Our observations cannot resolve the alongshore component of the wave number ($l$). However, if we assume that first-mode baroclinic Rossby dynamics accurately describe the westward propagation, then (1) provides a means of calculating $l$. Since (1) is quadratic in $l$, it gives two solutions for each observed value of $k$ and $\omega$, but one solution gives wavelengths of 900 km or longer. Since the westward propagation was not apparent off Point Conception, only 225 km up the coast, we consider only the larger solution for $l$, which gives shorter wavelengths. For the values of $k$ and $\omega$ corresponding to observed alongshore velocities and across-shore isopycnal slopes, the values of $l$ are $4.5 \times 10^3$ km$^{-1}$ (221 km wavelength) and $3.9 \times 10^3$ km$^{-1}$ (254 km wavelength), respectively. The resulting vector wave numbers for both alongshore velocity and isopycnal slope are oriented toward 292° true. The same calculation using modeled velocity at 300 m gives an alongshore wave number of $6.7 \times 10^3$ km$^{-1}$ (149 km wavelength) and a direction of propagation of 310° true.

### 3.3.4. Source of Westward Propagation

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The numerical simulation provides velocity information at all locations, so we can investigate the direction of propagation in the model without assuming anything about the dynamics by calculating velocity autocorrelations over a two-dimensional region. For the simulated currents at 300 m and in the region [122°W 119°W] × [30°N 33°N], we calculate the autocorrelation of velocity as

$$\text{cor}(\Delta x, \Delta y, \Delta t) = \frac{\langle \vec{u}(x, y, t) \cdot \vec{u}(x + \Delta x, y + \Delta y, t + \Delta t) \rangle_{x,y,t}}{\langle \vec{u}(x, y, t) \cdot \vec{u}(x, y, t) \rangle_{x,y,t}}. \tag{3}$$

Slices through the resulting autocorrelation matrix at fixed values of $\Delta t$ (Figure 12) show the area of highest correlation moving nearly due west with increasing temporal separation, in reasonable agreement with the 292° direction of propagation estimated from the observations with the assumption of first-mode baroclinic Rossby wave dynamics.

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Slices through the resulting autocorrelation matrix at fixed values of $\Delta t$ (Figure 12) show the area of highest correlation moving nearly due west with increasing temporal separation, in reasonable agreement with the 292° direction of propagation estimated from the observations with the assumption of first-mode baroclinic Rossby wave dynamics.
wind stress near the coast and positive wind stress curl within the SCB raised isopycnals within the SCB in spring. The relaxation of the wind forcing in summer allowed the density anomalies to propagate westward when model dynamics included the $\beta$ effect. Glider observations of the anomalous depth of the 26.5 kg m$^{-3}$ isopycnal show a similar uplift of the isopycnal in spring of each year along Line 90.0 with westward propagation of the density anomaly in the summer and fall (Figure 13b). Isopycnals shoal in spring at Point Conception, but there is no apparent westward propagation of the density anomalies in the summer and fall (Figure 13a). The timing of the setup of density anomalies within the SCB is then consistent with the annually periodic wind forcing suggested by Di Lorenzo [2003].

[46] In the absence of wind observations with sufficient coverage of the SCB, we turn to the numerical simulation and its adjusted wind stress field. Alongshore winds at Line 90.0 (not shown) are always equatorward with a maximum 200–400 km offshore and near zero velocity at the coast. Wind stress curl is positive inshore of the maximum with strongest positive curl inshore of the SRR (Figure 14a). Wind stress curl in the SCB is strongest in April–May (Figure 14b). Annual maxima in simulated wind stress curl within the SCB lead the observed shoaling of isopycnals within the SCB by about two months (Figure 14b). When the wind stress curl relaxes, isopycnal depth anomalies begin to propagate westward. The observed across-shore wave numbers (section 3.3.1) and the alongshore wave numbers estimated by assuming Rossby wave dynamics (section 3.3.3) give a spatial scale roughly equivalent to the size of the SCB. For forcing concentrated within the SCB as indicated by modeled wind stress curl, the resulting Rossby...
waves would be expected to have spatial scales similar to the size of the SCB [Debnath, 2007].

3.3.5. Rossby Waves or Eddies?

Recent work by Chelton et al. [2007, 2011] using SSH observations has shown that much ocean variability that had been attributed to linear Rossby waves [e.g., Chelton and Schlax, 1996] is actually due to nonlinear eddies that propagate westward with speeds close to the baroclinic Rossby wave speeds. Although westward propagating features in SSH typically have phase speeds faster than permitted by linear theory [Chelton and Schlax, 1996], the phase speeds that we observe (0.90 km d$^{-1}$ or 0.01 m s$^{-1}$) and model (1.03 km d$^{-1}$) are within the range allowed by the linear theory (Figure 11). The observed westward propagating features do not transport salinity along isopycnals (Figure 10), but nonlinear eddies with closed contours that form near the coast and move offshore across the background salinity gradient (e.g., Figure 8) would produce across-shore salinity transport. Though these findings are consistent with linear dynamics, the phase speeds are smaller than observed mean velocities in the region (Figure 4), so we may expect nonlinear effects to also be important, particularly as features evolve and move out of our observation region.

4. Conclusions

We combine long-term, high-resolution observations from a network of autonomous underwater gliders in the southern California Current System with a regional, numerical state estimate. This novel technical framework allows us to (1) characterize the narrow poleward flows in the region, particularly the recently detected poleward current offshore of the SRR, (2) show that the poleward current offshore of the SRR propagates westward, and (3) demonstrate that the observed westward propagation is largely consistent with linear Rossby wave dynamics.

We observe persistent poleward flows in three areas: within 100 km of the coast at Point Conception, inshore of the SRR within the SCB, and offshore of the SRR. We account for variability in the strength and position of poleward flows in the CCS by constructing averages in jet-following coordinates for each region of poleward flow. These jet-following averages reveal the poleward flows to be narrower and faster than suggested by long-term mean sections of alongshore flow. Mean poleward velocities are on the order of 0.1 m s$^{-1}$ in each region. Contrary to previous work, we find no distinction between subsurface and surface poleward flow within the SCB and, consequently, no support for separate naming of surface and subsurface currents. We show that the poleward current offshore of the SRR is deeper, denser, and saltier than the CU at Point Conception and that much of the poleward transport through the SCB and within the SCB. This westward propagation is apparent in both the velocity and density fields since the two are linked by the dominant geostrophic dynamics. Without the high-spatial- and -temporal-resolution observations provided by the network of gliders, we could not have detected these subsurface westward propagating signals.

The across-shore wave number, frequency, and phase speed of the westward propagation and the lack of westward transport of salinity along isopycnals are consistent with first-mode baroclinic Rossby wave dynamics. Observed isopycnal depth anomalies and adjusted wind stress from the numerical simulation show that local winds during spring and early summer raise isopycnals within the SCB. When these winds relax, density anomalies and the poleward current propagate westward. While consistent with linear dynamics, the relatively slow speed of propagation suggests that nonlinear effects are likely important.

Poleward undercurrents are a ubiquitous feature of midlatitude eastern boundary current systems, but the details of the flows vary by region [Neshyba et al., 1989]. By virtue of its location along the west coast of the United States, the CCS is by far the best observed eastern boundary current system. Without similarly thorough observations in other eastern boundary current systems, we can only speculate about the applicability of our findings to other regions. Typical poleward undercurrents in other eastern boundary current systems could be expected to be on the order of a few 10s of km wide with speeds near 0.1 m s$^{-1}$ as found in the CCS. Since eastern boundary current systems generally experience seasonally modulated wind forcing [Bakun and Nelson, 1991], westward propagation of density anomalies that subsequently affects the position of undercurrents may be found in other regions. The CCS and other eastern boundary current systems support highly productive ecosystems and associated fisheries [Botsford et al., 2006]. Strong and variable alongshore currents may be important to the distribution, dispersal, and retention of organisms in these systems. We anticipate that observations from networks of autonomous vehicles and continually improving numerical simulations will be essential to improving our understanding of eastern boundary current systems and their associated ecosystems.

Appendix A: Accuracy of Vertically Averaged Current Measurements From Gliders

Glider-based measurements of vertically averaged currents are used to reference ADP-derived shears and geostrophic shears throughout this analysis, so the accuracy of those measurements is critical to our results. For each glider dive, the true vertically averaged water velocity relative to the earth, $\vec{u}_w$, is related to the true vertically averaged glider velocity over the earth, $\vec{u}_g$, and the true vertically averaged glider velocity through the water, $\vec{u}_r$, by

$$\vec{u}_w = \vec{u}_g - \vec{u}_r,$$  \hspace{1cm} (A1)

where we have used complex variables for velocities with the eastward velocity as the real component and the northward velocity as the imaginary component. We denote our measurements of these three quantities as $\vec{u}_ws$, $\vec{u}_gs$, and $\vec{u}_rs$. Our measurement of the glider’s velocity over the earth, $\vec{u}_gs$, is based on GPS measurements at the beginning and end of each dive that have position accuracy of 10 m. Our measurement of horizontal glider velocity through the water, $\vec{u}_r$,
is based on an estimated angle of attack from a model of glider flight [see Sherman et al., 2001] and integrating measured heading, pitch, and vertical velocity over each dive, so it may be subject to accumulated errors in magnitude and direction. Since the glider’s vertical velocity is not perfectly constant, our measurements $u_v$, $u_p$, and $u_r$ are temporal averages during each dive rather than depth averages.

Previous authors have considered the main sources of error in calculating glider displacement through the water during a dive (e.g., vertical ocean velocities and errors in the model of glider flight) and found these factors to contribute to errors in vertically averaged current estimates of $O(0.01 \text{ m s}^{-1})$ for a single dive, with accuracy increasing when many dives are considered [Davis et al., 2002; Gourdeau et al., 2008]. With the exception of two deployments on Line 80.0 that failed to return complete data sets for detailed postdeployment calibration, all glider deployments used in this analysis have had heading-dependent compass calibrations applied, eliminating the largest source of error in the vertically averaged currents discussed by Gourdeau et al. [2008]. Eriksen et al. [2001] found a strong linear relationship with gain near unity between turnarounds to changes in angle of attack, $\gamma$, suggests an uncertainty in the angle of attack gives an increase in angle of attack of 0.50° between turnarounds. Directional bias in $\hat{u}_r$ remains negligible.

Increasing fouling during a deployment may affect the angle of attack of the glider. As fouling increases drag and disrupts flow over the glider, we would expect the glider’s angle of attack to increase and its vertical velocity to decrease, both of which cause its forward speed through the water to slow. Changes in vertical velocity are measured, but changes in angle of attack are not accounted for in our estimate of $\hat{u}_r$. For most deployments, the glider completes four transects, so two offshore turnarounds are available from the same deployment. The difference between values of $\gamma$ from the first and second offshore turnaround of each deployment gives some indication of the influence of fouling. For 16 deployments with two offshore turnarounds each, the mean change in $\gamma$ between the first and second turnaround is $-0.0263 + 0.0015i$. Attributing all of the change in $\gamma$ between turnarounds to changes in angle of attack gives an increase in angle of attack of 0.50° between turnarounds. Directional bias in $\hat{u}_r$ remains negligible.

Since the mean value of $\gamma$ found in this analysis is less than a standard deviation from unity, we have not applied any correction to vertically averaged velocity estimates. Any magnitude bias in $\hat{u}_r$ should result in errors in the along-track (across-shore) component of $\hat{u}_w$ which are not the focus of this analysis. In the mean over many sections, any bias should average to zero because the gliders survey in both the onshore and offshore directions.

Appendix B: Horizontal Currents From Glider-Mounted Acoustic Doppler Profilers

Each Spray glider deployed in the CCS is equipped with a custom Sontek Argonaut 750 kHz acoustic Doppler profiler (ADP) mounted in the tail. The instrument is oriented such that an upward pitch at the nominal 17° ascent angle and zero roll result in the central axis of the ADP pointing downward so that range bins are depth bins. The three beams of the instrument are aimed 25° off the central axis with one beam looking forward along the long axis of the glider.

The ADP collects a 16 ping ensemble average every 4 m in the vertical during the ascending portion of a dive (Figure B1a). Each ensemble average provides measurements of along-beam speed and return amplitude in five measurement cells for each beam. Each measurement cell extends 4 m in the vertical. For the settings used in the CCS, the first measurement cell is centered 10 m below the glider. The sampling parameters are such that cells from successive ensembles should align as indicated in Figure B1a.

Glider ADP measurements undergo several processing and quality control steps before profiles of ocean velocity are estimated. We calculate the depth of each measurement cell from records of the glider’s depth, pitch, and roll during each 16 ping ensemble. Data from ensembles when the glider is pitched or rolled enough to displace measurement cells from their nominal alignment are excluded from further
sampling pattern and estimation bins for glider ADP sampling. The position of the glider at the time of each of the 16-ping ensembles is shown by the black squares. For each ensemble, the glider measures along-beam speed and return strength in five measurement cells below the glider. The timing of ensembles is set such that measurement cells from successive ensembles align as indicated. The cells intersected by the black arrow are at the same depth and sort into the \( i \)th bin as indicated. Measurements from the shallowest cell for each ensemble (cell 1, light grey shading) are not used in estimates of velocity.

processing. The ADP can be used as an altimeter during the descending portion of a dive; data in cells that are deeper than the altimeter-derived bottom depth are excluded. Along-beam speeds are used to calculate eastward (\( u \)), northward (\( v \)), and upward speeds relative to the glider by successive rotations using records of the glider’s pitch, roll, and heading during each ensemble average. Velocities relative to the glider that exceed 0.75 m s\(^{-1} \) are considered to be erroneous since the speed of the glider through the water is approximately 0.25 m s\(^{-1} \) and the range of the ADP is too small (about 20 m) to expect very large relative velocities. We also exclude measurements for which the signal-to-noise ratio is less than 1.0.

63 Transducer failures have occurred during some deployments. We use the average return strength for each beam during each profile to detect failures of transducers. Any sudden drop in return strength of one beam relative to the other two indicates failure of the respective transducer, and data from that beam are not used in further calculations. The loss of data from one or more beams prohibits calculation of a velocity profile. Transducer failures are the primary cause of the missing ADP velocity profiles shown in Figure 2.

64 We let \( N \) be the number of ensemble averages during the ascending portion of a glider dive; the sampling geometry defines a set \( N + 4 \) estimation bins (Figure B1b). Measurements in the shallowest cell (cell 1) for each ensemble are not used because of ringing of the ADP transducers, so up to four measurements contribute to the estimate in each bin. The exclusion of data from cell 1 results in no data in the uppermost sampling bin, so we only estimate velocity in the \( N + 3 \) bins with data. The number of measurements contributing to the estimate in a bin is reduced if measurements are excluded during quality control. Because the glider sampling pattern is not perfectly regular, the depth of a given estimation bin is defined to be the mean depth of each good measurement in the bin.

65 The glider-mounted ADP functions similarly to a lowered acoustic Doppler current profiler (LADCP) deployed from a research vessel, and our calculation of ocean velocity profiles from the glider-mounted ADP data is based on the LADCP data processing scheme presented by Visbeck [2002]. For each valid measurement of horizontal water velocity relative to the glider, \( (u, v) \), we have an equation

\[
(u, v)_e = (u, v)_w - (u, v)_g,
\]

where \( (u, v)_w \) is the ocean velocity at the location of the measurement cell, and \( (u, v)_g \) is the velocity of the glider at the same moment. (Note that here the subscript \( r \) refers to water velocity relative to the glider, which is the opposite of the glider’s velocity through the water used in Appendix A.) Both terms on the right hand side of equation (B1) are unknown. There are \( N \) unknown glider velocities (one for each sampling depth), and \( N + 3 \) unknown water velocities (one for each estimation bin with data). Excluding data from the shallowest cell for each ensemble, we have at most \( 4N \) equations of the form of (B1). This system of equations can be written as a matrix equation of the form \( \mathbf{Gm} = \mathbf{d} \), where

\[
\mathbf{d} = [u_{1,2} \; u_{1,3} \; u_{1,4} \; u_{1,5} \; u_{2,2} \; \cdots \; u_{2,5} \; \cdots \; u_{N,5}]^T
\]

is the vector of observations of speed relative to the glider in one direction,

\[
\mathbf{m} = [u_{g,1} \; \cdots \; u_{g,N} \; u_{w,2} \; \cdots \; u_{w,N+4}]^T
\]

is the vector of unknown glider and water velocities in that direction, and

\[
\mathbf{G} = \begin{bmatrix}
-1 & 0 & 0 & \cdots & 0 & 1 & 0 & 0 & 0 & \cdots & 0 \\
-1 & 0 & 0 & \cdots & 0 & 0 & 1 & 0 & 0 & \cdots & 0 \\
-1 & 0 & 0 & \cdots & 0 & 0 & 0 & 1 & 0 & \cdots & 0 \\
-1 & 0 & 0 & \cdots & 0 & 0 & 0 & 0 & 1 & \cdots & 0 \\
0 & -1 & 0 & \cdots & 0 & 0 & 0 & 0 & 0 & \cdots & 0 \\
0 & -1 & 0 & \cdots & 0 & 0 & 0 & 0 & 1 & \cdots & 0 \\
0 & -1 & 0 & \cdots & 0 & 0 & 0 & 0 & 0 & \cdots & 0 \\
0 & -1 & 0 & \cdots & 0 & 0 & 0 & 0 & 0 & \cdots & 0 \\
\vdots & \vdots & \vdots & \ddots & \vdots & \vdots & \vdots & \vdots & \vdots & \ddots & \vdots \\
0 & 0 & 0 & \cdots & -1 & 0 & 0 & 0 & 0 & \cdots & 1
\end{bmatrix}
\]

is the matrix of coefficients when all measurements are good. When all measurements are used, \( \mathbf{d} \) has dimensions \( 4N \times 1 \) and \( \mathbf{G} \) has dimensions \( 4N \times (N + N + 3) \). Loss of measurements reduces only the number of equations in the system, so that, in practice, \( \mathbf{d} \) and \( \mathbf{G} \) have at most \( 4N \) rows, but \( \mathbf{G} \) always has dimension \( (N + N + 3) \times 1 \).

66 Though the number of equations exceeds the number of unknowns, we still require additional information to solve the system of equations since the ADP data alone can only provide the baroclinic portion of the ocean velocity [Visbeck, 2002]. We use the estimate of vertically averaged water velocity during each dive (Appendix A) to reference
the ADP shear. This measurement of vertically averaged velocity is valid from the surface to the maximum depth reached by the glider, a range that is offset from the sampling range of the ADP since the ADP samples below the glider. We account for this offset in two ways. First, we exclude ADP velocity estimates in the seven bins \( (\tau_{\text{bin}} = N - 2, \ldots, N + 4) \) that are deeper than the glider’s maximum depth (Figure B1b) from the constraint. Second, we assume that the near-surface portion of the water column that is not sampled by the ADP has uniform velocity. Under this assumption, we weight the uppermost estimation bin as if it extended to the surface in the vertically averaged velocity constraint. This constraint adds the row

\[
\begin{bmatrix}
0 & \cdots & 0 & \Delta z_2 & \cdots & \Delta z_{N-3} & 0 & \cdots & 0 \\
\end{bmatrix},
\]

(B5)

to the matrix \( \mathbf{G} \). The \( \Delta z_i \) are the vertical extents of the velocity bins, which are approximately 4 m, except for \( \Delta z_2 \) which is larger as discussed above. The corresponding element added to \( \mathbf{d} \) is \( U \Delta z_2^{-3} \Delta z_p \), where \( U \) is the estimated vertically averaged velocity. Since the ADP measures shear only on the ascending portion of each dive and vertically averaged velocity is based on the glider’s displacement throughout the entire dive, there is a mismatch in location and time between the shear profile and the barotropic constraint that is unaccounted for. At the 30 km and larger scales considered in this analysis, any errors due to this mismatch should not be significant. The agreement between ADP-derived currents after 30 h filtering and geostrophic currents (e.g., Figure 3) suggests that the induced errors are small.

Ideally, the overdetermined system \( \mathbf{Gm} \cong \mathbf{d} \) is now solvable by least squares techniques. However, the loss of equations due to bad measurements can make the system ill conditioned. To further constrain the problem and reduce noise in the solution, we apply the curvature-minimizing smoothness constraint of \( \text{Visbeck} [2002] \) to the horizontal ocean velocities and horizontal glider velocities. These constraints add \( N + 1 \) and \( N - 2 \) additional equations to the system, respectively. The additional rows added to \( \mathbf{G} \) are

\[
w \times
\begin{bmatrix}
0 & \cdots & 0 & -1 & 2 & -1 & 0 & \cdots & 0 \\
0 & \cdots & 0 & 0 & -1 & 2 & -1 & \cdots & 0 \\
\vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \ddots & \vdots \\
0 & \cdots & 0 & 0 & 0 & 0 & 0 & \cdots & -1 \\
\end{bmatrix},
\]

(B6)

and

\[
w \times
\begin{bmatrix}
-1 & 2 & -1 & 0 & \cdots & 0 & 0 & \cdots & 0 \\
0 & -1 & 2 & -1 & \cdots & 0 & 0 & \cdots & 0 \\
\vdots & \vdots & \vdots & \vdots & \ddots & \vdots & \vdots & \cdots & \vdots \\
0 & 0 & 0 & 0 & \cdots & -1 & 0 & \cdots & 0 \\
\end{bmatrix},
\]

(B7)

where \( w \) is a weight that determines the degree of smoothing. We choose \( w = 5 \) to produce sufficiently smooth velocity profiles. The data vector \( \mathbf{d} \) gains \( 2N - 1 \) rows of zeros since we seek to minimize curvature in the solution.

We then solve the system for the unknown glider velocities and horizontal ocean velocities using least squares techniques to minimize the \( L_2 \) norm of \( \mathbf{Gm} - \mathbf{d} \). The solution is

\[
\mathbf{m} = (\mathbf{G}^T \mathbf{G})^{-1} \mathbf{G}^T \mathbf{d}.
\]

(B8)

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