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The Columbia River Plume Study: Subtidal variability in the velocity and salinity fields

B. M. Hickey,¹ L. J. Pietrafesa,² D. A. Jay,³ and W. C. Boicourt⁴

Abstract. A comprehensive study of the strongly wind driven midlatitude buoyant plume from the Columbia River, located on the U.S. west coast, demonstrates that the plume has two basic structures during the fall/winter season, namely, a thin (~5–15 m), strongly stratified plume tending west to northwestward during periods of southward or light northward wind stress and a thicker (~10–40 m), weakly stratified plume tending northward and hugging the coast during periods of stronger northward stress. The plume and its velocity field respond nearly instantaneously to changes in wind speed or direction, and the wind fluctuations have timescales of 2–10 days. Frictional wind-driven currents cause the primarily unidirectional flow down the plume axis to veer to the right or left of the axis for northward or southward winds, respectively. Farther downstream, currents turn to parallel rather than cross salinity contours, consistent with a geostrophic balance. In particular, during periods when the plume is separated from the coast, currents tend to flow around the mound of fresher water. At distances exceeding about 20 km from the river mouth, the along-shelf depth-averaged flow over the inner to midshelf is linear, and depth-averaged acceleration is governed to lowest order by the difference between surface and bottom stress alone. In this region, along-shelf geostrophic buoyancy-driven currents at ~5 m (calculated from surface density) and along-shelf geostrophic wind-driven currents (computed from a depth-averaged linear model) are comparable in magnitude (~10–25 cm s⁻¹).

1. Introduction

The Columbia River is the largest river on the Pacific coast of North America, accounting for 77% of the total drainage along the coast between San Francisco and the Strait of Juan de Fuca. The Columbia plume provides an excellent natural laboratory in which existing numerical and analytical models for midlatitude river plumes [e.g., Chao and Boicourt, 1986; Chao, 1988a, b, 1990; Garvine, 1984, 1987; Zhang et al., 1987; Weaver and Hsieh, 1987; Oey and Mellor, 1993; Kourafalou et al., 1996a, b] can be tested and in which the importance of various physical processes can be assessed. Strong river flow, wind forcing, and tidal forcing cause the Columbia River plume to be an extremely dynamic feature where a wide variety of plume processes may be studied.

The prevailing shelf circulation in this region is reasonably well understood. Currents over the shelf are predominantly northward in fall and winter and southward in spring and summer [e.g., Smith and Hopkins, 1971; Hickey, 1989; Strub et al., 1987]. Fluctuations occur in all seasons with periods of 2–10 days. Such fluctuations are largely driven by the local wind during the winter storm period and by both local and remote winds (via propagating coastal trapped waves) in spring and summer [Hickey, 1981, 1984; Battisti and Hickey, 1984]. The Columbia plume itself also has been examined on a seasonal basis, so that the basic seasonal structure and orientation of the plume are well defined. The plume in winter is generally directed northward from the estuary mouth and is mostly confined to the shelf. In summer the plume turns to the southwest after crossing the shelf and is found seaward of the shelf off Oregon and California [Barnes et al., 1972].

The Columbia estuary has also been the subject of several studies [Giese and Jay, 1989; Hamilton, 1990; Jay and Smith, 1990a, b]. The width of the estuary at its mouth is about 4 km, and the depth over the bar is about 20 m. The ratio of the estuary width at the mouth to the baroclinic Rossby radius (~15 km) is typically about 0.25 (the Kelvin number). Plume volume varies between 2 and 11 × 10¹⁰ m³. The dominant timescales of plume formation are the diurnal and semi diurnal tidal modulation of estuary discharge, tidal monthly changes in stratification (strongest during the low-flow season, mid-July to October), and seasonal changes in river flow. The tidal prism (defined as the integrated volume between mean lower low and high waters) varies from about half the river flow volume (neap tides during a strong freshet) to 10 times the river flow volume (spring tides and low river flow). River flow into the estuary varies from about 3 to 17 × 10¹⁰ m³ s⁻¹ over a typical year. Maximum discharge occurs during late spring snowmelt freshets and during winter storms. The density field within the estuary normally alternates between two states: one, weakly stratified or partially mixed, which occurs during low-flow periods with strong tides; the other, highly stratified (nearly salt wedge), which occurs under most other conditions.

The Columbia River estuary and plume system provides important contrasts to other recently studied North American plumes. In particular, because of the relatively narrow entrance, large astronomical tides, and robust river flow, the

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Columbia plume is more strongly forced at the estuary boundary than other systems. For example, tidal currents are 2-3 times larger at the mouth of the Columbia River than at the mouth of the Chesapeake Bay, and river flow is 3-5 times that of the Chesapeake (Boicourt et al., 1987, 1994). Because the Columbia River entrance cross-sectional area is about one fifth that of the Chesapeake Bay and the Columbia River flow is also larger, the volume flux per unit area from the Columbia estuary to the ocean is an order of magnitude greater than that of the Chesapeake under mean conditions. The midlatitude Columbia plume, which is strongly affected by rotation, also offers an interesting contrast to the low-latitude Amazon plume, in which rotational effects are weak (Lenz, 1995).

A field study focused on the time-dependent formation of the Columbia River plume and its interaction with ambient currents and local wind stress was carried out in fall/winter 1990-1991. The objectives of the study were to examine the dominant processes affecting plume structure and time-dependent evolution over the shelf, namely, mesoscale and low-frequency advection of the plume, the role of mixing processes in plume evolution, and the effects of tidal advection and estuarine boundary conditions on plume structure. Cudaback and Jay (1996, also, lateral circulation and forcing in the Columbia River entrance, submitted to Continental Shelf Research, 1997) discuss plume formation, hydraulic control, and the lateral force balance near the estuary mouth. Jay and Flincham (1997) discuss barotropic tides in the plume near the river mouth. The present work addresses mesoscale plume processes, in particular, the effect of the plume on the velocity and salinity fields over the shelf.

The Columbia Plume Study is described in section 2. In section 3 the velocity data are presented, and the difficulty of separating buoyancy-driven and wind-driven processes is discussed. Examples of time-dependent plume behavior at the surface and a statistical analysis of velocity and salinity fields are presented in section 4. Deviations of the observed velocity field from that expected for purely wind driven dynamics are described in section 5. A first-order separation of along-shelf buoyancy-driven and wind-driven currents is presented in section 6. The paper concludes with a summary and discussion.

2. Columbia Plume Study

Satellite imagery obtained during the field study illustrates that the plume from the Columbia River is a dominant feature off the Washington coast (Plate 1). In this image the plume from the Columbia is visible in sea surface temperature and, because of its high sediment load, in surface albedo, from the estuary mouth to the Strait of Juan de Fuca. The Columbia Plume Study focused on the portion of the plume within about 100 km of the river mouth (see elements of the moored array shown on the infrared imagery in Plate 1). The experiment included simultaneous data from a 22-element moored velocity/temperature/conductivity/bottom pressure/meteorological sensor array, three 2-3 week conductivity-temperature-depth/ acoustic Doppler current profiler (CTD/ADCP) surveys, seven radio-tracked drifter surveys, satellite imagery, and ancillary coastal wind, sea level, and surface wave data.

The study took place from late October 1990 to early April 1991. The timing of the shipboard hydrographic surveys, drifter studies, and satellite data relative to ambient wind stress and currents is shown in Figure 1. The winter of 1990-1991 was extraordinarily stormy. Maximum daily significant wave height exceeded 3 m 42% of the days from mid-October through December. Wind stress was generally northward, reaching values of over 3 dyn cm^-2 on numerous occasions. During late winter, several extended periods of southward wind stress occurred, as is typical in the Pacific Northwest (Hickey, 1989).

Regional currents have a temporal pattern very similar to that of the winds: northward early in the season, with increasingly longer southward periods in late winter. Fluctuations in along-shelf wind and currents appear to be strongly correlated, as expected from earlier measurements in this region. River flow (which is an indication of the stratification and volume of the outwelling plume water) has a seasonal pattern, low in the fall, increasing as the winter season progresses. Shorter period variability in river flow is related to rainfall during storms and also to snowmelt.

2.1. Moored Array

The moored array consisted of 58 instruments supported on 14 surface toroidal moorings and eight subsurface moorings (Figure 2). Letters are used to identify moorings: from north to south, "W" indicates Willapa; "K," Klipsan Beach; "B," Long Beach; "N" and "S," moorings just north and south of the river mouth; "O," Oregon; and "EN" and "ES," moorings within the estuary itself on its north and south sides, respectively. Moorings located near the 10, 30, and 50 fathom (fm) (~20, 55, and 90 m) isobaths are identified with the numbers 1, 3, or 5 after their location label. Measurement depth in meters usually follows the mooring identifier. For example, "W3, 5 m" signifies 5 m data from the mooring on the 30 fm (55 m) isobath near the mouth of Willapa Bay. Final deployment depths of most moorings identified with a "1" were closer to 30 m than to 20 m. The 30 m isobath is on the inner shelf, and the 55 and 90 m (30 and 50 fm) isobaths are both on the midshelf (Figure 2). Because of the proximity of two submarine canyons, several of the 90 m "midshelf" moorings are near the shelf edge. However, these canyons are sufficiently narrow and the water column is sufficiently stratified that flow in the upper ~50 m of the water column does not "feel" the canyons (Hickey, 1997). Thus an "outer shelf" designation for the 90 m data would likely be inappropriate. The 55 and 90 m locations will be designated as "shallower" and "deeper" midshelf locations, respectively, in the remainder of the paper.

The array was designed to resolve the spatial scales of flow variability in the turning region of the plume, as well as in the downstream coastal current. Although the array emphasized northward tending plumes, moorings were also placed south of the river entrance to identify occasional periods of southward tending plumes and also to provide comparison for plume/nonplume conditions (O3, O5). Strong currents near the river mouth, location of shipping lanes, the presence of heavily fished areas, and the wave climate near the Columbia River bar placed constraints on the array design; the mooring closest to the river mouth (N1) was partially buried in sand during the experiment and had to be excavated by divers.

The basic measurement set for velocity within the plume consists of data at 5 and 10 m at all sites, with 1-2 m resolution at sites which included ADCPs. Near-surface currents were measured using vector-type current meters suspended from surface moorings (either an InterOcean S4 or an EG&G Vector Measuring Current Meter); ADCP current meters either mounted in a surface mooring looking downward (K3, N3) or on a taut-wire mooring looking up (S3); or by upward looking
Figure 1. Environmental setting of the Columbia Plume Study, as described by selected subtidal time series of along-shelf wind stress and 5 m currents as well as hourly time series of river flow into the estuary and total flow from the estuary (river flow plus tidal prism) expressed as volume per 12.4 hour tidal interval. Timing of CTD surveys (heavy bar), drifter deployments “D” and satellite imagery “S” are indicated along the x axis.

ADCPs mounted in bottom cages (N1, K1, K5). For redundancy, each ADCP was paired with an individual current meter mounted on the same mooring. Measurements beneath the plume and in the bottom boundary layer were made at S5, O3, and S3 at 4-20 m intervals. With the exception of measurements at O3, 20 m, which were made with an S4, deeper measurements were made using Aanderaa current meters on taut-wire moorings with top floats below 35 m.

At one midshelf site (K3), 5 m speed measured with a downward looking ADCP mounted on a surface toroid exceeded that from an InterOcean S4 by about 5-10 cm s⁻¹ during both northward and southward current events. The 5 m Doppler speeds at this site also exceeded speeds at other sites to the north, south, east, and west. We suspect that the surface-moored Doppler was, on average, sampling shallower-than-expected depths due to tilting of the surface toroid, but to date, we have been unable to substantiate this effect. Data from the S4 are used in all figures and calculations for the 5 m depth. Doppler data are displayed in vector time series at 10 m and are included in empirical orthogonal eigenfunctions (EOFs) (denoted “D” to differentiate from 5 m S4 data).

The basic measurement set for water properties consists of data at 1, 5, and 10 m, and at greater depths on moorings that included either temperature sensors (T-chains) or current meters. T-chains were mounted on five of the surface moorings to provide more detailed information on plume stratification at and near the river mouth as well as in the region downstream of the mouth (Figure 2). Vertical resolution of the T-chains ranged from less than a centimeter (in which long sensor strings were folded and refolded to keep the string above bottom) to a meter. Salinity data were obtained at a depth of 1 m on all surface moorings and at selected deeper depths at roughly half the sites.

Measurements in the moored array were made at 15 or 30 min intervals for most instruments. ADCPs were generally set to sample over hourly intervals. To remove inertial and dominant tidal oscillations, the hourly data were low passed using a Cosine-Lanczos filter with a 40-hour half-power point, and these data were decimated to 6-hourly values to form the “subtidal” data set used in this paper.

Because wind forcing was expected to dominate most time series and because spatial structure in the wind field can cause significant spatial differences in plume structure [e.g., Beardsley et al., 1987], wind speed and direction were measured at hourly intervals at seven sites in the expected region of the surface plume. Wind data were corrected to a nominal height of 10 m above the sea surface assuming a neutral stability log layer [Halliwell et al., 1986]. A variety of instrument types were used to measure wind speed and direction: Aanderaa (W3, K5, K1, N5, N1), Coastal Climate (K3), and National Data Buoy Center (NDBC) buoy 46010 (B10). Although records from the nonvector-averaging instruments were much noisier than those
Figure 2. Locations of moorings and types of sensors relative to the river mouth and bottom topography. Sensors that returned usable data are shown for each mooring. Depths of individual current meters (in meters) are given below the mooring code. Sensors at 1 m measured only temperature and conductivity. Bottom topography here and in all other figures is given in fathoms (1 fm = 1.8 m).
of the vector-averaging instruments, no significant differences were observed between the instruments after editing and filtering. Along-shelf surface wind stress was calculated from hourly wind data using the relationship $\tau'_x = \rho_a c_s |w| w_x$, where $w = (w_x, w_y)$ is the wind vector, $\rho_a$ is the density of air, and $c_s$ is the surface drag coefficient calculated according to Large and Pond [1981]. Hourly stress data were low-passed filtered and decimated to produce a 6-hour subtidal data set.

2.2. CTD Surveys

Thirteen CTD maps of the plume were obtained under a variety of environmental conditions. Conductivity, temperature, and depth measurements were made with either a SeaBird or a Neil Brown CTD. The survey procedure was to begin sampling offshore on the southern end of the plume region in an attempt to establish the southern boundary of fresh water for a northward tending plume. The survey vessel would then proceed toward the mouth of the Columbia River. After a rapid survey of the entrance region (designed to minimize tidal distortion), the survey proceeded with sequential cross-plume transects down the axis of the plume. Mapping nearshore regions in February at night was restricted by the density of crab pots.

2.3. Drifter Studies

Seven drifter deployments were interspersed among the hydrographic surveys. Typically, 10 “Davis” type surface drifters [Davis, 1985] were released near the river mouth during ebb
2.4. River Flow Data

Daily river flow into the Columbia estuary was synthesized from daily measured flow at Bonneville Dam (representative of the Eastern subbasin that contributes about 75% of the annual average flow) and for the Willamette River (representative of the coastal subbasin that contributes the remaining 25% of the annual average flow) [Jay, 1984]. Total estuary outflow (the sum of river flow over a 12.4 hour tide plus the tidal prism associated with that tide) was estimated for the mouth of the estuary. The tidal prism was calculated using Astoria tidal height predicted using techniques by Foreman [1977] and a barotropic one-dimensional, semi-analytical tidal model of the Columbia River and estuary [Giese and Jay, 1989].

3. Identification of the Columbia River Plume

A distinctive plume signature is observed in the shelf velocity field in results from available numerical models [e.g., Chao and Boicourt, 1986; Kourafalou et al., 1996a, b]. However, in spite of the volume of the Columbia River plume, time series of current vectors at 5 and 10 m from the sea surface over the shelf in the vicinity of the plume are remarkably similar to time series of currents at such depths anywhere in the Pacific Northwest during winter (Figures 3a and 3b) [e.g., Hickey, 1989]. For example, except at the site within the estuary mouth (ENS), the mean is generally northward, and fluctuations of several days' duration are superimposed on the mean. Fluctuations are highly correlated over the roughly 100 km long study region, and along-shelf current fluctuations at most sites appear to be correlated with along-shelf wind stress. Thus, with 5-10 m velocity data alone, it is difficult to detect specific plume effects in the time series. For example, much of the apparent northwestern tendency, which might be thought to be plume related, is actually due to the northwestern orientation of the local isobaths. Even the elevated mean northward flow observed at midshelf sites (N3, K3, W3) relative to inner shelf sites (K1, W1) could simply be due to lateral shear in the regional currents (e.g., the signature of a wind-driven coastal jet).

Comparison of velocity time series at 5 m (Figure 3a) with time series at 10 m (Figure 3b) and with data from depths deeper than 10 m (Figure 3c) shows that variability decreases with depth from the sea surface. Below 10 m, current speed and direction are relatively uniform with depth and are directed offshore of those at shallower depths, roughly parallel to the direction of the local isobaths. Evidence of veering in the bottom Ekman layer is seen at depths within 10 m of the bottom (e.g., 87 m in Figure 3c). The increase of variability in speed and direction toward the sea surface might be due to the presence of a buoyant plume. On the other hand, the increase could be due to other processes, such as frontal instabilities or surface Ekman layer dynamics.

In contrast to the velocity field, the plume from the Columbia is easily identified in maps of sea surface salinity. Mean surface (1 m) salinity and mean velocity at selected depths are shown in Figure 4. These means were calculated over the time period common to all instruments in the moored array, October 25 to November 28. Means calculated for time periods extending to December 16 (but at fewer locations) show spatial patterns almost identical to those shown for the period of common data. The freshest ocean water is observed west-northwest of the river mouth rather than adjacent to the coast. Maximum lateral gradients (~1 practical salinity unit (psu) km^{-1}) occur on the southwest side of the plume. At the two locations where data are available 5 m from the sea surface, the mean vertical salinity gradient is of the order of 0.25 psu m^{-1}. Minimum 1 m salinity at midshelf within 6 km of the river mouth is of the order of 24 psu, as compared with ambient surface salinity of about 32 psu and mean surface salinity at the river entrance of about 10 psu. These results suggest that significant entrainment and mixing occur between the river entrance and the first element of the moored array.

When velocity data are similarly presented, it is apparent that the mean near-surface velocity field is not as spatially uniform as expected for relatively narrow, wind-driven shelves (Figure 4). For mean northward along-shelf wind in the absence of a buoyant plume, the expected ambient flow would be northward, roughly following the local isobaths except in the surface and bottom boundary layers, where frictional effects cause the flow to cross isobaths. Mean wind-driven flow in the surface frictional layer would be expected to onshore relative to the local isobaths. In the region occupied by the Columbia plume, the mean flow at 5 m is onshore relative to the local isobaths only in regions more than 20 km north and south of the river mouth. Even at those sites, the direction of the mean flow is not to the right of the mean wind direction. The mean flow pattern at 5 m and, to a lesser extent, at 10 m roughly parallels the local salinity contours, turning outward from the river mouth and bending slightly shoreward again north of the mouth. With the exception of the station just south of the river mouth, flow decreases with depth. Largest mean speeds (~30 cm s^{-1}) are observed at 5 m over the 90 m (50 fm) isobath more than 25 km north of the river mouth. Mean flow is weakest over the inner shelf at all sites. At the inner shelf station 16 km north of the river mouth (K1), the mean flow is particularly weak at all depths, and the direction is weakly southeastward at a depth of 5 m, parallel to the salinity contours which bend shoreward near that station. The outflow velocity from the river mouth at 5 m is similar in magnitude to the 5 m flow over the shelf at sites near the river mouth. Mean flow at depths below 10 m is oriented roughly along isobaths at most depths and locations, consistent with weaker plume effects at those depths.

One explanation for the difficulty in identifying a plume velocity signal is that the majority of the plume may be confined above the shallowest measurement depth (4–5 m). Contoured salinity maps at 1, 5, and 10 m for four of the shipboard surveys demonstrate that this is indeed the case, particularly under conditions in which the plume axis separates from the coast north of the river mouth (Figures 5a and 5b). The best example of such a plume (January 20–22) was obtained during a period of weak southward stress and currents following 1.5 days of stronger southward wind stress and currents (about ~0.6 dyn cm^{-2} and ~20 cm s^{-1}, respectively). Although both February maps were obtained during periods of weak southward wind stress (approximately ~0.3 dyn cm^{-2}), the plume in the later survey is spread out farther across the shelf (Figure 5b). The difference in spatial structure between the two February surveys is likely due to the fact that the first survey began 2 days after a strong northward stress event (~2 dyn cm^{-2}), whereas the second survey occurred following 2.5 days of reasonably strong (~0.5 dyn cm^{-2}) southward
Figure 3a. Time series of subtidal velocity vectors 5 m from the sea surface for the Columbia Plume Study. Vector wind at B10 is shown at the top of the figure. Station locations are shown on the inset map.
stress. The surveys show that westward and northwestward tending plumes are relatively thin and strongly stratified within the plume (vertical salinity gradient along the plume axis of the order of 1 psu m$^{-1}$). The width of such plumes decreases with depth, extending over the continental slope at the surface, but only to about the shelf edge at 5 m from the sea surface in the examples shown. Maximum lateral salinity (and hence density) gradients also occur farther offshore at the sea surface than at greater depths. Plume orientation can differ with depth; in the January 20–22 example, the plume axis at 1 m is to the west of the axis at 5 m (Figure 5a); in the February 24 example, the plume axis is directed west-northwestward at the surface but southwestward at 5 m (Figure 5b).

A cross-shelf density section through the January west-northwestward tending plume illustrates in more detail the tendency for such plumes to be surface intensified, to be strongly stratified within the plume, and to have little or no contact with the shelf bottom downstream from the mouth of the estuary (Figure 6). These characteristics are repeated in several other examples of such separated plumes obtained in the Columbia River Study (not shown). (The term “separated” is used to describe a plume whose axis, except near the river mouth, is located at some distance from the coast.) Historical data suggest that water denser than about 24 $\sigma$, and saltier than about 32 psu is not associated with a specific river plume during early winter [Landry et al., 1989]. Using this criterion, total plume thickness for separated plumes ranges from 5 to 15 m. However, the majority of the plume volume and the strongest lateral density gradients are usually confined above 10 m.

The map of October 25–26 is the best example of a plume under strong northward wind stress conditions (>2.5 dyn cm$^{-2}$) (Figure 5a). Note, however, that this example was obtained early in the winter season, when river flow was relatively low (see Figure 1). Both river flow into the estuary and total outflow from the estuary in the October survey are smaller than in the other surveys. The northward October plume hugs the coast, barely extending to midshelf. The plume is thicker than the northwestward (separated) plumes, with a significant signature at 10 m, but less at 15 m (not shown), and relatively weak stratification within the plume (of the order of 0.25 psu m$^{-1}$). Like the northwestward plumes, the northward plume appears less continuous at depth. A vertical section showing density across the plume illustrates the tendency for lighter plume water to have more direct bottom contact than in separated plumes (Figure 6). Comparison of the vertical density section with that for the January plume illustrates that the northward plume has weaker within-plume stratification than the northwestward plume. Total plume thickness in this and other
examples (not shown) ranges from 10 to about 40 m, with most of the volume and spatial gradients confined to the upper 20 m.

Surface drifter data provide direct evidence of the existence of substantial velocity shear in the upper 5 m (Figure 7). Surface speeds and directions obtained from the drifters were compared with hourly (unfiltered) 5 m data from the moored array at times when drifters passed within 2 km of a mooring. Although directions at the surface and 5 m are reasonably similar, surface speeds are consistently higher than those measured at 5 m; surface speeds often approach 100 cm s$^{-1}$, whereas those at 5 m rarely exceed 50 cm s$^{-1}$. Vertical shear in the upper 5 m, likely one factor in the weaker-than-expected plume signal at 5 m, is being explored in a separate paper.

Another potential explanation for the absence of an obvious plume velocity signal at 5 m is that it is roughly phase locked with the along-shelf wind-driven currents. This idea is explored in later sections, where a lowest order separation of wind and buoyancy-driven contributions to the observed velocity field is presented.

4. Time-Dependent Plume Behavior
4.1. Selected Examples

When daily 5 and 10 m velocity data from the moored array are viewed with surface salinity maps, plume effects become apparent at 5 m and sometimes at 10 m: flow directions at 5 and 10 m are not spatially uniform and flow at 5 and 10 m is often in markedly different directions (Figure 8). Although the
latter result might be thought to be related to the existence of large surface Ekman currents at 5 m, the 5 m currents are not simply to the right of the interior flow direction as would be expected for simple Ekman dynamics. The most dramatic departures from expected wind-driven behavior occur during periods of "relaxation" (or slight reversal) of the seasonal northward winds, when the plume is directed northwest of the river mouth, crossing the shelf isobaths (e.g., on November 20 in Figure 8).

The temporal sequence shown in Figure 8 includes two northward wind stress events (November 22-25 and 29-30) and two southward wind stress events (November 20-21 and 26-28). The northward stress event with a maximum on November 24 is the largest wind stress event of the study period: measured hourly winds exceeded 20 m s$^{-1}$ during this storm. Following this storm, river flow doubled due to the rainfall associated with the storm (see also Figure 1). Therefore the map sequence encompasses periods of both high and low river flow as well as northward and southward wind stress. As deduced also from the CTD survey data, two spatial patterns dominate: plumes with axes separated from the coast during or just following periods of southward winds or weak northward winds ("relaxation events") (e.g., November 20 and 27) and plumes hugging the coast during periods of strong northward winds (November 24-25). The transitions between these two patterns are clearly delineated in the sequence of maps.

The sequence begins with a relaxation event that follows the northward wind event of November 17-19. Flow along the plume axis at a depth of 5 m is directed slightly south of the axis, consistent with the existence of wind-driven frictional flow in the upper 5 m driven by the weak east-southeastward wind stress. Flow on the inner shelf out to midshelf is southeastward at 5 m at most sites. However, flow at the inner shelf site closest to the river mouth (N1) is strongly southwestward during this and other relaxation events. On November 22, as northward wind stress begins, the flow at 5 m turns abruptly onshore at most sites, and 5 m flow is generally to the right of 10 m flow, as expected in the surface Ekman layer. The 5 m flow at midshelf at the northern part of the measurement grid (W3) has roughly the same magnitude and direction as that at midshelf at the southern part of the array (O3), where plume influence is minimal. Flow at 5 m on the west and north sides of the plume roughly follows the salinity contours, consistent with some degree of geostrophic balance, for which flow would be expected to parallel density (or salinity) contours. As the wind continues to build, 5 m vectors turn more northward (November 23). In contrast to the pattern observed during the weak southward wind stress event (November 20), 5 m flow along the plume axis is now directed slightly to the right of the plume axis, consistent with onshore surface Ekman layer flow. When wind stress is at a maximum (November 24), the plume as defined by the 30 psu contour is confined to the inner half of the shelf. The water on the outer half of the shelf and south of the plume attains salinities greater than 32 psu.

Maximum currents are generally observed November 25, when wind stress has already begun to weaken (or "relax"). At the onset of relaxation, 5 m vectors at most plume sites turn offshore (November 25). Currents at 5 m at the midshelf site off Oregon remain onshore, as expected during northward wind stress events in the absence of a river plume or other...
Figure 5a. Maps of salinity (psu) at selected depths as obtained from CTD surveys during the periods October 25–26, 1990, and January 20–22, 1991. Inset time series show the survey period (bold bar on the x axis) relative to along-shelf wind stress (solid curve) and along-shelf, midshelf currents (dashed curve) at 5 m (if available). The October survey occurred during a period of strong northward wind stress with low river runoff and low estuary outflow; the January survey occurred during a period of relatively high runoff and weak southward wind stress following a prolonged period of strong southward wind stress. Station locations are shown as dots. Regions in which salinity is less than 28 psu are shaded. Contour interval is constant over the maps, and the same interval (2 psu) is used at 1 and 5 m; an interval of 1 psu is used at 10 m.
Figure 5b. As in Figure 5a, for the periods February 21–22 and 24, 1991. Both surveys occurred during periods of weak southward wind stress and relatively high river flow. However, the first survey began 2 days after a strong northward wind stress event, whereas the second survey occurred just following 2.5 days of reasonably strong southward wind stress. Crab pot density prevented sampling close to the coast during the February 21–22 survey.
mesoscale phenomena. The shape of the plume on November 25 is that expected for a well-developed plume in the northern hemisphere in the absence of strong winds: an offshore bulge followed by a plume turning region, culminating in a narrower buoyantly driven coastal current. Rainfall increases, and both river flow and total estuary outflow increase following the November storm. Near-surface salinity within the estuary is less than 4 psu for the next several days. The area of freshest water increases significantly during this period.

As northward winds continue to weaken, currents begin to turn southward on the inner shelf north of the plume. The best example of the plume in a weak wind environment occurs on November 27. The plume is west-northwestward, and 5 m currents along the axis are oriented nearly parallel to the axis, crossing the salinity gradient. Speed increases down the axis, and speeds are highest in the region where the plume turns, as in the other examples of relaxation events. Currents at most sites on the inner to midshelf near the northern edge of the plume are southward at 10 m but onshore at 5 m, roughly parallel to salinity contours. The winds remain weak (although slightly southward) the following day (November 28), and the area of freshest water continues to increase.

Wind stress turns northward once again on November 29, and the 5 m vectors respond immediately with onshore flow; the plume moves onshore as on November 24. The wind immediately relaxes, and flow at all plume sites turns offshore on November 30, whereas 5 m flow at the Oregon site outside the plume remains slightly onshore. By the following day the plume is again directed northwestward rather than northward, and flow on the inner shelf is onshore or southward.

4.2. Statistical Analysis

To determine the spatial structure of velocity and salinity fluctuations, EOF analysis [Kundu and Allen, 1976] was performed for the time period common to all data records (October 25 to November 28) as well as on a longer time period (October 25 to December 25). Results from the two periods are not significantly different in either dominant spatial patterns or temporal relationships, and only results from the shorter time period are shown.

4.2.1. Velocity field. EOF analysis of the velocity field separates the flow field into a mode representing primarily along-isobath velocities (the first) and a mode with significant cross-shelf velocities (the second) (Figure 9). The amplitude of the first mode, which accounts for 70% of the total variance in the velocity field, increases from the inner to the midshelf but decreases to near zero at some sites closer to the shelf edge. The time series associated with the first mode is strongly correlated with the along-shelf component of wind stress ($r = 0.8$), lagging it by 0.5 days (Figure 10, upper left), suggesting that this EOF represents the variance of the regional wind-forced along-shelf currents. (A correlation exceeding 0.48 is significant at the 95% level of confidence [Koopmans, 1974].) However, EOF vectors are not strictly aligned with the isobaths, as is commonly found in strongly wind-driven regions with slowly varying topography, and the amplitudes and directions of the vectors have significant spatial variability. The amplitudes are unidirectional and decrease with depth at most
Figure 8. Daily velocity vectors and contoured maps of surface salinity from November 20 to December 1, 1990. Solid arrows indicate 5 m data, and dashed arrows indicate 10 m data (20 m data at O3). Salinity data were obtained from the moored sensor array at 1 m. The actual value of salinity at the estuary location is given in each figure. Regions with salinity less than 25 psu are shaded. Contours for 10, 20, 25, 30, and 32 psu are solid; additional contours are dotted. Contours extrapolated subjectively are dashed. Wind (W) and river flow (R) are shown with scaled arrows on the right side of each figure.

sites. The spatial structure of this EOF is significantly different from that obtained during summer in a region with no buoyant forcing [Winant et al., 1987]. In that study, the amplitude of the first mode of the velocity field decreased from midshelf to the shelf edge by about 60%. In the present case, the decrease at some sites to the stations closest to the shelf edge (about 10–20 km from the edge) is almost 90% (e.g., K5). The analysis presented in section 6 demonstrates that the first EOF mode includes both wind- and buoyancy-driven along-shelf currents.

The spatial structure of the second EOF is more complex than that of the first EOF. Flow is primarily across the isobaths at 5 and 10 m rather than along isobaths, as for the first EOF, and vertical structure is significant. Correlation with along-shelf wind stress is strong ($r = 0.7$), but the phase relationship is opposite for the first mode (Figure 10, lower left). Flow is more offshore at 5 m than at 10 m during periods of southward stress and more onshore at 5 m than at 10 m during periods of northward stress at most sites, consistent with Ekman veering with depth. For a spatially uniform wind stress such as is observed over the plume region (see section 5), we would expect a relatively uniform spatial current pattern at each depth and similar amplitudes at all sites. However, the EOF shows substantial spatial differences; e.g., during periods of southward wind stress and also weak northward stress, flow is more southwestward or southward at locations on the inner and shallow midshelf and more northwestward at locations farther seaward. During periods of northward wind stress, the pattern reverses, with flow more northward and onshore at locations close to the coast, and more southeastward at locations farther offshore and off Oregon. The southeastward flow, when added to the seasonal mean flow, represents a reduction in northward flow rather than an actual southward flow. Note that although this mode contributes only a few percent of the variance to along-shelf flow at midshelf, where the amplitude of the first mode is large, the mode dominates the along-shelf variance on the outer midshelf. Thus this mode expresses
Figure 8. (continued)

4.2.2. Salinity field. EOF analysis of the 1 m salinity data produces two dominant modes, together accounting for 75% of the variance. The first mode (57%) describes the plume when it is separated from the coast north of the river mouth; the second mode (18%) describes the plume when it hugs the coast (Figure 11). Time series of the salinity represented by the modes indicate that both modes, although not themselves correlated, are strongly related to the first mode of the velocity field (Figure 12). Correlations are 0.74 (mode 1) and 0.52 (mode 2), with zero lag in both cases. The zero lag is consistent with a geostrophic response of the velocity field to the presence of the plume and its lateral pressure gradients. Episodes of northward ambient currents are associated with lower salinities near the coast and higher salinities west-northwest of the river mouth. During episodes of southward ambient currents, when the plume axis is separated from the coast, salinity is lower north-northwest of the river mouth and higher near the coast.

5. Wind-Driven Dynamics Within and Near the Plume

Separation of time-dependent buoyancy-driven currents from regional wind-driven currents is not a simple task because the signals likely co-vary. (The term “buoyancy-driven” is used in this paper to describe currents related to the density field of the buoyant plume, without implying a particular type of dynamics.) For example, during a northward wind stress event, northward quasi-geostrophic wind-driven currents develop over the shelf. At the same time, the density field associated with any existing river plume is advected toward the coast by currents in the surface Ekman layer, and the density-related quasi-geostrophic currents (usually northward) are also observed. In this section a local wind-driven model is used to determine the amplitude and temporal variability of wind-driven along-shelf currents in the vicinity of the Columbia plume. In the following section a combination of currents predicted by this model and geostrophic currents estimated from

Figure 8.
gradients in measured surface density data are used to separate wind and buoyancy contributions to the along-shelf geostrophic flow in regions downstream of the plume turning region.

5.1. Spatial Variation of the Local Wind Field

Spatial gradients in wind stress can have significant effects on the spatial structure of wind-driven currents over the shelf [Beardsley et al., 1987; Brink et al., 1987; Enriquez and Fribe, 1995]. However, wind time series for the Columbia plume region show that both the amplitude and the direction of the wind field had no consistent spatial variability over the study site during this fall/winter period (Figure 13). EOF analysis confirms that the wind field is highly uniform over the region: 97% of the variance is contained in the first two EOF modes (74% and 23% for modes 1 and 2, respectively) for the period common to all four records. The amplitude is slightly greater at the site near the mouth of the Columbia (B10), perhaps because of topographic steering of winds by the Columbia River gorge. The first mode contains most of the along-shelf wind variance, and the second, the cross-shelf variance.

The above results have two important implications. First, spatial variation in the wind field cannot be responsible for a significant amount of the observed spatial variability in the velocity field and/or plume movement. Second, from a practical point of view, any one time series can be used to represent the wind field over the plume. Wind stress computed at the location closest to the measured currents (K5) is utilized in the model presented below.

5.2. Along-Shelf Flow

A comprehensive analysis of momentum within the Columbia plume is beyond the scope of this paper. However, with several simplifying assumptions, insight into the relative roles of buoyancy and wind forcing of along-shelf currents can be obtained for regions outside the highly nonlinear turning region of the plume. The along-shelf momentum equation can be expressed as

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} + f u = \frac{1}{\rho} \frac{\partial p}{\partial y} - \frac{1}{\rho} \frac{\partial \tau'}{\partial z}$$

(1)

where $u$, $v$, and $w$ are the cross-shelf, along-shelf, and vertical components of velocity (positive onshore, northward, and downward, respectively), $t$ is time, $\rho$ is density, $p$ is pressure, $f$ is the Coriolis parameter, and $\tau'$ is the along-shelf component of stress. With the assumption that the system is semigeostrophic downstream of the turning region (an assumption supported by models such as Chao and Boicourt's [1986]), the cross-shelf momentum equation can be expressed as

$$f u = \frac{1}{\rho} \frac{\partial p}{\partial x}$$

(2)

A simple observationally based geometry for along-shelf velocity vertical structure is shown in Figure 14. In the deep layer of thickness $H-h$ beneath the plume, along-shelf velocity $v_d$ will be assumed depth independent. In the upper (plume) layer of thickness $h$, along-shelf velocity ($v = v_p + v_d$) is vertically
Figure 10. (left) Comparison of first ($E_1^r$) and second ($E_2^r$) velocity EOF time series with along-shelf wind stress $\tau$. Note that the scale for $E_1^r$ is double that for $E_2^r$. (right, top) Comparison of the first-mode EOF time series at W3 at a depth of 5 m with depth-averaged along-shelf flow from a purely wind-driven model ($v_w$). (right, bottom) Comparison of the sum of the first and second modes with the sum of the adjusted wind-driven currents and geostrophic buoyancy-driven currents ($v + v_w$). The value of the resistance coefficient used for the wind-driven model was 0.05 cm s$^{-1}$. Note that a zero offset between the models and the EOF was required to compensate for the removal of the mean in the EOF analysis.

Figure 11. Amplitudes of the first ($E_1^s$) and second ($E_2^s$) EOFs of the salinity field for the period October 25 to November 28, 1990. Data locations are shown as small dots. Extrapolated contours are dashed.
sheared. In keeping with previous winter studies in this region [Hickey, 1984], we will assume that both nonlinear terms and the along-shelf pressure gradient term are negligible beneath the plume. Assuming negligible along-shelf pressure gradient below the plume is consistent with the observation that effects due to propagating waves are negligible during the winter storm period when local wind stress is large and increases toward the north (in the direction that waves propagate) [Hickey, 1989]. With these assumptions, averaging (1) over the entire water column gives

\[ \frac{\partial \hat{v}}{\partial t} + \frac{h}{H} \frac{\partial v}{\partial x} + \frac{h}{H} \frac{\partial v}{\partial y} + \frac{h}{H} \frac{\partial v}{\partial z} + f\hat{u} = -h \frac{\partial \rho}{\partial y} + \frac{\rho H}{\rho}\ ]

where \( \hat{v} \) and \( \hat{u} \) are the along-shelf components of surface and bottom stress and

\[ \hat{v} = \frac{1}{H} \int_0^H v \, dz \quad \hat{u} = \frac{1}{H} \int_0^H u \, dz \]

are the depth-averaged along-shelf and cross-shelf components of flow, respectively. An overbar is used to indicate a vertical average over the plume, and a circumflex indicates a vertical average over the entire water column.

Lowest order estimates of terms in (3) were obtained using 5 m data from locations W1, W3, and K3 to form gradients using simple differences. Depth-averaged acceleration was estimated from 5, 10, 20, 30, and 45 m data at midshelf. Based on results of CTD surveys, which show that most of the lateral density gradients are above 10 m, we use an average plume depth of 10 m. The plume-averaged along-shelf pressure gradient term was estimated according to methods described in detail in the next section. Results from the EOF analysis show that cross-shelf current fluctuations that contribute to the large-scale velocity patterns occur primarily in the surface and bottom boundary layers (see Figure 9). The cross-shelf Coriolis term was therefore estimated using data 5 m from the surface and 5 m from the bottom, each applied over a 10 m interval of the water column. A crude estimate of the term that includes vertical velocity was obtained using measured vertical shear and maximum vertical velocity taken from a plume model (0.02 cm s\(^{-1}\)) [Chao and Boicourt, 1986]. The depth-averaged magnitude of this term is of the order of the other two nonlinear terms (not shown).

Comparison of the estimated depth-averaged nonlinear terms with along-shelf surface stress and acceleration demonstrates that although nonlinear terms are likely important for the details of flow within the plume, wind stress is sufficiently large and the shelf sufficiently deep that nonlinear terms contribute negligibly to the along-shelf momentum balance averaged over the entire water column (Figure 15). Bottom stress and depth-averaged along-shelf pressure gradient are significant, although their magnitudes are less than half that of along-shelf wind stress. The Coriolis term often exceeds both the along-shelf pressure gradient and bottom stress terms. However, we note that this term is partially compensated by the along-shelf pressure gradient (\( r = -0.6 \)). This suggests that cross-shelf velocity within the plume in the region downstream of the mouth has a significant geostrophic component, as mentioned in the qualitative discussion of examples of the plume velocity field (Figures 8a and 8b).

If (3) is linear to lowest order, and recalling that the along-shelf pressure gradient was assumed negligible beneath the plume, we can separate the depth-averaged flow into buoyancy- and wind-driven components, where the wind-driven component \( v_r \) satisfies the equation

\[ \frac{\partial v_r}{\partial t} = \frac{\tau^y_r - \tau^z_r}{\rho H} \]

This equation has been used successfully to model depth-averaged wind-driven along-shelf flow in the absence of river plumes [e.g., Hickey, 1984]. If bottom stress is expressed as a linear function of depth-averaged velocity

\[ \tau^z_r = \rho r v_r \]

where \( r \) is a resistance coefficient, (4) becomes

\[ \frac{\partial v_r}{\partial t} = \frac{\tau^y_r - \rho r v_r}{H} \]

Integrating (6), the depth-averaged along-shelf velocity driven by along-shelf wind stress is given by

\[ v_r = v(0) e^{(-\rho r t / H)} + \int_0^t \frac{\tau^y_r}{\rho H} e^{(\rho - r) t / H} \, dt' \]

Brink et al. [1987] demonstrate that details of amplitude and phase relationships obtained from such a model are strongly dependent on the choice of resistance coefficient. Surface gravity waves were vigorous during the Columbia plume experiment, and hence the resistance coefficient would be expected to be greater than in the absence of waves [Grant and Madsen, 1979]. To include wave effects, the resistance coefficient was set at 0.05 cm s\(^{-1}\) for our standard case. For comparison, a
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Figure 13. (top) Time series of subtidal wind vectors and (bottom) time series of the along-shelf component of wind for several locations. Station locations and the amplitudes of the first and second wind EOFs are shown on the two inset maps.

Figure 14. Assumed vertical structure of along-shelf velocity and density gradient fields used to separate wind- and buoyancy-driven along-shelf velocities. Terms are defined in the text.
value of 0.04 cm s\(^{-1}\) has been used for conditions in which no gravity waves are felt on the bottom [Brink et al., 1987]. A number of runs were made with higher resistance coefficients (see section 6.2).

Figure 16 shows depth-averaged along-shelf currents obtained using (7) along with measured 5 m along-shelf currents at locations near the 55 m isobath that are frequently within the plume although well downstream of the plume turning region (W3, 5 m); south of the plume (O3, 5 m); and beneath the plume (S3, 41 m). Application of the model to other locations is deferred until section 6.2, where buoyancy-driven currents are considered explicitly. In spite of the fact that the measured currents may have baroclinic shear and that the 5 m currents likely also have a wind-driven Ekman component, the agreement between the modeled depth-averaged and observed currents at these midshelf sites at a variety of depths is remarkably good \((r \sim 0.8)\). Surprisingly, the highest correlation is observed at the site actually within the plume \((r = 0.9)\) rather than at sites beneath and outside the plume \((r = 0.8)\), where buoyancy-driven flow would be expected to be negligible. The only consistent pattern of deviation from modeled wind-driven flow appears to be an excess of southward flow observed at the plume site during periods of weak or southward wind stress.

The modeled wind-driven along-shelf flow predicts the timing of events with a high degree of accuracy at sites both within and outside the plume. Since (7) applies equally well to regions within the plume and to those outside the plume, where only wind-driven forcing occurs, the timing must be determined by the wind-driven forcing. For this to be the case within the

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Figure 15. Comparison of estimates of terms in the alongshelf momentum equation depth-averaged over the water column. Terms are described in the text. In each panel, the signs of terms on the right-hand side of (3) are reversed so that a mirror image between stress and other terms indicates a balance of momentum.

Figure 16. Time series of along-shelf depth-averaged velocity in a bottom depth of 55 m computed with a local wind-driven model \(^{(V_w)}\) compared with time series of observed along-shelf velocity at sites or depths that are predominantly within the plume (W3, 5 m), outside the plume (O3, 5 m), and beneath the plume (S3, 41 m). The model was run with a resistance coefficient of 0.05 cm s\(^{-1}\). In this and all subsequent similar figures, measured flow is shown with a solid line, modeled flow with a dashed line.
plume layer (e.g., at W3, 5 m), buoyancy-driven velocity must be either small or virtually phase locked with the depth-averaged wind-driven flow.

5.3. Cross-Shelf Flow
Modeling wind-driven frictional cross-shelf flow in an environment with time-variable stratification would require multiple assumptions about the vertical structure of both stratification and vertical eddy viscosity. A qualitative assessment of plume effects on the cross-shelf component of flow can be obtained by comparing the observed cross-shelf velocity with the along-shelf component of wind stress. At the 55 m site located most frequently outside the influence of the plume (O3), the cross-shelf velocity at 5 m is strongly correlated with the along-shelf wind stress, and the maximum correlation \( r = 0.7 \) occurs at zero lag (Figure 17, bottom). Intercept and slope values \( (0.94 \text{ cm s}^{-1} \text{ and } 5.2 \text{ cm}^3 \text{s}^{-1} \text{dyn}^{-1}) \) respectively) are similar to those obtained off northern California at 5 m in a bottom depth of 90 m \( (-0.5 \text{ cm s}^{-1} \text{ and } 4.2 \text{ cm}^3 \text{s}^{-1} \text{dyn}^{-1}) \) respectively [Winant et al., 1987]. The relationship between along-shelf wind and along-shelf velocity is reasonably uniform over time and is consistent with a predominance of simple frictionally driven flow at that site.

At a site that is frequently within the plume, on the other hand, a simple linear relationship between along-shelf wind stress and cross-shelf flow is not observed (Figure 17, top; \( r = 0.3 \), with wind lagging current by 0.25 day). Offshore velocity is several times higher than at the site outside the plume (typically 20–30 versus 5–10 cm s\(^{-1}\)). As mentioned during the discussion of Figure 8, large offshore velocities at plume sites appear to be related to the onset of relaxation of northward wind stress events. Velocities are onshore during the spin-up phase of a northward wind stress event. However, magnitudes are often 30–50% greater than those at the site outside the plume.

6. Geostrophic Buoyancy-Driven Velocity Field
Estimates of geostrophic buoyancy-driven velocity can be made from both CTD survey and moored array \( \sigma_r \) data. The former provide quasi-synoptic spatially comprehensive snapshots; the latter provide time series at individual sites.

6.1. Synoptic Maps
Dynamic height was calculated from shipboard survey data at selected depths relative to 15 dbar. This shallow reference layer, chosen so that stations on the inner shelf could be used, generally includes most of the river plume (see Figures 5a and 5b). The data were smoothed and interpolated to a uniform grid, and gradients were used to compute baroclinic velocity. Such a velocity field should be viewed with some skepticism, especially near the river mouth, where both tidal and nonlinear effects are likely significant.

Contoured dynamic height and the calculated baroclinic velocity fields are presented at the sea surface, 5 and 10 dbar for three of the CTD surveys (Plates 2a–2c). The dynamic height fields are colored to illustrate plume variability from survey to survey on a given pressure surface; i.e., the color scale is different for each pressure surface, but the same for all three examples. Results show that the plume produces a sea surface elevation signature of about 5–7 cm for both the northward and northwestern plume examples. Lack of data on the shallowest portions of the inner shelf likely leads to an underrep-
Plate 2a. Maps of geostrophic velocity (cm s\(^{-1}\)) and dynamic height (dyn m) relative to 15 dbar for January 20–22, 1991. Subtidal velocities measured in two depth ranges (20–30 m (solid lines) and 40–60 m (dashed lines)) beneath the plume, and at 5 and 10 m are shown in the upper, middle, and lower panels, respectively. Dots on the left-hand panel indicate the locations of CTD stations used to compute dynamic height. Velocities are shown for times closest to the time of the nearest CTD station pair. Note different scales for computed velocities and observed velocities. Color scales for dynamic height are different at each depth but are the same in Plates 2a–2c. Wind (W) and river flow (R) for the survey period are shown as scaled arrows on the left side of the left panel.

6.2. Time Series

To obtain estimates of depth-averaged flow within the plume, we make use of two assumptions introduced in section 5.2, namely, the flow is geostrophic in the along-shelf direction and the flow can be described by a vertically sheared plume layer with velocity \( v_p + v_d \) overlying a deeper, uniform flow layer with velocity \( v_d \) (see Figure 14). With these assumptions,
the along-shelf geostrophic balance at any depth \( z \) (2) can be expressed as

\[
\tau_V = \frac{\partial \eta}{\partial x} + \frac{\partial \rho}{\partial x} \int_0^z \frac{\partial \rho}{\partial x} \, dz'
\]  

(8)

where \( \eta \) is sea level height and \( g \) is the acceleration due to gravity. Averaging (8) vertically over the plume, assuming a constant plume thickness \( h \) and a constant plume density gradient \( \partial \rho / \partial x \), the equation for total flow depth-averaged over the plume \( \bar{v}_T \) becomes

\[
\bar{v}_T = g \frac{\partial \eta}{\partial x} + \frac{gh \, \partial \rho}{2 \rho \, \partial x}
\]  

(9)

At any depth beneath the plume, (8) can be expressed as

\[
\tau_v = g \frac{\partial \eta}{\partial x} + \frac{gh \, \partial \rho}{\rho \, \partial x}
\]  

(10)

Plate 2b. As in Plate 2a for October 25–26, 1990.
Subtracting (10) from (9) to eliminate the sea surface slope, we have

\[ f(\tilde{V}_p - v_d) = -\frac{gh}{2\rho} \frac{\partial \rho}{\partial x} \]  

(11)

However, since \( \tilde{V}_p = \tilde{v}_p + v_d \), then \( \tilde{v}_p \), the along-shelf depth-averaged geostrophic flow resulting from the plume pressure gradient field (the "buoyancy-driven" flow), can be expressed as

\[ \tilde{v}_p = -\frac{gh}{2\rho} \frac{\partial \rho}{\partial x} \]  

(12)

Using the fact that the fluctuating along-shelf wind-driven flow in this region is essentially barotropic in winter [e.g., Hickey, 1989], then \( v_d \) is equal to \( v_w \), the along-shelf wind-driven geostrophic flow that would occur in the absence of a plume. Thus the total flow depth-averaged over the plume is given by

\[ \tilde{V}_p = \tilde{v}_p + v_w \]  

(13)
Estimates of $\tilde{v}_p$ require a number of assumptions and should be regarded only as a lowest order approximation to the actual signal. Most of the limitations affect amplitude more strongly than phase. Rough estimates of the errors involved suggest that the estimated velocities are good to within about 30%. First, gradients in lateral frontal boundaries are clearly underestimated by the roughly 10 km scale of the moored array. This weakness is partially compensated by the use of 1 m $\sigma_t$ data to estimate the average density gradient over the entire plume. Also, plume thickness varies in space and in time as the plume moves over the shelf in response to the time-variable regional currents and as river outflow changes. Based on results from the CTD surveys, we use a plume thickness of 10 m.

Current observations at 5 m are used as a proxy for total flow depth-averaged over the plume ($\tilde{v}_p = \tilde{v}_p + v_p$). To estimate the magnitude of surface Ekman currents at 5 m, the along-shelf velocity difference was computed between 5 and 20 m at the midshelf location least affected by the plume (O3). If all the fluctuations at that site (but not the mean) are attributed to frictional shear, then an estimate of Ekman-related velocity differences in the along-shelf flow is 5–10 cm s$^{-1}$ for average to large wind stress events (Figure 18, lower left). This difference is more than a factor of 2 smaller than observed velocity differences over the same depth interval at a location where the plume occurs. Note that since velocity differences at O3 are not significantly correlated to along-shelf wind stress (not shown), this estimate is likely an upper bound.

To assess the skill of the model for geostrophic buoyancy-driven flow, we ask three questions: (1) how well does $v_p = v_p$ fit observed time series of along-shelf flow beneath the plume?, (2) how well does $\tilde{v}_p$ fit observed time series of along-shelf velocity differences between depths within and depths below the plume?, and (3) are the amplitude and phase of the model that includes buoyancy-driven currents improved over a model that includes wind-driven currents alone?

To address the first two questions, time series of $v_p$ are compared with observed below-plume (20 m) currents, and time series of $\tilde{v}_p$ are compared with observed plume (5 m) minus below-plume (20 m) velocity difference at midshelf (O3) and inner shelf (K1) (Figures 18 and 19). Results are remarkably good, confirming that the assumptions made were not unreasonable. For example, correlations between observed and modeled below-plume flow are significant at the 95% level ($r = 0.58$ at K3 and 0.61 at K1) and amplitudes of observed and modeled fluctuations are similar (Figure 18, upper right; Figure 19, lower right). Correlations are also significant between calculated ($\tilde{v}_p$) and measured plume to below-plume velocity differences (~0.8 at both sites), and amplitudes are similar (Figures 19 and 20, lower right).

To address the last question, lagged correlations and regressions were calculated between the observed along-shelf flow at 5 m at several plume sites (a proxy for total flow depth-averaged over the plume) and the purely wind-driven flow depth-averaged over the entire water column ($v_p$), buoyancy-driven flow depth-averaged over the plume ($\tilde{v}_p$), and total flow depth-averaged over the plume ($v_p + \tilde{v}_p$) for selected values of the resistance coefficient (Figures 21 and 22). Statistics are also presented for the EOF time series as well as for locations outside the plume (O3, 5 m) and beneath the plume (S3, 41 m). Statistics were calculated for the largest common time period of most records (52 days). Some records were shorter (O3, K1, and the EOFs, 35–40 days). All correlations exceeding 0.48 are significant at the 95% level, as estimated using decorrelation.
timescales calculated for this time period to determine the number of degrees of freedom.

Figures 21 and 22 show that both the phase lag and the slope of the regression between modeled time series and observations generally increase as resistance coefficient increases. The optimal model for each location was selected as the one for which the slope of the regression fell closest to unity (shaded values in Figures 21 and 22). In all cases that included a buoyancy-driven component, the phase lag of the prediction with respect to the observations was reasonable, either zero or a 6 hour lag. A slight observational lead is reasonable because the 5 m measured currents include contributions from surface frictional effects, which typically lead geostrophic currents by several hours (see discussion in section 5).

Results from the statistical analysis demonstrate that modeled time series which include a buoyancy-driven geostrophic component provide the best fit to the observations at sites within the plume (W3, W1, K3, K1, and the EOFs), whereas time series computed with the purely wind-driven model provide a good fit at sites outside or beneath the plume (O3, 5 m; S3, 41 m, respectively). Optimal fit requires resistance coefficients to increase as bottom depth decreases, consistent with the increase of gravity wave energy as the bottom shoals: 0.05–0.07 cm s⁻¹ at midshelf locations closest to the river mouth (W3 and K3) and 0.07–0.18 cm s⁻¹ on the inner shelf (W1 and K1, respectively).

At most sites, the model for which the slope of the regression is optimal includes significant nonzero intercepts ranging from a northward mean (~5 cm s⁻¹) at midshelf locations closest to the river mouth to a southward mean (approximately ~10 cm s⁻¹) at inner shelf locations north of the mouth (Figure 21). In contrast, at midshelf sites farthest from the river mouth (W3) and beneath the plume (S3, 41 m), the means required for optimal fit are smaller (<4 cm s⁻¹).

Examples of observed versus modeled time series at shallower midshelf locations show that plume and wind-driven along-shelf flows are nearly in phase, and both contribute significantly to the observed variance in along-shelf currents (of the order of 20–40 cm s⁻¹ for wind and 15–25 cm s⁻¹ for 80~---

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Figure 20. (top) Comparison of modeled and observed 5 m along-shelf currents at a midshelf site (W3) farther downstream from the estuary mouth than that shown in Figure 18. (bottom) Comparison of wind-driven (v_r) and buoyancy-driven (v_p) currents.
buoyancy) (Figures 18, 20, and 22). The signals are in phase because during a northward wind stress event when northward geostrophic as well as frictional wind-driven currents develop over the shelf, the surface frictional currents advect the density field of the existing plume toward the coast, where its associated geostrophic currents (usually northward) are also observed. The opposite occurs during southward wind stress events. At the inner shelf site farthest downstream (W1), both wind- and buoyancy-driven flows are similar to those at midshelf. On the inner shelf at the site closest to the river mouth (K1), on the other hand, wind- and buoyancy-driven flows are both smaller (of the order of 10-20 cm s$^{-1}$) with similar amplitudes (Figures 19 and 22). At the inner shelf site closest to the river mouth (K1), the sum of wind- and buoyancy-driven flow requires a larger resistance coefficient to fit the observations (the reason for the smaller wind-driven flow), and only a portion of the southward mean is reproduced by the model.

The sum of the first and second mode velocity EOFs reconstructed at a midshelf site (W3) at 5 m is also best represented by a model that includes geostrophic buoyancy-driven currents, consistent with the spatial structure of these EOFs as described in section 4.2 (Figure 10, lower right). Comparison between the first EOF (upper right) and the sum of the first two modes (lower right) shows that the second mode contributes very little to along-shelf velocity; i.e., the majority of along-shelf plume and wind-related variance are both contained in the first mode at this midshelf site downstream of the plume turning region. At a deeper midshelf location (K5), buoyancy-driven along-shelf geostrophic flow is also roughly equal to the magnitude of wind-driven flow (of the order of 15-25 cm s$^{-1}$), although the two are not strongly related (Figure 23). The sum of the two components underpredicts the northward mean as well as along-shelf flow during a number of northward events. Comparison of 1 m salinity with the difference between the observed along-shelf flow and the total modeled flow demonstrates that underprediction occurs when salinity is low (Figure 23, lower right). This occurs during periods of plume relaxation such as November 20 and 28-29, when this site is generally directly in the path of the plume (see Figures 8a and 8b). The poor model results at this site could be due to the existence of significant ageostrophic flow in this region; alternately, when a plume is present at this site, the stations used to compute the

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**Figure 21.** Intercept (upper left hand corners) and slope (lower right hand corners) for regressions between predictions ($v_T$, $v_P$, and $v_T + v_P$) and observations at selected sites. Symbols are defined in the text. With the exceptions of S3 and EOFs, all data are at 5 m. Predictions are given with various values of the resistance coefficient $r$ in cm s$^{-1}$. Values in each row for which the slope of the regression is closest to 1.0 are shaded.

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**Figure 22.** Maximum correlations between predictions and observations as shown in Figure 21. Lag at maximum correlation is given as a subscript, with each unit corresponding to 6 hours. A positive lag indicates that the observations lead the model. Correlations of 0.53 are significant at the 95% level [Koopmans, 1974]. Values shaded in Figure 21 are shaded here.
and buoyancy-driven contributions to the 5 m velocity vary with depth. In general, plumes appear to be more continuous at the sea surface than at greater depths. Plume thickness is roughly 5-15 m for plumes whose axes are separated from the shelf, although most of the plume volume is contained in the upper 10 m (separated plumes) or 20 m (northward plumes). Away from the river mouth, plumes are several tens of meters above the shelf, and within-plume stratification is generally weaker than across the shelf, at times well out over the shelf. However, the fact that the salinity and velocity fields appear related through advection makes it likely that at depths within the plume, nonlinear affects are significant even at some distance from the river mouth. The along-shelf pressure gradient usually fall on opposite sides of the plume axis, so that the local east-west pressure gradient is poorly resolved.

7. Discussion and Summary

The plume from the Columbia has two primary orientations during the fall/winter period: west to northwestward during periods of southward wind stress or light northward wind stress, and northward during periods of strong northward wind stress. The plume and its velocity field respond within hours to changes in wind speed and direction. Within-plume stratification of westward and northward plumes is strong, with vertical salinity gradients of up to 1 psu m⁻¹ in the upper 10 m. Northward plumes hug the coast, rarely extending beyond midshelf, and within-plume stratification is generally weaker than in northwestward plumes. Northwestward plumes extend across the shelf, at times well out over the slope at the sea surface. In the latter case, the area of the plume decreases dramatically with depth, rarely extending beyond the shelf edge at 5 m. The direction of these separated plumes can also vary with depth. In general, plumes appear to be more continuous at the sea surface than at greater depths. Plume thickness is roughly 5-15 m for plumes whose axes are separated from the coast and 10-40 m for plumes that hug the coast, although most of the plume volume is contained in the upper 10 m (separated plumes) or 20 m (northward plumes). Away from the river mouth, plumes are several tens of meters above the bottom except on the inner shelf during periods of reasonably strong northward wind stress.

Plume currents are relatively unidirectional within the low-salinity bulge that emanates from the river mouth. Currents in this portion of the plume are directed down the salinity gradient, and speeds are generally greatest in the plume turning region. Wind-driven frictional currents cause the flow along the plume axis to veer to the right (for northward winds) or to the left (for southward winds) of the plume axis. Farther downstream, currents turn to parallel, rather than cross, salinity contours, consistent with a geostrophic momentum balance. During events in which the plume axis is separated from the coast (deemed “relaxation” events), currents tend to flow around the mound of low-salinity water. Thus buoyancy-driven currents are directed northward or west-northwestward on the deeper part of the midshelf but southward or southeastward on the inner shelf.

At locations within the river plume but downstream of its immediate turning region, nonlinear contributions to the variance of the depth-averaged flow are negligible. Because of the relatively deep shelf and relatively strong wind forcing, the along-shelf flow depth-averaged over the water column is dominated by the difference between surface and bottom stress alone. However, the fact that the salinity and velocity fields appear to be related through advection makes it likely that at depths within the plume, nonlinear affects are significant even at some distance from the river mouth. The along-shelf pressure gradient and the Coriolis terms depth-averaged over the water column are greater than the depth-averaged nonlinear terms, although less than both wind stress and acceleration, and have a strong tendency to balance each other within the plume in a manner consistent with geostrophy.

Wind- and buoyancy-driven contributions to the 5 m velocity variance were successfully separated in regions where the depth-averaged flow was linear and along-shelf flow was geostrophic, using a linear model for wind-driven currents and buoyancy-driven currents calculated from the measured surface density field. The model successfully accounted for
roughly 70% of the variance in the observations of along-shelf velocity at depths and locations within the river plume (as judged statistically by the resulting amplitude and phase relationships). At locations outside or beneath the plume, on the other hand, optimal fit to the observations was obtained with a purely wind-driven barotropic model. Note that although the calculations made use of the fact that the depth-averaged flow is linear, the use of data (with the semigeostrophic assumption) to compute the plume flow implies that some nonlinear contributions to plume flow are included in the density-related estimates. A more detailed study of momentum balances for the Columbia plume is currently in progress.

Geostrophic along-shelf currents associated with the plume are of the order of 10−25 cm s−1 at 5 m and 30−50 cm s−1 at the sea surface. Superposition of wind- and buoyancy-driven along-shelf geostrophic flow in regions where both the semigeostrophic and linear depth-averaged flow assumptions are valid results in a signal that is northward in the mean (as are both wind- and buoyancy-driven means) but variable on subtidal scales. On the deeper part of the midshelf region, geostrophic buoyancy-driven along-shelf currents are often out of phase with wind-driven currents, and magnitudes are comparable (∼15−25 cm s−1). On the shallower midshelf and at the inner shelf location farthest from the river mouth, wind-driven along-shelf geostrophic currents at 5 m are larger than those of buoyancy-driven currents (∼20−40 versus ∼15−25 cm s−1). On the inner shelf closest to the river mouth, wind- and buoyancy-driven geostrophic currents are almost in phase and roughly comparable in magnitude at 5 m, although reduced in magnitude from those at other shelf sites (∼10−20 cm s−1). At the site closest to the river mouth, the calculated buoyancy-driven flow successfully accounts for a portion of the strong southward currents observed during plume relaxation events. However, in general, the calculated geostrophic flow underpredicts a portion of the northward mean flow at midshelf and the southward mean on the inner shelf.

The spatial structure of the Columbia plume and its associated velocity field are consistent with general features predicted by models [e.g., Chao and Boicourt, 1986; Garvine, 1987; Kourafalou et al., 1996a]. For example, the data demonstrate a bulge offshore of the river mouth and the existence of a higher speed turning region. However, reattachment to the coast and development of a geostrophic coastal current are observed only under conditions of relatively strong wind stress in the direction of the rotational tendency. In model studies, reattachment occurs in conditions with no coastal winds. One difficulty in comparing model results to observations lies in the fact that the timescale for development of a downstream geostrophic coastal current in models is of the order of 10 days, whereas winds over the Columbia plume in fall and winter change on much shorter timescales.

Geostrophic buoyancy-driven currents estimated for the Columbia plume are stronger than those deduced from most models, which have been developed for weaker outflows (e.g., Kourafalou et al. [1996a, b] for rivers in the South Atlantic Bight; Chao and Boicourt [1986] for the plume from Chesapeake Bay). Unlike those weaker plumes, the magnitude of buoyancy-driven along-shelf currents in the Columbia plume is comparable to that of wind-driven along-shelf currents.

Models predict enhancement of surface Ekman flow within the plume [Kourafalou et al., 1996a]. Cross-isobath currents in the upper 5 m of the Columbia plume frequently exceed those outside the plume by up to a factor of 2. However, in contrast to regions outside the plume, these currents are not well correlated with along-shelf wind stress and therefore cannot be attributed to simple amplitude enhancement of wind-driven Ekman currents by the plume. During the spin-up phase of northward wind events, the onshore currents within the plume are in roughly the same direction as those outside the plume, although larger in magnitude. However, when the wind begins to relax (although still northward), 5 m plume currents turn offshore, while those outside the plume remain onshore.

Weak southward flow inshore of plumes has been observed in model results [e.g., Kourafalou et al., 1996a] and is usually attributed to frictional effects; i.e., along-shelf flow responds first on the inner shelf to shifts in wind direction. Although this is also the case for the Columbia region (not shown), the southward flow commonly observed inshore of the Columbia plume is more consistent with the existence of geostrophic flow around the plume. In support of this hypothesis, buoyancy-driven along-shelf geostrophic flow calculated at stations closest to the river mouth is more strongly southward than that calculated farther downstream from the mouth (compare $\gamma_p$ at K1 and W3; Figures 19 and 20). Several examples demonstrate that this southward flow often occurs prior to wind reversal. Moreover, the wind-driven model consistently underpredicts the magnitude of southward flow during relaxation events, suggesting that the observed southward flow is not due only to cross-shelf differences in bottom friction.

No evidence for counterflow beneath the plume was observed within the limitations of the Columbia Plume Study subtidal data set. An undercurrent is predicted by models beneath the bulge near the river mouth [Chao and Boicourt, 1986]. Estuarine inflow must occur near the mouth of the estuary, so it seems likely that the Columbia moored array did not sample the bulge region with sufficient resolution to detect undercurrents. The Columbia plume appears to have little effect on currents underneath the plume except at locations within about 5 km of the river mouth. Currents below 30 m were relatively depth independent, with the exception of flow reduction due to bottom Ekman layer effects.

Results from the Columbia Plume Study differ significantly from those of the recent study of the Amazon river (AMASEDS). Because of the low latitude of the Amazon plume, the Coriolis term in the cross-plume momentum balance is of the order of the horizontal wind stress and local acceleration terms. In the case of the midlatitude Columbia plume, cross-plume dynamics in the region downstream of the river mouth appear to be close to a geostrophic balance. One of the most striking differences between the two plumes is that wind-driven flow in the Amazon plume is completely decoupled from a lower layer [Lenz, 1995]; in the Columbia plume, wind driving controls the depth-averaged along-shelf flow and contributes roughly half the along-shelf variance to the total flow within the plume and almost all the variance below the plume. Analyses in the two studies are performed on data from similar water column depths and in similar bottom depths. Both the Columbia and Amazon plumes are usually detached from the bottom; density contrasts (plume to ambient water) are similar; the plumes are about the same thickness when offshore of the coast. The difference in coupling between the plume and below-plume layers may be partly because the Amazon study took place 100 km from the coast, where wind-driven coastal upwelling/downwelling, which drives the strong along-shelf quasi-barotropic currents off the Washington coast, is not a dominant process.
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