Modeling the generation of the Juan de Fuca Eddy

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Numerical simulations with the Regional Ocean Modeling System are used to study the generation of the cyclonic Juan de Fuca Eddy located off the entrance of Juan de Fuca Strait in summer. An initial simulation forced with average summer upwelling favorable winds, tides, and buoyancy boundary conditions that maintain an estuarine flow in Juan de Fuca Strait produces an eddy and currents that are in reasonable agreement with observations. Sensitivity studies are then carried out to explore the importance of these three forcing mechanisms. The relative proximity of dense water in the bottom estuarine flow entering the strait is shown to lead to enhanced upwelling off Cape Flattery when either wind or tidal forcing is applied. This upwelled water then mixes with the estuarine outflow and is advected offshore. The tidal upwelling arises through three mechanisms: M2 vertical excursions of nearly 20 m at 50 m depth west of the cape on the flood tide; strong tidally rectified vertical velocities west of the cape; and the spilling of denser bottom water over the western wall of Juan de Fuca Canyon on the ebb tide. The cyclonic eddy is a consequence of geostrophic adjustment to the doming isopycnals that arise from the upwelling. These model simulations refute an earlier hypothesis that the eddy is generated when California Undercurrent water is drawn up Tully Canyon and onto the Vancouver Island shelf, suggesting instead that these canyon dynamics play only a secondary role in maintaining the eddy once it is formed.


1. Introduction

The Juan de Fuca Eddy (also known as the Tully Eddy) is a cyclonic feature located off the entrance to Juan de Fuca Strait in summer (Figure 1). Although Freeland and Denman [1982] demonstrated that this upwelling eddy is composed of California Undercurrent (CUC) water and proposed a mechanism that could transport this water onto the Vancouver Island shelf via the Juan de Fuca and Tully (Spur) Canyons, they were vague on the precise eddy generation mechanism. Unlike the Sitka and Haida Eddies [Tabata, 1982; Crawford, 2002; Crawford et al., 2002; Di Lorenzo et al., 2005] that are formed regularly further north along the British Columbia and Alaskan Shelves and that detach and migrate into the Gulf of Alaska, the Juan de Fuca Eddy is a quasi-permanent feature that is always located in approximately the same location [Freeland and McIntosh, 1989], though its shape does vary slightly [MacFadyen et al., 2005]. This suggests a strong link to topography and/or persistent seasonal current features.

The upwelled water in the Juan de Fuca Eddy plays a significant role in making the southern Vancouver Island and northern Washington continental shelves one of the most productive fisheries regions (in terms of metric tons per unit area) in the northeast Pacific [Ware and Thomson, 2005]. Understanding the ecosystem dynamics that cause this region to be so productive is important both in the context of present fisheries management and with regard to the changes that are anticipated under global warming scenarios projected for the next century [Canadian Parks and Wilderness Society, 2005]. Given the role that the eddy appears to play as an initiation site for harmful algal blooms [Tully, 1942] and described in detail by Freeland and Denman [1982] (henceforth FD82), Denman and Freeland [1985], and Freeland and McIntosh [1989] (henceforth FM89). The hypothesis...
proposed by FD82 is that an eddy appears about the time of the “spring transition” when prevailing alongshore winds switch from southeasterly to northwesterly and drive a southeastward current over the outer continental shelf and slope. In the upper layer, the eddy is nearly geostrophic with a pressure gradient toward the eddy center balancing the outward Coriolis force. Further down the water column, this momentum balance is disrupted because the narrow Spur (Tully) Canyon blocks circulating flow and causes the velocities and the Coriolis term to approach zero. However, the inward pressure gradient remains and it drives an up-canyon flow that draws in water from the shelf edge. As noted by FM89, Allen [2000], and Allen et al. [2001], any southward flow that crosses a canyon will drive an up-canyon flow. Thus the presence of an eddy is strictly not necessary for this type of upwelling.

It should be noted that although the FD82 analysis demonstrated that the bottom water in Juan de Fuca Canyon originated with the CUC, this may not always be the case. In times of marked upwelling it probably is, though certainly any CUC water will mix with regional water masses as it moves up the canyon. In a seasonal water mass analysis recently performed for the Straits of Juan de Fuca and Georgia, Masson [2006] referred to this dense water as “deep inflow from the shelf”. In a similar analysis for the western continental margin of Vancouver Island, Mackas et al. [1987] found both CUC and “coastal deep” water to be present in Juan de Fuca Canyon. On the basis of these results, we will assume that the denser water that is upwelled to form the eddy is not solely composed of CUC water.

Although there have been previous numerical investigations of the Juan de Fuca Eddy, none yielded a conclu-
Weaver and Hsieh [1987] demonstrated that a buoyancy flux emanating from Juan de Fuca Strait might result in the formation of counter-rotating eddy pairs on the continental shelf. Klinck [1988] showed that a barotropic flow passing over a narrow canyon that cuts across the shelf will produce an eddy over the canyon. The simulations of Masson and Cummins [2000] generated a cyclonic eddy at the entrance of Juan de Fuca Strait when a northwest wind interacted with the estuarine flow emanating from the strait. However, they didn’t examine the eddy generation dynamics further as the Vancouver Island Coastal Current [VICC; Thomson et al., 1989] was the primary focus of their model investigations. Also, finally, though the average summer circulation simulated by Foreman et al. [2000b] included the Juan de Fuca Eddy, this model was spun-up from, and strongly nudged to, three-dimensional climatological temperature and salinity fields that reflected the presence of that eddy. So the model did not generate an eddy from first principles.

The objective of this study is to employ numerical experiments partially validated with observational data to better understand the flow dynamics near the entrance of Juan de Fuca Strait, with particular attention paid to the generation of the Juan de Fuca Eddy. The paper is organized as follows: Section 2 provides some details on the numerical model; section 3 compares observations with the results of a baseline simulation that generates the eddy with tidal, estuarine, and average summer upwelling favorable winds; section 4 describes sensitivity studies to examine the role of each of these three primary forcings; and section 5 describes the upwelling dynamics leading to eddy formation. Finally, the results are summarized and discussed in section 6.

2. Model and Forcing Details

The model employed in this study is the Regional Ocean Modeling System version 2.2 (ROMS, http://www.myroms.org/index.php) that has garnered wide usage in regional and coastal circulation studies [Chassignet et al., 2000; Haidvogel et al., 2000, 2008; She and Klinck, 2000; Marchesiello et al., 2001a; Di Lorenzo, 2003; Shchepetkin and McWilliams, 2005; Di Lorenzo et al., 2005; Warner et al., 2005; Wilkin, 2006]. The model domain chosen for our application is bounded by approximately 45.5°N to 50.0°N and 123.5°W to 128.5°W (Figures 1 and 3). There are 20 levels in the vertical with enhanced resolution in the surface and bottom boundary layers as given by the S-coordinate parameters $\theta_s = 0.8$ and $\theta_b = 5.0$ [Song and Haidvogel, 1994]. (As a test with 40 vertical levels and the same S-coordinate parameters produced essentially the same results as those soon to be described for the baseline case, this vertical resolution was deemed sufficient.) A stretched coordinate rectangular grid (Figure 3) with resolution as coarse as 5 km adjacent to the western boundary and as fine as 1 km near Cape Flattery and the entrance to Juan de Fuca Strait was employed to obtain an accurate representation of topographic and coastal features in the region of interest without the computational demands that would be needed to maintain the same resolution everywhere. The model bathymetry is a smoothed interpolation of data from a variety of sources including: ETOP02 and the NOAA database for deep values offshore black regions are clouds.

Figure 2. MERIS satellite fluorescence for 6 June 2003 (courtesy of the European Space Agency and provided by J. Gower and S. King). Offshore black regions are clouds.

Figure 3. The horizontal model grid and 200 m depth contour. Rectangles represent 8 grid cells in each direction.
Average summer upwelling winds computed from the MM5 atmospheric model. Vectors denote direction and the underlying colors denote speed (m s\(^{-1}\)).

Figure 4. Average summer upwelling winds computed from the MM5 atmospheric model. Vectors denote direction and the underlying colors denote speed (m s\(^{-1}\)).

3. Initial Summer Simulations

Initial conditions for the model were computed from a combination of the Levitus et al. [1994] temperature and salinity monthly climatologies and a similar database with much higher spatial resolution that has been constructed at the Institute of Ocean Sciences using all available historical data for the Alaska, Washington, Oregon, and British Columbia shelves (http://www.pac.dfo-mpo.gc.ca/sci/osap/data/alaska/default_e.htm). Though the data archive is more extensive and the smoothing techniques were slightly different than those described by Foreman et al. [2006], the resultant average summer sigma-t contours at 50 m depth are very similar to those shown in Figure 4a of that publication. It should be noted that although the model initial conditions include temperature and salinity signatures of the Columbia shelves, all rivers are not explicitly included in the model forcing.

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[11] Wind-forcing was computed using output from the University of Washington mesoscale atmospheric model (MM5) with 4-km resolution (http://www.atmos.washington.edu/mm5rt/). Tinis et al. [2006] recently compared 2003 and 2004 wind and fall winds computed by this model, and a similar larger domain 12-km version, with observations from twelve offshore meteorological buoys ranging from British Columbia to northern California. Mean summer winds were generally in good agreement with the 12-km model speeds, ranging from 81 to 101% of their observed counterparts. Corresponding directions differed, on average, by only 8°. Using archived MM5 data, an average upwelling wind stress field was computed by averaging 2003 – 2005 summer (July – September) stresses associated with offshore high pressure atmospheric systems. This restriction eliminated northeasterly winds associated with infrequent summer storms. These stresses were first computed at each MM5 grid point and then interpolated bi-linearly to the ROMS grids. Winds associated with these average stresses (Figure 4) have speeds of over 8.0 m/s off Vancouver Island and southern Oregon, and increasing magnitudes eastward in Juan de Fuca Strait. Offshore directions are generally parallel to the coast and show strong topographic steering in Juan de Fuca Strait. Though a heat flux consistent with these average upwelling favorable winds could have also been computed from MM5 output, this was not done. Rather, in all simulations the surface ocean temperature was nudged back to initial values with an e-folding timescale of 5 days.

Tidal forcing at the model boundaries was interpolated from the Foreman et al. [2000] northeast Pacific Ocean model that assimilated tidal harmonics computed from TOPEX/Poseidon altimetry [Cherniauskaya et al. 2001]. Only the four main constituents, \(M_2\), \(S_2\), \(K_1\), and \(O_1\) were used as they account for approximately 70% of the elevation range at the entrance to Juan de Fuca Strait. The relatively small latitudinal extent of the model domain means that tidal potential forcing will be small and can be neglected [Foreman et al., 1993].

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forested and remote wind-forcing in the estuarine circulation, in order to determine if they can capture the predominant summer flow patterns near the entrance to Juan de Fuca Strait.

15 The imposition of local wind and tidal forcing is straightforward. The former is applied as a surface stress at each horizontal grid cell while the latter is applied only at the boundary cells. As mentioned earlier, estuarine flows are imposed by initializing with 3D climatological temperature and salinity fields and strongly nudging to these fields adjacent to the Juan de Fuca boundary. The implication is that, in the absence of poleward coastal winds, these fields include the appropriate baroclinic pressure gradients to drive a “normal” estuarine flow. Initial conditions were the climatological temperature and salinity fields described earlier. The model was run for sixty days and a harmonic analysis [Pawlowicz et al., 2002] of stored hourly values was performed over the last fifteen days to extract the largest components of tidal and subtidal energy. As an average over this same time period produces virtually the same result as the constant (zero frequency) term, $Z_0$, arising from a harmonic analysis, the fields arising from these two calculations will be used synonymously. The model will be validated by comparing these harmonics or averages with analogous values arising from the analysis of observations.

16 To illustrate the effectiveness of the strong boundary nudging in the eastern Juan de Fuca Strait in accurately generating the estuarine circulation, we compare average $Z_0$ (zero frequency) model and observed flows along transect A (Figure 1) which crosses the entrance to Juan de Fuca Strait (Figure 5). All values were converted to their along-channel components (using $-27^\circ$ clockwise from east, 117°T, as the along-strait direction) with positive values denoting flow into the strait. All observed values shown in black, except the one at 225 m depth, were computed from analyses of moored current meter time series taken between June and August of 1984 [Hickey et al., 1991]. Values at 225 m at the central mooring site were from observations taken in June 1985 [Dewey and Crawford, 1988]. The model shows a surface outflow that is stronger toward the middle of the strait than is suggested by these and other observations further to the east [Labrecque et al., 1994]. The model inflow is weaker than the observed values in mid-channel and the model inflow/outflow interface is shallower on the southern side and has too much tilt across the strait. However, as the model values are forced with average summer climatology in eastern Juan de Fuca Strait, rather than conditions specific to the time period of the observations, it is not reasonable to expect perfect agreement. In fact, as demonstrated in Table 1 of Foreman et al. [2000], the 1984 subtidal flows had considerable variability about their mean values. (More recent observations mid-way along the strait also show the core position of the estuarine flow to slowly wander cross-channel [Thomson et al., 2007].) As suggested by the 1985 inflow value at 225 m relative to the 175 m value from 1984, there is also considerable interannual variability. Further evidence of this variability comes from analyses of observations taken by a downward-looking University of Washington (UW) RDI Workhorse ADCP mounted 6.5 m below a toroidal buoy deployed, as part of the ECOHABPNW project, at essentially the same location as the 1984 mid-channel site. These ADCP data were collected at hourly intervals and harmonically analyzed for the period of June to August 2003.

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Figure 5. Contours of the alongstrait ($-27^\circ$ clockwise from east) $Z_0$ (zero frequency) currents (cm s$^{-1}$) along transect A (Figure 1) for the baseline simulation. Positive values denote approximately eastward. Except for the value at 225 m depth [Dewey and Crawford, 1988], all black values at the dots are from analyses of current meter time series between June and August 1984 [Hickey et al., 1991]. White values to the left of the dots are from analyses of observations from a downward ADCP between June and August 2003.
Results, shown in white on Figure 5, indicate a deeper surface outflow in 2003 than in 1984, and this is in better agreement with the model estimates.

It should be noted that the strong nudging to climatological temperature and salinity values along the Juan de Fuca model boundary are essential to accurately reproducing both the estuarine flows in the strait, and the eddy. When this initial summer simulation was repeated with that nudging timescale reduced from 1 day to 30 days, much weaker estuarine flows and a weaker eddy resulted.

Figure 6a shows the average horizontal $Z_0$ currents at 35 m depth superimposed on average salinity field for the same fifteen-day period. Observed $Z_0$ currents computed from analyses of current meter time series taken at 30 m between June and August 1984 [Hickey et al., 1991] and at 40 m between March and June 1985 (FM89) are also shown to demonstrate that the model has captured typical summer subtidal circulation features for the region [Thomson et al., 1989]. (These two observational data sets were chosen because together they provide the best spatial views of mean currents off the entrance of Juan de Fuca Strait.) Although there is a relatively large variability associated with summer currents in this region [Foreman et al., 2000a, Table 1], the model captures the following features with reasonable accuracy: an estuarine outflow from the strait, an eddy, a VICC, and a southeastward Shelf Break Current that, in this case, is driven by the local winds.

Figures 6b and 6c show the analogous $Z_0$ model currents for the surface and at 100 m depth. The presence of wind-driven currents means that the surface fields show a much weaker eddy than at 35 m depth. Superimposed on the Figure 6b model results are average observed currents for the period of 1 June to 31 August 2003 as measured by S4 meters moored at 3 m depth as part of the ECOHABPNW project. Not only are the model currents in excellent agreement with these observed averages, but also the pattern of fresher water emanating from the strait and curling around the eddy in a counterclockwise manner is similar to that shown in Figure 2. Figure 6c also has superimposed average observed currents, in this case from the 1984 and 1985 moorings at 100 m as described by Hickey et al. [1991] and FM89. Much poorer agreement is now seen between the model and observed values, especially in the eddy region where the observations do not display much correlation with each other. Nevertheless, model fields at 100 m do show a weak eddy and saltier water whose location is consistent with the eddy seen at 35 m.

Table 1 by Foreman et al. [2000b] gives standard deviations for many of the current meter (as opposed to ADCP) $Z_0$ values presented here. Though standard deviations could also be computed for the model $Z_0$ values, any comparison with their observed counterparts would be a case of “apples versus oranges” because the observations experienced variable winds and estuarine flows while the model was forced with temporally constant winds and estuarine flows. Thus the model standard deviations would be much smaller.

Though a quantitative comparison with tide gauge harmonics (analogous to that in FT97) was not performed, contours of tidal elevation amplitude and phase for $M_2$ and $K_1$ were similar to those obtained in that paper. Figure 7 compares model and observed major semi-axes for the $M_2$...
and $K_1$ tidal ellipses along transect A (Figure 1) crossing the entrance of Juan de Fuca Strait. As with the $Z_0$ currents shown in Figure 5, all values shown in black except those at 225 m depth were computed by harmonically analyzing observations taken between June and August of 1984 [Hickey et al., 1991] while those at 225 m at the central mooring are from June 1985 [Dewey and Crawford, 1988].

**Figure 7.** Contours of the (a) $M_2$ and (b) $K_1$ major semi-axes (cm s$^{-1}$) along transect A (Figure 1) at the entrance of Juan de Fuca Strait for the baseline simulation. Values at the black dots at 175 m depth and above are from analyses of current meter time series taken in the summer of 1984 [Hickey et al., 1991] while those at 225 m at the central mooring are from June 1985 [Dewey and Crawford, 1988].

O$_1$ major semi-axes show similar patterns with their magnitudes being approximately 25% and 55% of $M_2$ and $K_1$, respectively.

[22] Plots of the model $M_2$ and $K_1$ major semi-axes at 35 m depth are shown in Figure 8 along with values computed from archived current meter time series moored at either 30 or 40 m depth. The $K_1$ model values are similar to those calculated by the FT97 barotropic finite element model and in particular, show evidence of the diurnal shelf waves along the western Vancouver Island shelf that have been extensively studied [Crawford and Thomson, 1984] and modeled [Flather, 1988; Cummins et al., 2000]. The present model values are in reasonable agreement with the observed major semi-axes and those from previous models. However, the $M_2$ speeds attain much larger values westward of the strait entrance than those in both the FT97 results and an analogous plot (not shown) arising from a barotropic (homogeneous density) run with the present model. A plot similar to Figure 8a but for the surface $M_2$ major semi-axes shows that the region of enhanced current speeds forms a narrow “tongue” extending offshore from Cape Flattery parallel with the southern shoreline of the strait. The fact

**Figure 8.** Contours of the model (a) $M_2$ and (b) $K_1$ major semi-axes (cm s$^{-1}$) at 35 m depth for the baseline simulation. Values in black are from analyses of current meter time series taken in the summer of 1984 [Hickey et al., 1991] at 30 m depth and those in white are from the spring of 1985 (FM89) at 40 m depth.
that the 35 m model major semi-axes values are also in reasonable agreement with the observations suggests that the larger values extending beyond the strait entrance are real features that must arise from the stratification.

4. Sensitivity Simulations

[23] The simulation in the previous section demonstrates that the Juan de Fuca Eddy can be generated with average summer upwelling favorable winds, an average summer estuarine flow in the strait, and the four major tidal constituents. Comparisons with analyzed current meter and ADCP observations near the entrance to the strait demonstrated that the model currents are reasonably accurate, thereby giving credibility to the model dynamics and suggesting that the dynamics generating the eddy in the model are real. In this section, we carry-out sensitivity runs to investigate the role that each of these forcing mechanisms plays in the eddy formation. A summary of these runs is given in Table 1.

[24] The first sensitivity run (B in Table 1) was carried out to determine the roles of nudging along the northern, southern, and western boundaries of the model domain. For this simulation, both tidal and wind-forcing were turned off so that the only forcing was the strong nudging to climatological temperatures and salinities adjacent to the Juan de Fuca model boundary (a 1-day e-folding timescale) and the relatively weak nudging (a 30-day e-folding timescale) along the northern, southern, and western boundaries. The salinities and velocities at 35 m depth averaged over days 46–60 are shown in Figure 9. (The velocity and salinity scaling are the same as in Figure 6 to permit direct comparisons.) Clearly visible are a strong estuarine flow in the strait, a northward VICC along the Vancouver Island coast, a northward coastal current along the Washington shelf resulting from the initial pool of fresh warm water off the mouth of the Columbia River, and an offshore current flowing to the north and northwest. Although the southwestern edge of the VICC developed a meander and filament by day 15 of this run and part of this filament subsequently broke apart to form both cyclonic and anticyclonic eddies, these features either gradually decayed or migrated offshore. Though Figure 9 shows a hint of a cyclonic eddy off entrance to Juan de Fuca Strait, it is much weaker than that in Figure 6 and is probably a just a remnant from the initial conditions. (The lack of wind and tidal forcing leads to less mixing and a slower elimination of the initial conditions than in subsequent runs.) The offshore currents flow toward the northwest, the opposite direction of the Shelf Break Current that FD82 suggested might play a role in eddy formation and that, in the absence of wind-forcing, might be expected to develop as a result of the

![Figure 9](image)

Figure 9. Salinity and currents at 35 m depth averaged over days 46–60 of the sixty-day simulation with only boundary nudging (run B in Table 1). To facilitate comparison, the color bar and vector scaling are the same as in Figure 6a.

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A second sensitivity run (C in Table 1) was conducted with tidal and upwelling wind-forcing but no estuarine flow. In this case, identical initial temperature and salinity profiles, typical of summer conditions off the Washington coast, were specified over the entire model domain. The wind and tidal forcing again produced upwelling along the Washington coast (Figure 10) and since the absence of an estuarine flow also removed the VICC, there is now upwelling over the inner portion of the Vancouver Island shelf. An eddy is apparent off Cape Flattery but it extends further to the south, its center is less saline relative to the values off Washington (and Vancouver Island), and its cyclonic velocities are generally weaker than in Figure 6a. The results of this sensitivity run suggest that the estuarine flow in the strait plays an important role in the eddy formation seen in the baseline run (A in Table 1). In the presence of upwelling favorable winds the lighter near-surface outflow curls around the upwelled water restricting its southern extent, while the proximity of dense water in the bottom estuarine inflow seems to produce stronger
upwelling off Cape Flattery. This second implication is now examined with the next run.

The third sensitivity simulation (D in Table 1) removed tidal forcing from the baseline run, but retained the upwelling favorable winds and estuarine flows in the strait. In this case, an animation (view via link from http://www-sci.pac.dfo-mpo.gc.ca/osap/people/foreman_e.htm) of the surface salinity and velocity fields clearly shows the development of upwelling along the Washington coast and an eddy that spins up off Cape Flattery. Figure 11 shows average 35 m salinities and velocities for the same fifteen-day period as in Figures 6a and 9. A cyclonic eddy that is even stronger and more clearly defined than the one arising from the baseline simulation is evident. In order to illustrate eddy development and to demonstrate that the signature of an eddy in the initial temperature and salinity conditions has not pre-disposed all simulations to develop an eddy, Figure 12 shows isohalines along transect B at the end of days 1, 5, 10, 30, and 60 of this run. The first panel shows a subtle doming of the isohalines in the initial conditions. However, by day 5, a much more prominent doming has developed to the west of Cape Flattery. The simulation continues, the dome continues to grow and moves further westward so that by day 10, this feature has grown and spread across the full width of Juan de Fuca Canyon. As the simulation continues, the dome moves to the west side of Juan de Fuca Canyon and by day 60, that center is over Tully Canyon. This latter position may explain the rationale behind the FD82 and FM89 hypothesis that the upwelling originates in this second canyon. However, the present simulations suggest that this is not the case; rather the eddy develops off Cape Flattery and under steady upwelling favorable winds grows and moves westward to lie over the Tully Canyon. It should be mentioned that although this simulation reaches a relatively steady state by day 60, the eddy and doming seen in Figure 12e do change their shape and position slightly over time. Run D therefore demonstrates that although the presence of steady upwelling favorable winds and an estuarine flow in the strait will produce a cyclonic eddy that is centered over Tully Canyon, the eddy does not originate there. Rather, it is generated by strong upwelling off Cape Flattery and moves to this location as it grows and detaches from the coast.

A fourth (E in Table 1) sensitivity run was conducted to reinforce the conclusion from the previous sensitivity analysis that our initial conditions are not pre-disposing the simulations to form an eddy. In this case, initial temperature and salinity fields in the region approximately bounded by 48.0°N to 48.7°N, and 124.5°W to 126°W, were replaced with values that were interpolated from climatological values outside this rectangle. This effectively eliminated any evidence of the eddy. The same forcing as in the baseline study (i.e., tides, upwelling favorable winds, and strong nudging to the temperature and salinity values adjacent to the eastern Juan de Fuca boundary) was then applied and the results compared with that run. Apart from the fact that the eddy took a little longer to develop, the results were very similar. In fact, by the end of the sixty-day simulation animations (not shown) the surface salinity fields were almost identical. Both simulations showed the same enhanced upwelling and tidal pumping off Cape Flattery and a growing bulge of upwelled water that spread westward and partially detached to form a cyclonic eddy. On the basis of this sensitivity run and the sequence of snapshots in Figure 12, we therefore conclude that the initial conditions are not affecting eddy formation.

The fifth sensitivity simulation (F in Table 1) was forced with a steady estuarine flow and tides but no wind. As might be expected, the tidal harmonics arising from harmonic analyses of time series from this run were virtually identical to those from the baseline case described in section 3. Interestingly, the average salinities and velocities shown in Figure 13 (whose scaling is again identical to that in Figures 6a, 9, 10, and 11) illustrate that the tides can also generate sufficient upwelling to produce an eddy, albeit

**Figure 10.** Salinity and currents at 35 m depth averaged over days 46–60 of the sixty-day simulation with upwelling winds and tides but no estuarine flow in Juan de Fuca Strait (run C in Table 1). Color bar and vector scaling are the same as in Figure 6a.

**Figure 11.** Salinity and currents at 35 m depth averaged over days 46–60 of the sixty-day simulation with upwelling winds and an estuarine flow but no tides (run D in Table 1). Color bar and vector scaling are the same as in Figure 6a.
weaker than what is produced with only wind-forcing (Figure 11). Although the absence of upwelling favorable winds has greatly reduced the saltier water along the southern Washington coast, there is still some present due to remnants of the initial conditions, local tidal mixing, and the southward propagation of upwelled water from Juan de Fuca Strait and Cape Flattery. Results from a run analogous to C (Table 1) but with no winds showed saltier water along the Washington coast moving southward after being produced by tidal upwelling and mixing off Cape Flattery and in the strait. The fact that this water is denser than that further offshore causes a setdown in the coastal elevations, geostrophic flow to the south, and continued southward propagation of the upwelled water. From this fifth sensitivity run we conclude that upwelling favorable winds are not necessary for the generation of the Juan de Fuca Eddy and the presence of upwelled water along the Washington coast. Tides and an estuarine flow alone are sufficient, although the eddy is intensified by the upwelling favorable winds.

[20] The sixth and final sensitivity simulation (G in Table 1) was similar to the baseline case except that average summer upwelling favorable winds were replaced by average winter downwelling winds computed from the MM5 4 km model. (Note that this was not a complete winter simulation as a climatology of winter temperature and salinity fields has yet to be constructed from a sparser set of observations and thus could not be used as initial conditions. Instead, this run was initiated with the same summer climatology as all the other runs.) A plot (Figure 14) of average salinity and velocity at 35 m depth shows a weaker, but nonetheless still present, eddy than for runs A, D, and F. It therefore seems that downwelling winds (at least those employed here) are not sufficient to suppress tidal upwelling and eddy formation.
There are two caveats that need to be pointed out for the final sensitivity run. The first is that despite the downwelling winds, the offshore currents in Figure 14 are seen to be southward and southeastward. This arises because the combination of these winds with tidal forcing produces spurious enhanced upwelling at the eastern corner of the southern boundary. This upwelling depresses the elevation there and sets up a north-to-south downward pressure gradient that forces a flow to the south. Several parameter adjustments associated with the radiation/nudging conditions along the southern boundary were unsuccessful in removing this effect. Though it will be important to resolve this problem for future realistic winter simulations, we do not feel that the incorrect offshore flows are affecting the local upwelling off Flattery. So the problem was left unsolved.

The second caveat is that the forcing did not include a sustained Columbia River discharge or pool of low density water that would arise from the discharge of lesser rivers along the Washington coast. A pool of fresh water was present off the mouth of the Columbia in the initial salinity field and it dispersed and moved northward along the Washington coast as the simulation progressed. (For the upwelling wind cases, the plume moved southward with the prevailing wind.) Though the average winter winds computed from the MM5 model were not sufficiently strong to transport this fresh pool from the vicinity of the Columbia River to Cape Flattery, in an analogous simulation with stronger winds, the plume did enter Juan de Fuca Strait, consistent with the observational results of Thomson et al. [2007]. The fifteen-day average salinity field at 35 m depth for these stronger downwelling winds showed virtually no upwelling off the cape. We, therefore, hypothesize that with an average Columbia River discharge and average winter winds, the plume will reach the strait [Thomson, 1981; Thomson et al., 2007] and provide sufficient near surface stratification to essentially cap the tidal upwelling off Cape Flattery. Though this hypothesis requires further testing, we speculate that a northward flowing Columbia plume is the primary reason that the Juan de Fuca Eddy has not been observed in winter.

### 5. Eddy Generation Dynamics

The preceding numerical simulations have demonstrated that the Juan de Fuca Eddy can be generated by: i) average summer upwelling favorable winds and an estuarine flow; ii) tides and an estuarine flow; and iii) all three forcing mechanisms working together. Even winter downwelling favorable winds, when combined with tides and an estuarine flow (but no Columbia River discharge), produced a weak cyclonic eddy. In all instances, the simulations suggest enhanced (stronger) upwelling off Cape Flattery; the nature of which will be examined now.

Figure 15 displays the $Z_0$ vertical velocities along transect B for the baseline run with wind, tidal, and estuarine forcing (run A in Table 1); and the sensitivity runs with estuarine flows and winds but no tides (run D); and estuarine flows and tides but no winds (run F). With wind, tidal, and estuarine forcing (Figure 15a), average upwelling speeds are seen to range from about 0.1 cm/s at 150 m depth to 0.03 cm/s at 75 m depth on the eastern side of the canyon. Downwelling regions with speeds as large as 0.1 cm/s are seen over the rises north of Cape Flattery and separating Juan de Fuca and Tully canyons. Alternating bands of upwelling and downwelling seen across Juan de Fuca Strait suggest the presence of vertical recirculation cells that may be partially caused by Herlinveaux Bank (Figure 1), just to the south of this transect. (A plot (not shown) similar to Figure 15a but crossing Herlinveaux Bank along transect C (Figure 1) displays strong vertical recirculation cells on both the eastern and western flanks of the bank.) However, the average vertical velocity within the Juan de Fuca Canyon region of Figure 15a was computed to be 0.0067 cm s$^{-1}$, so upwelling predominates. Also, the strongest upwelling is seen to be along the eastern flank of the canyon, just to the west of Cape Flattery.

Figure 14. Salinity and currents at 35 m depth averaged over days 46–60 of the sixty-day simulation with downwelling winds, tides, and an estuarine flow (run G in Table 1). Color bar and vector scaling are the same as in Figure 6a.
The analogous plot in the absence of tides (Figure 15b) generally shows much weaker vertical velocities. In particular, upwelling on the east side of the canyon is weaker, there is virtually no upwelling on the west side of the canyon, and there is now a larger region of upwelling in the western part of Juan de Fuca Strait. Despite these smaller speeds, the upwelling to west of the cape must still be sufficiently large to produce the isohaline doming seen in Figure 12 and to generate the eddy seen in Figure 11.

The patterns of upwelling and downwelling for the case with tides but no winds (Figure 15c) are very similar to those for the case with both forcing mechanisms (Figure 15a). This suggests that the vertical velocities seen in Figure 15a are largely of tidal origin; i.e., due to tidal rectification. (This, in itself, is an interesting result as normally tidal residuals are presented for only the horizontal velocities. Here we see that they can also be significant in the vertical.) However the values in Figure 15c are generally larger than those in Figure 15a indicating that the wind must be reducing this rectification.

At first impression, the magnitude of the mean vertical velocities seen in these three panels might seem to be at odds with the corresponding size and salinity of the eddies seen at 35 m depth in Figures 6a, 11, and 13. Specifically, the strongest eddy (Figure 11) arises from the weakest upwelling west off Cape Flattery (Figure 15b). As seen from the average vertical viscosity coefficients shown for runs A and D in Figure 16, the reason is that tides not only provide additional upwelling mechanisms through both rectification and vertical displacements, but they also...

Figure 15. \( Z_0 \) vertical velocities (cm s\(^{-1}\)) along transect B (Figure 1) arising from harmonic analyses over days 46–60 for runs (a) A, (b) D and (c) F.

Figure 16. Vertical viscosities (m\(^3\) s\(^{-1}\)) averaged over days 46–60 along transect B (Figure 1) for the sixty-day simulations with (a) upwelling winds, tides, and an estuarine flow (run A in Table 1), and (b) upwelling winds and an estuarine flow (run D in Table 1). Note the scaling difference in the color bars.
provide a major source of mixing. Figure 16 indicates that in regions such as the bottom of Juan de Fuca Strait and along the eastern side of Juan de Fuca Canyon, tides have increased the vertical viscosity by approximately an order of magnitude. Thus the primary reason that the eddy seen in Figure 11 is more saline and perfectly formed than the one in Figure 6a is that it was not exposed to any dilution or dispersion from tidal mixing.

[37] As seen in Figure 12, the result of the enhanced wind-driven upwelling off Cape Flattery is a dome of more saline water that grows westward and detaches to form an eddy when it reaches a sufficiently large diameter [Di Lorenzo et al., 2005; Crawford et al., 2002; Cenedese and Whitehead, 2000]. The eddy’s cyclonic sense of rotation is determined primarily by the geostrophic balance which develops when a pool of denser water rises in the water column. However, the conservation of potential vorticity as the upwelled water spreads from being over the narrow shelf at Cape Flattery to over the canyon, and the south-easterly currents flowing along the continental slope, the so-called Shelf Break Current [Foreman et al., 2000b], may also play lesser roles. The likely reason that the upwelling is enhanced off the cape is the proximity of readily available cold saline water that moves up Juan de Fuca Canyon and into the strait as part of the bottom estuarine flow. The western side of Tatoosh Island off Cape Flattery (Figure 1) is only about 4 km to the east of the 200 m depth contour that delineates the edge of the canyon. Consequently, there is a shorter, more direct path for upwelling water off the cape than exists further to the south along the Washington coast.

[38] Figure 17 shows the vertical displacement amplitudes for both the \( M_2 \) and \( K_1 \) tidal constituents along transect B. Following the Cummins and Oey [1997] definition, these displacements are simply computed as the vertical velocity amplitudes divided by the constituent frequency. (Hence vertical velocity amplitudes comparable with those for \( Z_0 \) can be easily computed from the values in Figure 17. For example, a 20 m displacement for \( M_2 \) is equivalent to a vertical velocity amplitude of approximately 0.3 cm/s.) At 100 m depth, the \( M_2 \) vertical displacements are approximately 25 m to both the east and west of Cape Flattery while at 50 m depth, the displacements to the west of the cape are about 20 m. These tidal displacements provide another mechanism for transporting denser water in the canyon and strait closer to the surface. \( M_2 \) vertical displacements along the western side of the canyon are about 20 m at 150 m depth but as they are generally smaller than those to the east, they would be less effective in upwelling denser water.

[39] The \( K_1 \) displacements seen in Figure 17b are equally interesting. In this case, maximum displacements are 35 m at 100 m depth on the eastern side of the cape and 20 m at 50 m depth on the western side. However, displacements are much larger on the western side of the canyon, reaching 60 m at 175 m depth and 40 m at 125 m depth. Although these displacements will not be as effective for upwelling nutrient-rich water from the bottom estuarine flow to the surface as the \( M_2 \) vertical excursions off Cape Flattery, they do suggest intriguing links to the generation of baroclinic diurnal shelf waves off Vancouver Island. Note that both constituents also show vertical displacements along the continental slope. For \( M_2 \), these are associated with internal tides [Drakopoulos and Marsden, 1993] while for \( K_1 \) they are associated with continental shelf waves [Crawford and Thomson, 1984].

[40] As Figures 16 and 17 suggest several potential regions for wind-driven and tidal upwelling, it is important to ascertain which are the primary ones. Figure 12 demonstrates that for wind-forcing, the primary upwelling center is to the west of Cape Flattery. Figure 18 suggests that this is also the primary region for tidal upwelling. Two snapshots of the surface salinities and currents separated by seven hours on day fifty-one (during spring tides) of the baseline (run A) simulation are shown. The first panel corresponds to a flood tide off Cape Flattery while the second is for the preceding ebb. Both panels show evidence of strong upwelling along the Washington coast and to a lesser extent off southwestern Vancouver Island. There is also the suggestion of more saline water streaming off Cape Flattery, south of the fresh estuarine outflow. On the flood tide, there is a larger patch of upwelled water to the west of the cape while on the ebb it is reduced (at least on the surface). The direction of the velocities off the cape in the first panel

Figure 17. Vertical displacement amplitudes (m) along transect B (Figure 1) for (a) \( M_2 \) and (b) \( K_1 \).
suggests that this upwelling arises as the flooding tide encounters shallower bathymetry (Figure 1). This would certainly be consistent with the large $M_2$ vertical displacements seen in Figure 17a; in particular, an upward transport on the flood tide.

[41] It should be noted that the tidal upwelling seen in these model runs may be similar to the topographic upwelling that was simulated by Tee et al. [1993] off Cape Sable, Nova Scotia. In that case, a combination of upwelling induced by tidal rectification and strong tidal mixing was shown to reproduce an observed cold water anomaly. Garrett and Loucks [1976] demonstrated that the centrifugal force arising from alternating tidal currents with maximum strength of approximately 1 ms$^{-1}$ was the likely cause of this Cape Sable upwelling. It may also be the primary mechanism producing the larger (tidally rectified) vertical velocities seen in Figures 16a and 16c than those in Figure 16b. In this case, the radius of curvature associated with Cape Flattery, being much smaller than that for Cape Sable, produces a centrifugal upwelling that is much larger than the opposing Coriolis downwelling associated with the bottom estuarine flow. So the effect should be larger here. It should be noted that this same centrifugal upwelling may also be relevant to the tidally rectified eddies studied by Thomson and Wilson [1987] off Cape St. James, British Columbia.

[42] In addition to the flood tide upwelling off Cape Flattery, an animation (view via link from http://www-sci.pac.dfo-mpo.gc.ca/osap/people/foreman_e.htm) of salinity along transect B indicates that the ebb current flowing out of Juan de Fuca Strait may in some cases (such as during spring tides) be strong enough to overcome the eastward estuarine bottom flow and push the denser water lying along the bottom of the canyon and strait up onto the shelf adjacent to the western side of the canyon (Figure 1). This is illustrated in Figure 19 with snapshots from the animation at the same simulation times as those in Figure 18. The first panel shows that on the flood tide, water with salinity 33 psu almost reaches the surface just to the west of

![Figure 18. Snapshots of surface salinity and currents off Cape Flattery on day 51 of the sixty-day simulation with tides, upwelling winds and estuarine flow (run A): (a) hour 23 at flood tide, (b) hour 16 at ebb tide. Circled regions highlight the greatest difference in salinity.](image1)

![Figure 19. Snapshots of salinity along transect B on day 51 of the sixty-day simulation with tides, upwelling winds and estuarine flow (run A): (a) hour 23 at flood tide, (b) hour 16 at ebb tide. The circled region in Figure 19a highlights water with salinity 33 psu almost reaching the surface while in Figure 19b, it highlights water with salinity 33.7 spilling over the rise between Juan de Fuca and Tully Canyons.](image2)
the cape. As the tide reverses to ebb, all the isohalines on the eastern side of the canyon fall and set off a wavelike propagation westward that culminates in steep salinity (density) gradients along the western side of the canyon (Figure 19b), with water of 33.7 psu spilling onto that portion of the Vancouver Island shelf separating the Juan de Fuca and Tully Canyons. Throughout this process, a doming of the isohalines, characteristic of the eddy, remains evident over the Tully Canyon and Vancouver Island shelf.

[43] The wavelike pattern propagating westward across the bottom of the canyon may be internal tides that are generated in the vicinity of Cape Flattery on each ebb tide. The animation associated with Figure 19 reveals that these waves attenuate relatively quickly along the mid-VI shelf (approximate depth of 125 m), probably as a result of the mixing that ensues when this denser water is pushed onto the shelf. As the tides reverse to flood, the steep isohaline surfaces along the western side of the canyon relax downward and the process repeats. This same animation shows that, whereas the primary period of the isohaline oscillations at the entrance of Juan de Fuca Strait is semi-diurnal, further to the west and off the continental shelf, it is diurnal. Again, this is consistent with the vertical displacements shown in Figure 17b and the diurnal continental shelf waves that are known to be the predominant tidal currents in this region [Crawford and Thomson, 1984; Flather, 1988; Cummins et al., 2000]. Though internal tides may not be captured accurately with this hydrostatic model, there are insufficient observations in the canyon to quantify the simulation errors. Nonetheless, ROMS can be expected to perform as well as the hydrostatic Princeton Ocean Model that Cummins and Oey [1997] used to reproduce internal tides off the north coast of British Columbia with reasonable accuracy.

[44] In summary, the tides seem to provide three mechanisms for advecting the denser water that constitutes the bottom estuarine flow in the canyon and stratify onto the shelf just west of the strait entrance. The first is the flood tide upwelling off the cape that forms a dome (and eddy) of denser water which grows and migrates westward; the second is the spilling of denser bottom water over the western wall of the canyon on the ebb tide; and the third is the tidally rectified vertical velocities that, as illustrated in Figure 15, enhance the upwelling in the canyon off Cape Flattery.

[45] Given this apparent role of the tides in upwelling denser water from the lower depths of Juan de Fuca Canyon and Strait, it should not be surprising that the model simulations show fortnightly variations in this upwelling as the tides progress through spring to neap cycles. Spring tides cause stronger upwelling and tend to push the upwelled water and the eddy further to the west. However, this process is complicated because the relatively strong diurnal currents not only mean that the two floods (ebbs) generally arising each day do not have equal strength, but also the two spring tides per month are not equal. Similarly, it is likely that the wave-like feature propagating westward across the bottom of the canyon on ebb tide may not be sufficiently strong during neap tides to spill over onto the Vancouver Island shelf as seen in Figure 19b. Though beyond the scope of the present study, this hypothesis also warrants further investigation.

[46] Subsidiary runs were also carried out with only the diurnal or semi-diurnal tides to determine their relative importance. Both produced sufficient upwelling to generate an eddy, though the eddy arising with only semi-diurnal tidal forcing was larger. Snapshots (not shown) similar to those in Figure 19 showed that the semi-diurnal run also produced sharper isohaline gradients along the western edge of the canyon than the diurnal run. However, neither constituent on their own displayed sufficiently steep gradients that the 33.7 psu isohaline spilled over onto the adjoining Vancouver Island shelf, as it does in Figure 19b. It therefore seems that both the diurnal and semi-diurnal constituents play a role in this phenomenon and they may have to be in phase (e.g., during spring tides) for the spill-over to happen.

[47] In order to determine if the cross-channel seiching associated with Ekman transport dynamics that was recently identified further eastward in Juan de Fuca Strait [Martin et al., 2005] might be playing a role in upwelling off Cape Flattery, the salinity across transect A (Figure 1) for days 50–52 of the baseline simulation was examined more closely. Consistent with the hypothesis put forward by Martin et al. [2005], bottom boundary layer isohalines tilted upward along the north side of the strait at flood tide (corresponding in time to the surface salinities and velocities shown in Figure 18a), but the isohalines above 100 m depth tilted downward, consistent with an estuarine outflow that hugs the north side. However, a snapshot corresponding to the ebb tide (Figure 18b) also had the bottom isohalines continuing to tilt upward to the north side of the strait while near surface salinities near Cape Flattery were slightly fresher than for the flood tide. This would seem to be in disagreement with the Martin et al. [2005] hypothesis that the near-bottom Ekman transport should reverse and have bottom isohalines tilting upward to the south on the ebb tide. Consequently, it does not appear that the phenomenon identified by Martin et al. [2005] is playing a role in upwelling off Cape Flattery.

6. Summary and Discussion

[48] The numerical simulations discussed in the preceding sections demonstrate that the Juan de Fuca Eddy can be generated by: average summer upwelling favorable winds and an estuarine flow; tides and an estuarine flow; and all three forcing mechanisms working together. A weak eddy was also generated by average winter downwelling favorable winds, tides, and an estuarine flow, although this simulation was not initiated with winter climatology and did not include a Columbia River discharge that would be expected to further suppress upwelling off Cape Flattery. Comparisons of tidal and average model current harmonics with those obtained from the analysis of current meter records verified that the model currents were reasonably accurate, thereby establishing credibility for the model dynamics and confirming that the model resolution is sufficient to capture important topographic effects.

[49] There appear to be several mechanisms (Figure 20) that contribute to the enhanced upwelling of continental slope (primarily CUC) water off Cape Flattery and the subsequent formation of the eddy. All are contingent on the proximity of dense water to the cape and this is
Figure 20. Schematic showing mechanisms contributing to the generation of the Juan de Fuca Eddy.

guaranteed by the presence of an estuarine flow in Juan de Fuca Strait and the fact that the 200 m contour of the canyon comes as close as 4 km to Tatoosh Island off Cape Flattery. This makes dense slope water more readily accessible for upwelling here than at locations further south along the Washington coast. Upwelling favorable winds, in combination with this estuarine flow, were shown to produce enhanced upwelling off the cape and form the Juan de Fuca Eddy. Tidal forcing in combination with the estuarine flow was also demonstrated to be sufficient to produce an eddy. In this case, $M_2$ vertical excursions of up to 25 m at 100 m depth and nearly 20 m at 50 m depth along the east side of the canyon transport dense water toward the surface where it is mixed with fresher estuarine outflow and transported offshore by the surface estuarine flows and tides. Lesser, but by no means insignificant, vertical excursions were also computed on the west side of the canyon due to both the $M_2$ and $K_1$ tidal constituents. In particular, snapshots of isohaline movement over a tidal cycle showed that during spring tides, deeper saltier water moves across the bottom of the canyon with the ebb tide, up the western flank of the canyon, and spills over the canyon edge onto the adjoining shelf. Though these cross canyon motions are suggestive of internal tides, further investigations (beyond the scope of the present study) are needed to better understand the underlying dynamics. It is likely that the large $K_1$ vertical excursions seen (Figure 17b) along the western side of Juan de Fuca Canyon are associated with the diurnal shelf waves that are known to exist along the Vancouver Island continental shelf [Crawford and Thomson, 1984; Flather, 1988; Cummins et al., 2000]. Further study is also needed to establish this link.

[50] The enhanced upwelling mechanism described above may be applicable to other regions where the continental shelf is narrow and deep water is more readily accessible for upwelling through either tidal or wind-forcing. Two highly productive regions where it may apply are Brooks Peninsula off northwestern Vancouver Island, and the Kaikoura Canyon off eastern New Zealand. Needless to say, in both cases process studies analogous to those conducted here are needed to determine the primary dynamics.

[51] Although the preceding results differ markedly with the eddy formation mechanism suggested by FD82 and FM89, their hypothesis that an approximate geostrophic balance between an inward pressure gradient and outward Coriolis over the Tully Canyon will cause water to flow up the canyon may be valid. Indeed, an analysis of average term balances in the momentum equations over the last fifteen days of the baseline run (A) confirmed that near the head of the Tully Canyon, there was a strong balance between the Coriolis and pressure gradient terms at 35 m depth. Snapshots of salinities along transect B culminating in Figure 12e suggest that FD82 and FM89 may have identified a secondary mechanism that partially sustains the eddy once it moves to this location. In fact, it may be that these canyon dynamics lock the Juan de Fuca Eddy in place and prevent it from drifting further westward like the Haida Eddy [Di Lorenzo et al., 2005].

[52] There is also the associated question of whether a new eddy grows off the cape when one is already established off Tully Canyon. The fact that there is no observational evidence pointing to more than one eddy suggests that newly upwelled water off the cape must move into the existing eddy; and this would be consistent with the large tidal excursions in the region. In their simulations, Di Lorenzo et al. [2005] found that an existing Haida Eddy had to be sufficiently far offshore before a new eddy formed. When it was too close, the buoyant water just fed the existing eddy and it got bigger. Clearly, further simulations are needed to confirm all these speculations. Nevertheless, the results presented here do indicate a different eddy formation mechanism than is suggested by FD82 and FM89 and, contrary to ruling “out a tidal origin for the eddy”, they actually suggest that the tides play an important role in its formation.

[53] Our winter wind simulations did not include a sustained Columbia River outflow and the presence of a Columbia plume near the entrance to Juan de Fuca Strait can be expected to restrict the tidal upwelling off Cape Flattery. Thus even though our simulation with average winter winds, tides, and an estuarine flow did show a weak eddy, we suspect that the presence of Columbia plume water off the cape would provide sufficiently strong stratification to further suppress the tidal upwelling so that an eddy would not be formed. This is consistent with observations not showing any evidence of an eddy in winter.

[54] In addition to the follow-up studies mentioned previously, additional simulations are required to better understand eddy dynamics under the influence of time varying wind fields. Summer winds are highly variable and observations indicate that when upwelling favorable winds are interrupted by storm (downwelling) events lasting a few days, the surface water property characteristics of the eddy disappear. This would be consistent with our simulations with winter winds and the Hickey et al. [2005] findings that the Columbia plume moves northward along the Washington coast during these events. These summer storms also transport eddy water that may include the domoic acid produced as part of harmful algal blooms [Trainer et al., 2002] to the Washington and Vancouver Island coasts where they can impact shellfish beds. Simulations to better under-
stand these phenomena are presently underway and will be reported in future manuscripts.

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