Altimeter-derived variability of surface velocities in the California Current System: 2. Seasonal circulation and eddy statistics

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Abstract

Satellite surface height and temperature fields are used to examine the seasonal surface circulation in the California Current System. In spring and summer, an equatorward jet develops next to the coast, with an initial latitudinal structure that responds to the latitudinal distribution of equatorward alongshore winds. This jet moves offshore from spring to fall and develops meanders and closed eddies that contribute eddy kinetic energy and water mass properties to the nearby deep ocean. Higher EKE appears first in spring near the coast and moves offshore with the seasonal jet. Along altimeter tracks parallel to the coastline 200 and 400 km offshore, the greatest EKE is found in summer and fall, respectively, with wavelengths of approximately 300 km. These are the wavelengths of the largest meanders and the distance between the largest eddies visible in altimeter “snapshots” of the jet. At 600–800 km offshore, the wavelengths are generally smaller (100–150 km), similar to eddy sizes reported by others for the quiescent northeast Pacific. Some energy with longer wavelengths arrives at the track 600 km offshore in winter and 800 km offshore in spring, which we interpret as the most offshore extension of the EKE generated by the jet and eddy system closer to the coast in the previous spring and summer. A sequence of snapshots from October 1992 to October 1993 shows details of the evolution of the jet and eddy system over one complete annual cycle. © 2000 Elsevier Science Ltd. All rights reserved.

1. Introduction

The California Current (CC) flows equatorward from approximately 50°N to 20°N, connecting the eastward North Pacific Current (also called the West Wind Drift) at...
approximately 50°N to the westward North Equatorial Current at approximately 20°N. The northern region of the California Current (off Oregon and Washington north of about 43°N) has a straight coast and two main sources of fresh-water input: the largest is the Columbia River at 46°N (Fig. 1), followed by the Strait of Juan de Fuca at 48°N between the US and Vancouver Island. Wind forcing in this region is moderately upwelling-favorable in summer and strongly downwelling-favorable in winter. The middle region (35–43°N) has a number of capes and the strongest seasonal contrast in winds, with strong and persistent upwelling-favorable winds in summer and poleward, downwelling-favorable winds during winter storms. The influence of storms decreases south of approximately 35–37°N, where monthly mean winds remain upwelling-favorable all year. The coast turns sharply eastward at 35°N, forming the Southern California Bight between 32 and 35°N, a region sheltered from the strong wind forcing found elsewhere in the system. Off Baja California between 22 and 32°N, winds are stronger than in the SC Bight and upwelling-favorable the year round, although weaker than the summer winds off northern California. There are several capes along Baja California, the most prominent at Punta Eugenia, 27–28°N (Fig. 1).
The surface circulation of the California Current System (CCS) on the largest scales has been described by climatologies of ship drift over the entire system or composites of surveys from cruises covering some portion of the system (Hickey, 1979; Chelton, 1984; Lynn and Simpson, 1987). These show broad, equatorward surface flow in summer (i.e., the normal California Current) and poleward flow in the Inshore Countercurrent (ICC) next to the coast in winter north of about 33°N. The Inshore Countercurrent is often called the Davidson Current in certain locations, but we use the notation of Lynn and Simpson (1987) to refer to a general current structure, no matter where it is found. Field surveys find a subsurface Poleward Undercurrent (PUC) over the continental slope during most seasons and locations, which the satellite data cannot detect. Depending on the degree of averaging, mean surface velocities are of order 5–15 cm s⁻¹. Individual hydrographic, ADCP and drifter surveys find much larger velocities, exceeding 1.0 m s⁻¹ in some cases (Paduan and Niiler, 1990; Brink and Cowles, 1991; Brink et al., 1991; Huyer et al., 1991; Strub et al., 1991). See Hickey (1979, 1998) and Huyer (1983) for more detailed reviews of the circulation in the CCS.

Strub and James (1995) used two years of Geosat altimeter data and sequences of sea surface temperature (SST) images to provide a preliminary estimate of the large-scale surface circulation in summer. Here we extend that analysis to look at the complete seasonal cycle using six years of altimeter data from three satellites: the first two years of Geosat data from the exact repeat mission (ERM), the first four years of TOPEX altimeter data, and ERS-1 data coincident with the first 15 months of TOPEX data. The TOPEX data in 1993–1994 coincide with a period of extensive field data off northern California (current moorings, drifter releases and hydrographic surveys), which were collected during an Office of Naval Research initiative. These have allowed verification of the accuracy and along-track spatial resolution of the altimeter-derived velocities (Strub et al., 1997), which gives us greater confidence in the interpretation of the remotely sensed data. Following our previous approach, (Strub and James, 1995), we combine the altimeter height fields with other satellite data (primarily SST) and relate these to past field observations to describe the seasonal evolution of the large-scale circulation in the California Current System, along with its eddy statistics (which include all temporal variability, whether caused by eddies, meanders or other structures).

2. Data and methods

2.1. Altimeter, tide gauge, SST and color data

Altimeter surface height fields are analogous to surface dynamic-height fields, calculated from hydrographic data (assuming the level of no motion is correct). However, the sea-surface heights measured by satellite radar altimeters include both the heights associated with geostrophic currents and the heights associated with the marine geoid. Since the marine geoid dominates the signal and is not well enough known to be removed independently, a temporal mean altimeter height
field is constructed and subtracted from each along-track data point during every cycle to eliminate the geoid signal. This also eliminates the mean height and velocity fields, leaving height and velocity “anomalies”. Contours of the height anomalies are approximate streamlines of the geostrophic velocity fields. True streamlines would account for the latitudinal gradient of the Coriolis parameter, which varies by approximately 30% between the center and the top/bottom latitudes.

To partially compensate for the loss of the circulation associated with the mean height field that is removed, a climatological mean dynamic-height field is calculated (relative to 500 m) from the long-term mean temperature and salinity fields of Levitus and Gelfeld (1992) and added to each height field. This climatological dynamic-height field depicts slow and broad equatorward flow, with a height field that slopes gently down toward the coast and represents equatorward velocities of approximately 3–4 cm s\(^{-1}\).

Altimeter data (processed by the NASA/NOAA PATHFINDER project) include approximately two years of Geosat data (November 1986 through October 1988), four years of TOPEX altimeter data (October 1992 to October 1996), and fifteen months of ERS-1 data overlapping the first TOPEX year (October 1992 to mid-December 1993). Data from the POSEIDON altimeter are not used due to higher noise levels present in this data set. The data have 6–7 km horizontal resolution along nadir tracks of the three satellites, presented in Fig. 1, where the higher cross-track resolution of the ERS-1 and Geosat orbit geometry is evident. During the 15 months of ERS-1 and TOPEX overlap, the highest spatial resolution is available and we make use of this to show “snapshots” of the height fields, centered on 35 days of altimeter data from the two altimeters. To remove offsets between the data sets before combining data from the different satellites, we remove the spatial mean from each cycle of each altimeter field over our area of interest (Fig. 1). By setting the spatial mean of each cycle to zero, we also remove the large-scale change in steric height associated with seasonal heating and any other long-term changes, leaving the gradients of height that show geostrophic circulation features with length-scales of 100–1000 km. The addition of the dynamic height climatology restores the mean height gradients at the largest scales.

Along altimeter tracks that cross the coast, data from the several pixels closest to the coast are usually flagged as suspect for various reasons, leaving a gap of 30 or more km. To extend the height fields to the coast, tide-gauge data are used from a number of coastal locations, separated by approximately 200–300 km (black filled circles in Fig. 1). These data are treated like the altimeter data, except that tides are removed by filtering, rather than using tidal models. In addition to the tidal filter, a 20-day filter is used to eliminate the effects of coastal trapped waves, which produce fluctuations in sea level with periods of around 10 days. After these filters are applied, the mean is removed over the same period as the altimeter data and the climatological dynamic height from the nearest available offshore location is added to the tide-gauge height anomalies. The tide-gauge data only affect the fields in the 50 km closest to the coast, removing “bulls eyes” in the nearshore height contours and providing smoother alongshore currents next to the coast.
Besides fields of height, we also present statistics of the cross-track geostrophic velocities, derived from the along-track slope of the altimeter height fields. Only the four-year record of TOPEX data is used to analyze the eddy statistics, since it is the least noisy and longest of the altimeter data sets. The velocities are calculated using the same method as in Strub et al. (1997), who conclude that these velocities are representative of measured velocities that are spatially averaged over approximately 80–100 km, with uncertainties of 3–5 cm s$^{-1}$.

Field data from summer off northern California often show colder water lying inshore of a jet that flows along a sharp gradient in SST. We make use of this relation by combining altimeter height contours with SST fields from the Advanced Very High Resolution Radiometer (AVHRR). These show the relation between the location of the temperature front and the seasonal jet. The climatological AVHRR data come from the Jet Propulsion Laboratory, consisting of monthly global fields with 18 km resolution during 1981–1986. The AVHRR data shown with altimeter snapshots are from individual images collected along the west coast with resolution of approximately 1–2 km. These higher resolution SST fields are not calibrated to absolute temperature but are simply radiance temperatures from the 11 μm band, which can be several degrees colder than absolute SST. We also show the seasonal cycle of satellite-derived surface pigment concentration fields, estimated from the Coastal Zone Color Scanner (CZCS). CZCS data extend from late 1978 to mid-1986 and have been discussed in detail by Strub et al. (1990) and others. Here they are presented to show the relationship between the seasonal jet and the more productive regions of the CC. These data come from the global dataset available from JPL, with approximately 20 km resolution.

The method of successive corrections (Bratseth, 1986; Vazquez et al., 1990) is used to construct two-dimensional horizontal fields of sea-surface height anomalies from the combination of altimeter and tide gauge data. In this method, heights are estimated at each point on a regular grid (0.5° is used here). A weighted mean of the data surrounding each grid point forms a first guess of the height at the grid point, which is modified during three iterations, reducing the spatial scale of the quadratic weighting function from 1.25° (first guess) to 1.0°, 0.75°, and 0.5° (Vazquez et al., 1990). In forming the snapshots, the data are also weighted in time, giving the data from the central time greater weight and using data from a period extending only 17.5 days from the central time in both directions. The resulting field is smoothed with a final Laplacian filter. This process produces a field that is slightly smoother, but similar in features, to a loess filter with half-power cutoff scales of approximately 1.4° and 35 days. The final smooth field has values everywhere that data exist within the largest spatial window, with slightly higher resolution in regions where the observations are more numerous. In the seasonal fields, the four years of more frequent but widely spaced TOPEX data define the large-scale circulation features. The Geosat and ERS-1 data contribute to the large-scale pattern and also provide an estimate of the smaller features (down to approximately 50–100 km). Thus, the larger-scale features in the seasonal height fields are defined by six years, while the smaller-scale features (less than 200 km) represent conditions during only three years. Features smaller than 100–150 km should be interpreted with caution.
2.2. Wind forcing — Coastal Upwelling Indices

An estimate of the alongshore surface wind forcing over the CC is provided by the Coastal Upwelling Index (CUI) produced by the NOAA/NMFS Pacific Fisheries Environmental Laboratory in Monterey, California (Bakun, 1975; Schwing et al., 1996). This index is constructed at coastal locations separated by $3^\circ$ of latitude every six hours. Geostrophic wind velocities are first calculated from the surface atmospheric pressure fields produced by the US Navy Fleet Numerical Meteorological and Oceanographical Center (FNMOC). The winds are rotated $15^\circ$ counterclockwise and reduced by a friction factor of 30%, then used to calculate the alongshore surface wind stress from a bulk formula with constant drag coefficient of 0.0013.

The offshore Ekman transport is then calculated, i.e., $M = (1/\rho f) \tau_a$, where $\rho$ is the density of water, $f$ is the Coriolis parameter (changing with latitude), and $\tau_a$ is the alongshore component of the wind stress. The Upwelling Index is the offshore transport, with units of $m^3/s$ per 100 meters of coastline. The daily average of the 6-h CUI values serves as an index of the alongshore wind stress (a value of 100 corresponds roughly to 0.1 N m$^{-2}$), modified by the latitudinal change in $f$. We use this index primarily to demonstrate that the seasonal cycle of large-scale, alongshore wind-forcing during the six years of altimeter data is relatively normal. It also describes wind forcing during 1992–1993, when the snapshots of the evolving seasonal circulation are presented.

3. Results

3.1. Surface wind forcing

Fig. 2a shows a 29-year (1967–1996) mean annual cycle of the CUI along the US west coast. Poleward winds (negative values, darker shading) occur in fall and winter north of $35^\circ$N, stronger in the north, while equatorward winds (positive values, lighter shading) are dominant from late April through September, strongest south of $40^\circ$N. A systematic error in the CUI (Bakun, 1975) places the maximum upwelling in summer at $33^\circ$S, while measured summer winds are maximum between $38^\circ$–$40^\circ$N and are weak near the coast between $32^\circ$–$35^\circ$N in the sheltered SC Bight (Halliwell and Allen, 1987). This systematic error does not affect our use of the upwelling index to assess the “normalcy” of the period of altimeter data and to describe the large-scale forcing.

In Fig. 2b, a mean annual cycle of the CUI is presented for the six years when altimeter data are available. During this period, poleward winds were generally weaker in early fall and stronger during much of winter, while equatorward winds were weaker in spring and summer. The magnitudes of the anomalies (not shown) represent only 12–25% differences from the 29-year means for this 6-year period, some of which is caused by a long-term weakening trend in the summer CUI. The source of this trend (real winds or model) is not known.
Fig. 2. Time-latitude plots of daily Coastal Upwelling Indices: the annual cycle from late-September to early October. A 10-day running mean has been applied to all time series before further processing. Upwelling conditions (equatorward winds) are indicated by light shading, downwelling conditions by darker shading: (a) Twenty-nine year mean from October 1967 through September 1996. (b) Six-year mean from October through September, 1986–1988, 1992–1996. (c) The single year from October 1992 to October 1993. Units are cubic meters per second per 100 m of coastline.
Fig. 3. Mean seasonal height fields (two-month means) calculated from altimeter plus tide gauge data. The long-term mean dynamic height relative to 500 m has been calculated from the Levitus and Gelfeld (1992) temperature and salinity climatology and added to this and all other altimeter height fields. In the climatological dynamic height field, heights between 84 and 90 cm (shaded) enter the California Current, while lower dynamic height contours enter the Alaskan Gyre.
3.2. Mean seasonal circulation, SST and surface pigment concentration

Fig. 3 presents two-month mean surface height fields, constructed from the six years of altimeter and tide gauge heights plus a mean annual dynamic height field (Levitus and Gelfeld, 1992). As stated above, features of approximately 100–150 km and less should be cautiously interpreted, as they derive from the shorter three-year record of more closely spaced Geosat and ERS-1 tracks (Fig. 1). The region between 84 and 90 cm (shaded) shows the approximate path of water brought into the CCS from the North Pacific Current. Although temporally changing streamlines do not show the exact paths of individual parcels, they provide a general indication of which locations are likely to be influenced by water properties brought in by the “West Wind Drift” (Section 4). We note again that the height contours are only approximate geostrophic streamlines.

The mid-winter (January–February) pattern shows equatorward flow in the CCS far offshore south of 47°N. High sea levels next to the coast between approximately 38–52°N correspond to the poleward ICC. Hydrographic data from the California Cooperative Fisheries Investigation (CalCOFI) data set show poleward flow extending from 33–37°N in winter (Lynn and Simpson, 1987), the northern limit of their analysis. Our previous analyses of tide gauge data also show high sea levels to extend along the entire coast (33–48°N) in early winter (Strub and James, 1988). Thus, we consider it likely that the altimeter fields miss the poleward flow between 33 and 37°N, due to frequent Geosat data dropouts in winter and coarse spatial sampling by TOPEX. The snapshot of heights constructed from TOPEX and ERS-1 data in
January 1993 (Section 4.4) shows the ICC extending from 31°–52°N. In Fig. 3a, we believe that the altimeter depiction of the ICC in winter is correct north of 38°N, but that the poleward flow originates at least as far south as 33°N.

In late winter and early spring, lower sea levels and equatorward flow next to the coast appear first in the south (Lynn and Simpson, 1987; Strub et al., 1987a; Strub and James, 1988; Harms, 1996; Harms and Winant, 1998). The March–April field of altimeter heights reproduces this southward flow through the Southern California Bight and demonstrates its strong connection to equatorward currents next to the coast off central and northern California, with a weaker connection to currents as far north as 44°N. In May and June, the equatorward flow originates as far north as Vancouver Island, while sea levels next to the coast begin to rise in the south and form the cyclonic circulation typical of summer conditions in the SC Bight.

The concentrated equatorward jet migrates 100–300 km offshore in mid-summer (July and August) and extends from Vancouver Island (51°N) to southern California, where part of it enters the southern end of the strengthened cyclonic gyre that fills the SC Bight (31°N). The jet remains closer to the coast off Oregon and Washington but meanders far offshore of the California coast. This six-year mean does not show the rich field of meanders and eddies that appear in individual altimeter and SST fields presented below and in regional studies (Rienecker and Mooers, 1989; Brink and Cowles, 1991; Strub and James, 1995). The six-year mean summer altimeter field does capture the general placement of the equatorward jet, including its greater offshore displacement south of approximately 43°N. The shading indicates that water from the central north Pacific, carried in the North Pacific Current, is kept on the offshore side of the jet and first intersects the coast within the SC Bight (see discussion below). It also shows the observed beginning of poleward flow off central California in summer, while winds continue to be equatorward (Huyer and Kosro, 1987; Strub et al., 1987a; Winant et al., 1987; Chelton et al., 1988; Largier et al., 1993).

Fields from September–October and November–December show the transition from mid-summer to mid-winter. Poleward currents develop off northern California in September and October as the equatorward jet continues to move offshore. The cyclonic circulation in the SC Bight becomes weak and poorly defined. In November and December, winter storms begin in the north and the strong, poleward ICC reappears north of approximately 40°N. The equatorward jet that began next to the coast in spring becomes the offshore, equatorward part of the “U-shaped” flow constituting the winter circulation pattern drawn in schematic form by Hickey (1989, 1998).

In Fig. 4, mean three-month seasonal height fields have been constructed from the altimeter and tide gauge data and overlaid on similar three-month seasonal cycles of SST, using AVHRR data from 1981 to 1986. In these three-month seasons, winter is composed of January–March, etc. In Fig. 5, the same three-month mean height fields from Fig. 4 are overlaid on three-month seasonal mean CZCS fields from 1979–1986, providing estimates of the seasonal changes in the patterns of surface pigment concentrations. The absolute values of pigment concentrations from the CZCS are only accurate to a factor of 2 or worse (Strub et al., 1990; Chavez, 1995), but we are primarily interested in seasonal changes in the patterns of concentration, not the absolute values.
Fig. 4. Mean seasonal height fields from the California Current (three-month means) overlaid on mean seasonal SST fields from AVHRR satellite data, 1981 through 1986. Winter is January–March, etc. Contour intervals are 2 cm.
The height fields in Fig. 4 include the prominent features from Fig. 3. The SST fields in winter and early spring generally consist of zonal bands, due to the north–south gradient of seasonal heating. Thus, the east–west orientation of isotherms is not aligned with the more north–south orientation of the height contours in winter. As water is warmed in spring and summer, upwelling replaces nearshore water with colder water from below and maintains the pattern of colder water in the region next to the coast. The equatorward jet occurs along the density front that coincides with the edge of the colder water, most evident in Fig. 4c (the colder water in Fig. 4b is hidden by the height contours). Both the front and the jet continue to move offshore in fall. The jet is more diffuse but still evident in the height field in winter.

In Fig. 5, highest pigment concentrations are found next to the coast in spring and remain inshore of the equatorward jet in summer. The combination of Figs. 4 and 5 shows that in summer the jet separates colder, richer upwelled water from the offshore, warm and oligotrophic water. A similar relationship between phytoplankton biomass and an alongshore jet has been documented on smaller scales in field studies along the coastal upwelling jet off Oregon (Small and Menzies, 1981) and around the mesoscale jet off central/northern California (Hayward and Mantyla, 1990; Chavez et al., 1991; Kosro et al., 1991). These satellite data make it clear that this relationship is a basic characteristic of the CC in summer from southern California to the Canadian border. They also show that warm and oligotrophic offshore water is carried into the southern part of the SC Bight by the cyclonic turn of the current, while cooler and richer water is advected into the northern part of the Bight. This sets up north–south gradients of temperature and chlorophyll in the Bight, which may provide conditions favorable for the creation of strong temperature and chlorophyll fronts in the Bight, as observed by Pelaéz and McGowan (1986), Thomas and Strub (1990) and others.

As the jet moves offshore in fall, the region of higher surface pigment concentrations also moves offshore and appears to become less concentrated. Some of this apparent pigment may be detrital products of dead phytoplankton, but it still serves as a tracer of the more productive waters (through recycled nutrients). The spatial relationship between high pigment and contours of height persists longer than that of cold SST and height, since isopleths of high pigment do not take on the more zonal orientation of SST in fall and winter. Mechanisms that may maintain the higher pigment concentrations within and inshore of the jet include geostrophic adjustment in meanders and eddies, raising the nutricline into the euphotic zone (Hayward and Mantyla, 1990) and an expanding field of positive wind stress curl, creating upwelling through Ekman pumping. A seasonal expansion of the positive wind stress curl is evident from February through September in the climatology presented by Bakun and Nelson (1991), while Strub et al. (1990) find a small but significant correlation between wind stress curl and CZCS surface pigments.

3.3. Eddy statistics of the seasonal circulation

Using only the four years of high-quality TOPEX data, we examine the temporal variability in the geostrophic currents, noting that “eddy statistics” from the two years
Fig. 5. Mean seasonal height fields from the California Current (three-month means, as in Fig. 4) overlaid on mean seasonal surface pigment fields from the CZCS satellite, late 1978 through mid-1986.
of Geosat data produce similar results, but with much higher levels of background noise. Assuming velocities to be constant over the 10-day TOPEX repeat cycles, vector velocities can be calculated at crossover points for every cycle. These provide estimates of the eddy kinetic energy (EKE) and principal axis (variance) ellipses. To test the effects of possible aliasing by the ten-day sampling, Strub et al. (1997) compare statistics from TOPEX velocities to those from current meters located under a TOPEX crossover and show that: (1) the altimeter ellipses are representative of those from surface velocities averaged over distances of approximately 100 km; (2) altimeter estimates of EKE at crossover points agree well (within 10%) with those from current meters; (3) variances of the altimeter cross-track geostrophic velocities agree just slightly less well (within 20–30%) but give more spatial information on seasonal changes in velocity variance than available by only sampling at crossovers; and (4) cross-track velocities along altimeter tracks give a reliable estimate of the size of velocity fluctuations with wavelengths greater than 60–80 km.

In contrast to the accuracy of the above statistics, frequency spectra calculated using the altimeter vector velocities at crossover points are strongly contaminated by aliased errors in the modeled diurnal tides, which create fictitious signals with periods near 60 days (Schlax and Chelton, 1994; Strub et al., 1997). For this reason, frequency spectra from the altimeter data are not presented here. Strub et al. (1997) qualitatively estimate periods of 100–150 days for fluctuations in 1–2 year time series of the measured and altimeter velocities at a point 400 km offshore of northern California, while Kelly et al. (1998) find a similar dominant period of 120 days in frequency spectra of the same measured currents used by Strub et al. These long periods validate the assumption that velocities are relatively constant over periods of 10 days (above).

Statistics calculated from the complete four-year data set, including the seasonal variability, are shown in Fig. 6. The four-year mean principal axis ellipses (Fig. 6a) show the spatial pattern in changes in EKE (the areas of the ellipses) and the surface velocity polarization (differences between major and minor axes). Variance ellipses with major axes larger than those that would result from 7 cm s\(^{-1}\) random noise in the cross-track velocities are darkened. Strub et al. (1997) estimate the likely level of RMS “noise” in the TOPEX velocities to be 3–5 cm s\(^{-1}\). Thus the darkened principal axis ellipses are significantly larger than expected from random noise in the data.

The spatial pattern of four-year mean cross-track velocity variances are shown in Fig. 6b by the lengths of the cross-track bars. If velocities were isotropic, these bars would be equivalent to EKE. Fig. 6a indicates that variability along the major axis at the more energetic locations is usually no more than 2 times as energetic as along the minor axis. A factor of two difference will cause errors of only 25–33% in estimates of EKE from a single cross-track variance. Figs. 6a and b provide evidence that the highest energy in temporally varying currents occurs within approximately 600 km of the coast south of 44\(^{\circ}\)N and within 100 km of the coast north of 44\(^{\circ}\)N, with maximum energy of around 400 cm\(^2\) s\(^{-2}\) at approximately 38\(^{\circ}\)N, 126\(^{\circ}\)W. Offshore of 130\(^{\circ}\)W, the ocean is extremely quiet.

Fig. 6c presents along-track wavenumber spectra (in energy preserving form) from the four descending tracks that parallel the coast between approximately 34–41\(^{\circ}\)N (separated by approximately 200 km). All wavenumber spectra are calculated from
Fig. 6. Statistical properties calculated from the full four years of TOPEX data. (a) Variance ellipses at crossover points (blackened ellipses are considered statistically different from those produced by 7 cm s$^{-1}$ noise in cross-track velocities); (b) Variances of the cross-track velocities at all points along the TOPEX tracks; (c) Wavenumber spectra for the full descending tracks closest to the coast (parallel to the section of coast between 34 and 41°N) (wavenumbers in cycles per km).
cross-track velocities on the sections of the tracks between 25 and 50°N. These include the more energetic region between approximately 30–45°N and less energetic regions farther north and south. Each track has approximately 120 realizations, with a nominal Nyquist wavelength of approximately 12 km. However, the smallest alongtrack wavelength that can actually be resolved is approximately 80 km, due to the low-pass filtering characteristics of the finite difference geostrophic velocity calculation (Strub et al., 1997). The units of wavenumber (κ) are (cycles km⁻¹); the corresponding length-scale, κ⁻¹ is the wavelength (km cycle⁻¹).

Along the two tracks approximately 200 and 400 km from the coast, there is a clear peak of energy with wavelengths of approximately 300 km, the level of energy decreasing by approximately 30% from 200 to 400 km offshore. The energy levels decrease further along tracks that are 600 and 800 km from the coast. The dominant wavelengths also reduce to approximately 150 km along the two more offshore tracks. Energy levels and wavelengths along the shorter track next to the coast south of 33°N (not shown) are similar to those from the longer track closest to the coast between 34 and 41°N (Figs. 6b and c). This suggests that the energy levels and spatial characteristics of features causing the velocity variability are similar at distances 200 km from the coast between 25 and 45°N. Figs. 6a and b suggest further that energy levels at a distance of 200 km from the coast decrease north of 45°N. The 300-km wavelengths found 200 and 400 km from the coast represent the size of the most prominent meanders in the alongshore jet and the distance between closed eddies of the same rotational direction surrounding that jet (next section). While the energy in features with wavelengths of 300 km continues to decrease as one moves from 400 to 800 km offshore, the energy in features with wavelengths of approximately 150 km remains constant and thus becomes dominant. This smaller wavelength is similar to the typical 100–140 km “eddy scales” found by Stammer (1997) in the NE Pacific offshore of the CCS, using gridded TOPEX data. We interpret the shift in wavelengths and energy along our TOPEX tracks to indicate the transition from the larger meanders and eddies in the more energetic CCS to the background EKE in the much less energetic deep ocean of the NE Pacific.

Within-season variability is addressed by removing the seasonal mean of the cross-track velocities from each point during each season. The seasonal changes in the within-season variability are then examined in two ways. In Fig. 7, the cross-track variances are shown for each three-month season along each track, after removing the mean velocity of the season at each point. Principal axis ellipses at crossovers are also shown using the same data. The along-track variances provide greater spatial coverage than the ellipses, which can be formed only at the widely separated crossovers. For instance, Fig. 7b shows an increase in energy next to the coast along the most nearshore portions of the cross-shelf tracks between 25 and 40°N in spring. This is not easily seen in ellipses at the crossover points closest to the coast (along the track 200 km offshore) north of 34°N. To show the wavelength characteristics of this energy, Fig. 8 presents wavenumber spectra of the same velocities (seasonal mean removed) along the four tracks shown in Fig. 6c (approximately thirty cycles in each season). Together, Figs. 7 and 8 provide insight into the energy carried by the meanders and eddies as the entire system moves into the more quiescent offshore deep ocean.
Fig. 7. RMS values of cross-track geostrophic velocities at each point along TOPEX tracks for each 3-month season. Principle axis ellipses are also drawn for each season. Based on four years of TOPEX data (no POSEIDON data used).
The track 200 km from the coast is offshore of most of the springtime seasonal increase in energy noted in Fig. 7b. The spring wavenumber spectrum along this track (Fig. 8b) shows only a slight increase in energy (area under the spectrum), although its dominant wavelength moves from about 200–250 km in winter to 300 km in spring,
more characteristic of its mean value. This may be the first sign of the new jet that is forming inshore of the track. While the velocity variances next to the coast begin to increase in spring (Fig. 7b), the energy 400 km offshore reaches its seasonal minimum, as shown by the area beneath its springtime wavenumber spectrum (dashed line in Fig. 8b). Figs. 7b and 8b combine to depict the region offshore of 200 km in spring as fairly uniform and low in energy.

In summer within 400 km of the coast, variance values appear to reach their peaks (Fig. 7c), with a maximum of more than 600 cm$^2$ s$^{-2}$ located 200–300 km offshore of Point Arena (39°N) and secondary maxima (200–300 cm$^2$ s$^{-2}$) off Punta Eugenia (27–30°N) and southwest of Point Conception (31–33°N, 122–124°W). These locations are similar to locations of maximum EKE reported from one year of Geosat data (White et al., 1990). The summer maximum in energy along the track 200 km from the coast clearly appears in its wavenumber spectrum (Fig. 8c), which shows a wider range of wavelengths (150–300 km) than during other seasons. Along the track 400 km offshore, the spectrum shows a significant increase in energy in summer, compared with spring, with wavelengths of 250 km. Fig. 8c also makes it easier to see the summertime seasonal minimum in energy along the track 600 km offshore.

In fall, the maximum energy values decrease 200 km offshore and the wavelength range narrows around the 300 km peak. Energy levels actually increase slightly 400 km offshore (the area under the spectrum in Fig. 8d) and the dominant wavelength increases to 300 km. We interpret this to show the arrival of energy in the core of the meandering jet and eddy system 400 km offshore in fall. Along the track 600 km offshore, energy increases in both the 300 and 150 km wavelength bands in fall. In winter, Figs. 7a and 8a show that energy 200 km from the coast is at a seasonal minimum with shorter wavelengths (200–250 km). Winter energy levels 400 km from the coast at wavelengths of 300 km are still near their peak (fall) values but the range of wavelengths is narrower, similar to the narrowing of the peak between summer and fall 200 km from the coast. Winter energy levels with wavelengths of 300 km reach a peak along the track 600 km from the coast, causing an overall seasonal maximum in energy. We interpret this to show the winter arrival of energy from the jet, 600 km offshore.

There is little apparent seasonal change in the low energy with wavelengths of 150 km along the track 800 km from the coast (Fig. 7). This track lies in the low-energy region often portrayed as representative of eastern subtropical gyres (Stammer, 1997). Careful examination of the wavenumber spectra in Fig. 8, however, reveals slightly higher energy levels with wavelengths between 300 and 400 km along this track in spring and summer. This could be interpreted as the continuing westward movement of energy generated in the seasonal jet a year earlier near the coast. If this interpretation is correct, the amount of energy appearing 800 km offshore has been reduced to only a few percent of that present 200 km from the coast in summer.

3.4. Snapshots of the large-scale circulation in 1992–1993

Fig. 9 presents individual altimeter height fields from TOPEX/ERS-1 data in 1992–1993 (centered on 35-day periods), overlaid on satellite SST fields.
features (less than 150 km) are better represented by the SST data (when the features have an SST signature), since irregular altimeter sampling in space and time may misplace features of this size and cause local errors of 5–10 cm in height (Greenslade et al., 1997). The altimeter data give the sense of the circulation (direction and strength), resolve the larger features, and qualitatively demonstrate the nature of the flow (meandering jets, cut-off eddies, etc). Agreement of the two types of data (jets along fronts, cold [warm] SST in cyclonic [anticyclonic] eddies, etc.) increases confidence in the inferred circulation of the features. The combined fields, along with results from coincident and previous field studies, show the types of features that create the mean seasonal cycles and statistics presented above. The alongshore wind stress during September 1992–October 1993 is indicated by the CUI in Fig. 2c, after application of a 10-day running mean to bring out the seasonal patterns from the 1–3 day synoptic variability.

In October 1992 (Fig. 9a), the SST field shows that cold filaments extend 400–500 km offshore of northern California and Oregon, with offshore-onshore cyclonic circulation indicated by the altimeter height fields near the ends of filaments at 38°N, 40°N and 44°N, creating wavelengths of 200–400 km. The main equatorward jet lies approximately 200–400 km offshore, with height differences of 10–20 cm across the jet. Within 100 km of the coast, flow is primarily poleward over the entire region (25–52°N). Fig. 2c shows that there were only weak poleward winds in late September/early October 1992, confined to the region north of 40°N. Thus, the October flow field presents an example of the tendency for poleward flow next to the coast in late summer and early fall, even before the onset of strong poleward winds.

Poleward winds become quasi-continuous (remembering the 10-day filter used on the upwelling index) in the north by mid-October and extend as far south as 30°N in January (Fig. 2c). The strongest poleward winds occur at the end of January. The altimeter heights in Fig. 9b show a continuous poleward ICC next to the coast in mid-January from 31 to 52°N. A strong jet (10–20 cm height difference) flows from northern California to north of Vancouver Island, connecting the subtropical and subarctic gyres along the boundaries. This current carries relatively warmer water northward off Oregon and Washington. Offshore of the poleward jet, meandering equatorward flow is still evident south of 45°N, at a distance of approximately 500–600 km offshore. The meanders continue to have wavelengths of 200–400 km, similar to peaks in the wavenumber spectra.

Winds become weakly equatorward south of 40°N in late February and strengthen in March and April, while winds continue to be poleward (strongly so at times) north of 40°N. The altimeter and tide gauge height fields depict a strong equatorward jet next to the coast in mid-March south of 40°N (not shown), about the time

Fig. 9. Snapshots of the altimeter (T/P and ERS-1) surface height fields from single 35-day ERS-1 cycles, overlaid on SST fields (11 μm radiance temperatures) from one or two AVHRR images during the cycle. Dates for the AVHRR images and the central date for the altimeter fields are shown. The contour intervals are 5 cm and color bars show the temperature palettes, which change with seasonal changes in SST. The black dashed line at 45°N in Fig. 9a is caused by missing data.
Fig. 9 (continued)

AVHRR:
Apr 25 & 27, 1993

Altimeters:
Apr 28 1993

T °C
18 -
16 -
14 -
12 -
10 -
8 -
AVHRR:
Aug 26 & 27, 1993

Altimeters:
Aug 26 1993

Fig. 9 (continued)
equatorward winds strengthen in the same region (Fig. 2c). The currents remain poleward north of 40°N, as do the winds.

In late April, the SST image in Fig. 9c clearly shows cold, upwelled water inshore of the equatorward jet between 35 and 40°N (the image is obscured by clouds north of 40°N). There are several small (50–100 km) undulations and colder regions in the band of cold water next to the coast between 35 and 39°N, which are not resolved by the altimeter. Enlargements of the SST features (not presented) more clearly show cyclonic curvature and are similar in size and location to SST features found in spring in other years (Kosro et al., 1991; Strub et al., 1991). Cold water is advected from the north into the SC Bight south of Point Conception to about 32.5°N, where a long filament extends to the southwest. The altimeter indicates the cyclonic rotation around the end of this filament near 31.5°N, 122.0°W. This cyclonic feature persists over the next several months, along with the anticyclonic feature to its southeast. Warmer water is advected into the Bight between 30 and 32°N, south of the filament. Hand contouring (rather than by computer) might have extended the inshore 90 cm contour around the cyclonic

In late-April and early-May, equatorward winds extend all the way to 48°N, during an event that appears to mark the “spring transition”, but is followed by a 3-week period of poleward winds north of 37°N (Fig. 2c). Wind stress continues to be upwelling favorable south of 37°N but is only half the “normal” strength. During this period (not shown), the heights next to the coast off northern California rise 15 cm from late-April to mid-June, weakening, but not eliminating, the equatorward flow south of 40°N.

Strong equatorward winds resume in June, extending from 30 to 48°N and remain upwelling-favorable over the entire region until October (Fig. 2c). The altimeter height field from late June (Fig. 9d) indicates an equatorward jet that extends from Vancouver Island to the SC Bight. Cold, upwelled water extends from Cape Blanco (43°N), where it is confined to a narrow coastal band (less than 25 km wide) to Point Conception, where it continues to be advected into the northern SC Bight. Eastward flow into the SC Bight is concentrated between 30 and 32°N (clouds obscure the SST in this region). The cyclonic and anticyclonic meanders near 28–32°N, 120–123°W are in locations similar to the eddies in the April field.

A number of filaments are evident south of Cape Mendocino (40°N) in June. To the west of Point Arena (39°N), the flow appears to go around a cyclonic eddy. This feature is actually an eddy dipole centered at approximately 38.5°N, 125°W, with a cyclonic eddy offshore of an anticyclonic eddy, as revealed by a June research cruise (Huyer et al., 1998). A second survey in August found a similar dipole eddy near 37–38°N, 126–128°W (Huyer et al., 1998). The altimeter and SST fields from late August (Fig. 9e) show a large cyclonic meander and eddy near 37–38°N, 127–128°W around this eddy pair. Gridding the altimeter height data with 0.2° grid spacing (rather than the 0.5° used here) resolves the eddy pair (not shown).
maintain their positions relative to each other as the system moves offshore between June and August.

In August (unlike June) upwelled water is evident next to the coast off Oregon (43°–46°N), but tapers to a very narrow coastal band north of 45°N in Fig. 9e. The region of cold water widens south of Cape Blanco (43°N), where a long filament extends 200–300 km to the west. This narrowing of the upwelling region north of Cape Blanco and expansion south of the cape in summer are consistent with the climatological mean SST field in Fig. 4c, as is the cyclonic meander and filament west of Cape Mendocino (40°N) and southwest of Point Arena (39°N). The filaments between 38 and 40°N in August and between 38 and 39°N in June demonstrate how meanders in the jet pull cool filaments offshore on the northern sides of cyclonic meanders, the cold SST becoming less visible as it travels around the meander and moves back onshore, perhaps due to subsidence of the cold water and surface heating during the journey (Brink and Cowles, 1991). The largest meanders and filaments appear to be associated with the largest capes, as suggested by some numerical models (Ikeda and Emery, 1984; Haidvogel et al., 1991; Batteen, 1997). The warm anticyclonic meander west of Point Arena is similar in location to the anticyclonic eddy reported by Lagerloef (1992). Evidence that it repeatedly appears on different years is given by the narrowing of the mean summer SST field (Fig. 4c) in this location. The wavelengths of the largest meanders in August continue to be approximately 200–400 km.

A cold cyclonic eddy can be seen separating from the flow in Fig. 9e, southwest of Cape Mendocino (39.5°N, 128°W), offshore of the core of the jet. Another separating cold eddy northwest of Cape Blanco (43.5°N, 126.5°W) is indicated by the SST but is not well resolved by the altimeter field (other than a divergence of the 80 and 85 cm isopleths). These eddies contribute colder and richer water properties from inshore of the jet to the deep ocean, along with their greater EKE. The cyclonic and anticyclonic meanders southwest of Point Conception in August are in the same locations as in June, with additional meanders downstream. While previous studies have documented “recurrent” anticyclonic eddies in this region in summer (Simpson et al., 1986), a series of cyclonic and anticyclonic meanders and eddies can be found in the area between 28 and 35°N, 118°–126°W at most times of the year (Fig. 9). In spring and summer, these locations are repeatable enough on different years to be visible as cyclonic and anticyclonic meanders southwest of Point Conception in the six-year mean height fields of Fig. 4b and c.

The fields from late October 1993 (Fig. 9f) have some features in common with those from October 1992 (Fig. 9a). Most of the circulation features next to the coast show poleward flow. The most concentrated equatorward flow occurs 300–500 km from the coast south of 43°N, roughly following the outer edge of the convoluted region of cool water farther inshore. The largest convolutions occur to the southwest of Cape Mendocino (40°N) and Point Arena (39°N) (wavelengths remaining around 200–400 km) and can be traced back in time to features (smaller meanders) that began close to the coast. West of Point Arena, a dipole eddy creates a mushroom structure in the SST between 37 and 40°N, 129°W. This dipole (anticyclonic to the north, cyclonic to the south) is not the same as the dipole found inshore of the jet during June and August at around 38–39°N (the cyclonic member of that pair still sits inshore of the jet
near 36°N, 126°W). Rather, it appears to have evolved from the interaction of the cyclonic eddy which separated from the jet southwest of Cape Mendocino in August and anticyclonic flow north of the separated cyclonic eddy. By creating this new eddy dipole, the separating cyclonic eddy has formed a structure that is efficient in transporting water mass characteristics from the jet to the deep ocean.

4. Summary and discussion

4.1. Seasonal evolution of the structure in the CCS — A summary schematic

The snapshots in Fig. 9 demonstrate how the offshore expansion of the meandering jet creates the seasonal evolution of mean fields and eddy statistics presented in Figs. 3–8. In Fig. 10 we offer a schematic diagram of the seasonally variable surface circulation in the California Current System, based on the results presented above and on previous studies. This depicts the poleward ICC next to the coast in winter, stretching from the SC Bight to north of Vancouver Island. Beginning in spring, an equatorward jet is created next to the coast that migrates offshore from spring to winter.

The seasonally migrating jet is not smooth, as depicted by the mean fields in Figs. 3–5, but highly meandering, with (meridional) wavelengths of around 300 km (Fig. 9). These dominant wavelengths help to differentiate the EKE associated with the jet from the background EKE in the NE Pacific, which has wavelengths of 120–150 km. At a fixed point 400 km offshore, the periods associated with these meanders and jets are qualitatively estimated to be 100–150 days by inspection of altimeter and current meter velocity time series (Strub et al., 1997) and quantitatively estimated to be 120 days by frequency spectra of the same current meter records (Kelly et al., 1998). These long periods are created by the slow movement of the larger meanders and eddies past a fixed point (Fig. 9). If this were pure wave motion, the dominant period and wavelength would produce phase speeds of 2–3 km day⁻¹ (as found by Kelly et al., 1998).

Cyclonic eddies and dipole eddy pairs form inshore of the jet’s cyclonic meanders, while anticyclonic eddies form offshore of the jet between the cyclonic meanders. At times, cyclonic eddies separate and move offshore of the jet, where they may form dipoles with the offshore anticyclonic eddies. These dipoles enhance westward transport of cool, enriched water.

The altimeter quantifies the offshore movement of energy in the jet and eddy system by showing the arrival of peak energy with wavelengths around 300 km. At 200, 400, 600 and 800 km offshore, maximum energy with this wavelength occurs in summer, fall, winter and spring, respectively, decreasing in magnitude as it moves offshore, until the increase in energy 800 km offshore is only a few per cent of the summer peak values 200 km offshore. If this offshore movement of energy were a result of wave propagation, it would indicate a group velocity of approximately 2.2 km day⁻¹. It is this observed westward movement of energy associated with the meandering jet and eddy system, as well as instances of enhanced westward transport, to which we refer.
when we say that the CCS “contributes” EKE and transport of coastal water mass properties to the deep ocean within 600–800 km of the coast.

In late summer and fall, poleward flow develops next to the coast at many locations despite continued equatorward winds. This poleward flow intensifies in the north in
late fall, in response to storms with poleward winds. As poleward winds extend farther south in winter, the circulation next to the coast returns to the large-scale poleward ICC, isolating the much weakened and fragmented meandering equatorward jet and associated eddies offshore and creating a “U-shaped” circulation pattern. The jet and eddies dissipate from fall to summer in the region between 500 and 800 km offshore.

4.2. Processes responsible for the observed structure in the CCS

The remaining discussion focuses on three specific aspects of the circulation related to the satellite observations: the initiation and westward movement of the seasonal jet; the processes that create the regions of high EKE; and the sources of freshwater in the core of the CC. In each case, questions are posed and then followed by hypothetical answers, based on the combination of altimeter/SST fields, previous observations and models. The first two questions are related to one another; the third is a long-standing issue in the CCS.

**Q1. What processes create the seasonal jet and control its offshore movement?**

**H1.** The seasonal jet begins as one or more local upwelling-front jets, which move off the shelf and unite to become a free, deep-ocean jet. This jet continues to move westward to approximately 130 W, where it weakens and dissipates between fall and spring. The dynamics governing the westward movement and dissipation have not been demonstrated convincingly, but the speed of the westward movement is consistent with Rossby wave dynamics, which have been proposed as the underlying mechanism.

The argument that the seasonal equatorward jet begins with the creation of local upwelling-front jets is based on the combination of observations from over the shelf with field and satellite observations from farther offshore. Over the shelf, observations from many locations show that an upwelling-front and jet develop quickly after the local onset of equatorward winds in spring. Off Oregon and Washington, the jet remains over or close to the shelf and slope (Huyer, 1983; Hickey, 1989). Off northern California it moves rapidly off the shelf (Strub et al., 1987b; Huyer, 1990) at approximately the same time that vigorous jets appear in surveys offshore of the continental slope (Kosro and Huyer, 1986; Huyer and Kosro, 1987). This is best illustrated by observations off northern California in 1987, where surveys offshore of the slope in April, May and June found a vigorous jet and front (Hayward and Mantyla, 1990; Kosro et al., 1991) that was not present in February and March. Those sampling over the shelf and slope (same latitude) in mid-March, however, found a strong jet along the slope, shortly after the onset of strong equatorward winds (Magnell et al., 1990). The combined observations can be interpreted as showing the jet leaving the shelf of northern California in mid-March and arriving offshore of the slope in April.

The mesoscale SST and altimeter height fields during 1987 and 1988 (Geosat) and 1993 (TOPEX/ERS-1) provide another view of the arrival of the jet approximately 50–100 km from the coast, with a larger alongshore field-of-view than provided by the ship surveys. During these three years, the latitudinal distribution of equatorward
currents in the springtime altimeter fields have the same latitudinal pattern as the onset of equatorward wind forcing. In early March 1987, equatorward winds extend from southern California to 48°N. Shortly thereafter (by the beginning of April), an equatorward jet appears “next to the coast” (50–100 km offshore) in the altimeter fields everywhere south of 47°N (Strub et al., 1998), similar to the local observations from surveys off northern California (Kosro et al., 1991). This contrasts to the situation in both 1988 and 1993, when winds become equatorward south of 40°N in March but remain poleward north of 40°N until May and June of both years. In TOPEX/ERS-1 fields from 1993 (Fig. 9) and Geosat fields from 1988 (not shown), an equatorward jet appears in the altimeter fields south of 40°N in April but does not appear north of 40°N until June.

Thus, the appearance of the jet in the deep ocean offshore of the shelf has a spatial and temporal distribution similar to the onset of local equatorward winds. By themselves, such winds do not normally create strong fronts and jets in the mid-latitude deep ocean, as they do next to the coast. This is the basis for hypothesizing that the front and jet found in the offshore SST and altimeter fields are most likely to have been created by wind forcing at the coast. The across-front structures of the frontal jets are consistent with this hypothesis. Field surveys show that horizontal density gradients extend from the surface to approximately 200 m in an upwelling front/jet observed to move from the shelf to offshore of the slope at 39°N in April–May 1982 (Strub et al., 1987b). These are similar to those observed farther from the slope in 1987 (Kosro et al., 1991) and 1988 (Huyer et al., 1991). Velocities associated with upwelling jets over the shelf off northern California range from 0.5–0.7 m s\(^{-1}\) at the surface to 0.1–0.2 m s\(^{-1}\) below 100 m (Huyer, 1990). In the offshore, deep-water jets, similar surface velocities (0.5–1.0 m s\(^{-1}\)) drop to 0.1–0.2 m s\(^{-1}\) between 100–200 m depth. The observation that the upwelling-jet moves quickly off the shelf off northern California but stays closer to the shelf off Oregon and Washington, is consistent with the patterns of SST and altimeter heights in 1993 (Figs. 9d and e) and 1987–1988 (Strub et al., 1991,1998), which show cold SST and/or altimeter height gradients closer to the coast north of 43°N and several hundred km offshore of the shelf off northern California. The boundary between these regions appears to be Cape Blanco (43°N), where recent studies have observed the separation of the coastal jet (Barth and Smith, 1998; Barth et al., 2000).

Once the front and jet appear west of the continental slope in the deep ocean, their westward movement between spring and fall is well established by the satellite fields and also by the hydrographic surveys in 1993 (Huyer et al., 1998). The two primary mechanisms that have been proposed to explain the westward movement are: (1) propagation as a first mode, annual baroclinic Rossby wave (White et al., 1990); and (2) direct surface forcing by the large-scale curl of the wind stress, which also may move westward on a seasonal time scale (Kelly et al., 1993).

Wavenumber-frequency spectra presented by White et al. (1990) from one year of Geosat altimeter data are similar to the expected dispersion relation of first mode, annual, linear Rossby waves. Unfortunately, aliasing of errors in the semidiurnal tidal model (caused by Geosat’s sampling pattern) creates fictitious westward propagating signals with properties similar to annual Rossby waves (Schlax and Chelton, 1994).
This does not mean that these Rossby waves are not present, it simply means that the Geosat sampling cannot detect them with statistical confidence. We note that the magnitudes of the approximately equal phase and group velocities estimated from the four-years of TOPEX observations in Section 4.1 are consistent with the proposed Rossby wave propagation, but this does not prove that these waves govern the dynamics of the offshore movement. As an alternative, Kelly et al. (1993) use a simple model to suggest that the apparent westward propagation in Geosat heights during 1987–1988 can be explained by the similar westward propagation of ECMWF wind stress curl during the same period. However, Kelly et al. (1998) repeat this analysis using three years of TOPEX data and wind stress curl from both ECMWF and the ERS-1 scatterometer and find that the wind fields do not show westward propagation similar to the altimeter height field. Thus, the dynamics responsible for the westward movement of the jet and eddy system remain to be determined with greater certainty.

The other first-order question related to the westward movement is why much of the EKE does not appear to continue to move past 130°W. We note that Kelly et al. (1998) find a similar drop in EKE west of 130–132°W in three years of surface drifter velocities, consistent with the altimeter data. One possible mechanism for trapping energy in the region next to the coast was suggested by McCreary et al. (1992), who used a layered model to suggest that non-linear effects can create mode-2 Rossby waves with eastward propagation, which maintain the surface and subsurface jets next to the coast. The model uses steady wind forcing and does not reproduce the seasonal aspects of the CCS as found in the altimeter data, leaving realistic tests of the mechanism for future work.

A simpler explanation for the low level of EKE west of 130°W is that the energy in the jet continues to propagate offshore but dissipates almost entirely in the year that it takes to do so. The very small amount of energy that the altimeter shows to arrive 800 km offshore in spring is consistent with this interpretation. This “simple” explanation leaves unanswered questions about the processes responsible for the dissipation.

Q2. What is the source of the seasonally developing EKE field in the CCS?

H2. Dynamical instabilities of the jet are the immediate source of EKE around the core of the seasonal jet; however, the ultimate source of energy in the jet is the formation of the density front, which may depend on proximity to the coast.

Numerical model studies have been used to investigate the source of the EKE in the CCS. These often try to identify the processes that determine the dominant wavelengths of perturbations or investigate the location of maximum values of EKE in relation to the core of the jet. Comparisons to altimeter data provide some guidance to these models studies. The mechanisms investigated include: coastal geometry (capes), bottom topography, wind forcing (wind stress and wind stress curl) and inherent dynamical instabilities.

Using a quasi-geostrophic (QG) model without bottom topography, Ikeda and Emery (1984) argue that the spacing of coastal capes (200–300 km along the California
coast) determines the wavelength of the meanders, linking the generation of the EKE back to wind forcing at the coast. This argument also has been made using primitive equation (PE) models, either including bottom topography (Haidvogel et al., 1991) or without it (Batteen, 1997). When capes are included, both models produce more vigorous meanders and filaments. Since Haidvogel et al.’s model includes bottom topography, the inference is that coastal geometry (capes) is more important than bottom topography (ridges). Batteen reports that the meanders are “anchored” by the locations of the capes, when present.

To examine the role of dynamic instabilities in a deep ocean jet that is not influenced by the coast, Pierce et al. (1991) and Allen et al. (1991) used QG models in idealized periodic channels without topography or coastlines, but with density and velocity structures similar to the summertime California Current jet. They found that dynamical instabilities within the jet (both barotropic and baroclinic in nature) create preferred wavelengths of approximately 260 km (the wavelengths of perturbations with maximum growth rates).

Pares-Sierra et al. (1993) attempt to differentiate between the effects of coastal processes and dynamic instabilities by comparing two models. A layered QG model represents both barotropic and baroclinic instabilities but does not include effects of coastal upwelling, since it excludes Ekman dynamics. A 1.5 layer PE model represents coastal upwelling but cannot represent baroclinic instability processes. As expected, the PE model produces higher variability near the coast and the QG model produces greater variability in the core of the CC. Because White et al. (1990) reported that the highest variability in one year of Geosat altimeter data occurs in regions “next to the coast” off California, Pares-Sierra et al. conclude that coastal upwelling is the dominant source of the EKE. However, the TOPEX data presented here and analyses of longer records of Geosat data (Kelly et al., 1993) make it clear that the region of highest variance (height and velocity) is around the core of the CCS. Lynn and Simpson (1987) also concluded that the highest variability of dynamic height in CalCOFI hydrographic data lies along the core of the CCS south of 37°N. Thus the model test of Pares-Sierra et al. actually favors dynamic instabilities in the CCS as the source of the EKE. This is further supported by direct current measurements over the slope off central California (Tisch and Ramp, 1997), where energy conversion estimates indicate that eddies are generated through baroclinic instabilities.

Combining Q1 and Q2. A hypothetical “explanation” of the seasonal development.

Above, based on the temporal and spatial correspondence of equatorward winds and the appearance of the equatorward currents in the offshore altimeter and field data, we hypothesize that nearshore upwelling (driven by wind stress at the coast) is an important process early in the spring-summer season in establishing local density fronts and jets, which move off the shelf. We also support the importance of instability processes in converting the mean potential and kinetic energies of the initial front and jet into the EKE of meanders and eddies surrounding the core of the jet. This is based on the similarity of the dominant wavelengths of 300 km revealed by the altimeter and
the preferred wavelengths for dynamical instabilities (260 km), along with altimetric evidence for the collocation of the maxima in EKE and the core of the jet. In addition, the altimeter observations show a monotonic decrease in EKE as the system moves west between summer and spring, which suggests that there are no significant further sources of potential or kinetic energy, once the jet leaves the coast.

We suggest the following hypothetical sequence of events, which are consistent with these observations. We stress that this is a hypothesis, not an established conclusion. Moreover, even if correct, our suggested sequence only “explains” what happens, not the dynamics of why it happens (why the jet leaves the shelf, etc.). (1) The primary source of energy to the jet is the potential energy set up during the initial creation and intensification of the density front, in proximity to the coast in spring. (2) This energy is converted to EKE (in the form of meanders and eddies) around the core of the jet by instability processes as the system moves offshore — which weakens the density front; (3) Unknown dissipation processes further diminish the EKE in the meanders and eddies; and (4) The instability and dissipation processes that weaken the density front and jet in the offshore, deep ocean are much greater than any processes that work to strengthen and maintain it, leaving little energy in the field as it continues to move past 130°W. The initial growth of instabilities may be enhanced by the similarity of the preferred wavelengths of dynamical instabilities and the distances between major capes (200–300 km), which set the locations and wavelengths of alongshore perturbations in the jet near the shelf.

Although the processes governing the westward movement of the system over the seasonal cycle have not been convincingly demonstrated, Rossby wave dynamics remain logical candidates. The observed speeds with which heights and energy tracked by the altimeter move offshore (2–3 km day$^{-1}$) are consistent with theoretical Rossby wave phase velocities (White et al., 1990). Seasonal movement of surface wind forcing may influence the apparent westward propagation (Kelly et al., 1993) but does not seem to be necessary to cause it (Kelly et al., 1998). A more thorough study is needed of linear and non-linear Rossby wave dynamics in the presence of typical eastern boundary topography and realistic seasonal wind forcing to establish the role of these dynamics with greater confidence. Dissipation processes and other details of the dynamics also need to be understood better.

Q3. On the basin scale, what are the sources of the fresh water in the core of the CCS and the interannual variability that is observed in the CCS transport?

H3. In addition to inflow from the North Pacific Current (the usual hypothesized source), other possible sources of fresh water in the core of the CCS include coastal inputs of the Pacific Northwest — the Columbia River, the Strait of Juan de Fuca, and the British Columbia coast.

In presenting the results, we noted that in both summer and winter, there is a direct connection between the CC and the coast of British Columbia. This raises a question that has existed for several decades, as to the source of fresh water in the core of the CC. A number of investigators have noted that the cool, fresh water in the core of the
CC off central and southern California can not come from either closer to the coast (cool but salty upwelled water) or farther offshore (warm and not as fresh) (Wickett, 1967; Chelton et al., 1982). Thus, it must come from upstream and these investigators have attributed it to the subarctic gyre.

Chelton et al. (1982) looked at interannual variability in transport and water mass characteristics and found that periods of higher equatorward transport in the CC are usually periods with lower salinity in the CC core. Chelton and Davis (1982) inferred (from tide gauge data around the NE Pacific Rim) that the strengths of the Alaska Coastal Current and the CC covary out of phase. Combining these results, Chelton and Davis hypothesized that the mechanism that directs more or less subarctic water into the CC is a change in position (or strength) of the North Pacific Current (the West Wind Drift). Similar changes in position and strength of the North Pacific Current have been hypothesized to be responsible for interannual to interdecadal changes in physical characteristics and species abundance/distributions (Hollowed and Wooster, 1992; Brodeur et al., 1996), but there has been no way to measure the hypothesized changes in the interior ocean.

The altimeter provides some insight into the pathways of transport between the two gyres. In Fig. 3, the shaded heights between 84–90 cm (the part of the North Pacific Current that enters the CCS, on average) show that water that enters from the northwest lies on the outside of the seasonal jet during much of spring-fall. This water has little impact on water properties on the inner half of the jet or the coastal region north of the SC Bight. During this period of the year, the most direct connection between the core of the CC off central California and regions with fresher water is to the US Pacific Northwest, where the Columbia and Fraser rivers create water characteristics similar to the core of the CC. During summer, there is also a direct connection to the boundary of the subarctic gyre along the coast of British Columbia, where fresher water is also found. Thus, fresh water from the upstream eastern boundaries of both gyres are possible sources for the anomalous low salinities measured off central and southern California, especially during spring and summer. However, flow into the SC Bight during most of the year creates a fairly direct connection between the SC Bight and the central north Pacific, via the inflow from the North Pacific Current (shaded regions in Fig. 3). During fall and winter, anomalous conditions delivered to the SC Bight from the North Pacific Current during the rest of the year could be advected northward and affect the rest of the system.

This provides at least three possible source regions for anomalously fresh water (1) transport from the central north Pacific in the North Pacific Current along the outer edge of the CC and into the SC Bight; and transport along the coastal boundaries from (2) the US Pacific Northwest and/or (3) the coast of British Columbia. Altimeter data give us the ability to monitor the detailed surface geostrophic circulation in the interior and along the boundaries of the NE Pacific Ocean upstream of the CCS. Continued field observations (to determine water properties and subsurface fields) in combination with longer altimeter records should be able to determine the sources of anomalous water in the CCS, if these observations can be maintained over a long enough period.
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