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Coastal Upwelling in the California Current System

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Abstract–Coastal upwelling in the California Current system has been the subject of large scale studies off California and Baja California, and of small scale studies off Oregon. Recent studies of the winds along the entire coast from 25° N to 50° N indicate that there are significant alongshore variations in the strength of coastal upwelling, which are reflected in the observed temperature distribution. Active upwelling appears to be restricted to a narrow coastal band (about 10–25 km wide) along the entire coast, but the region influenced by coastal upwelling may be much wider. Intensive observations of the upwelling zone during summer off Oregon show the presence of a southward coastal jet at the surface, a mean vertical shear, a poleward undercurrent along the bottom, and persistently sloping isopycnals over the continental shelf; most of the upwelling there occurs during relatively short periods (several days long) of upwelling-favorable winds. During the upwelling season off Oregon, the offshore Ekman transport is carried by the surface Ekman layer, and the onshore return flow occurs through a quasi-geostrophic interior. It is not known whether the structure and dynamics observed off Oregon are typical of the upwelling zone along the entire coast, though some of the same features have been observed off Baja California. Current and future research will eventually show whether the Oregon results are also applicable in the region of persistently strong upwelling-favorable winds off northern California, and in the region of complex bathymetry off central and southern California.

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1. INTRODUCTION

COASTAL upwelling has long been known to occur along the entire west coast of the United States and Baja California. This upwelling of subsurface waters is the result of the offshore Ekman transport of surface waters away from the coast because of equatorward winds along the coast. Large-scale systematic studies of upwelling and its consequences in the California Current system were initiated by the California Cooperative Fisheries Investigations (CalCOFI) program in 1949. Since 1950 the CalCOFI studies have been limited mainly to the region south of San Francisco, with the greatest efforts concentrated south of Pt. Conception. Small-scale studies of the upwelling regime off Oregon began about 1960 with repeated hydrographic sections off Newport; these were reinforced with sporadic current meter observations beginning in 1965, and culminated in the intensive Coastal Upwelling Experiments, CUE-I and CUE-II, conducted during the summers of 1972 and 1973. The wind-driven circulation over the continental shelf along the west coast continues to be a subject of active research, and we will therefore continue to learn about coastal upwelling in the California Current system.

In this paper, I shall first summarize what we know about this upwelling region as a whole; then discuss coastal upwelling off central Oregon where it has been studied intensively; summarize briefly what is known about coastal upwelling off southern California and Baja California; and finally discuss very briefly some areas that might be covered by current or future research.

2. THE LARGE-SCALE PICTURE

2.1. *The forcing function*

Upwelling along the west coast of North America is forced by atmospheric circulation around the North Pacific High which varies seasonally in both strength and position. The High migrates from its most southerly position at 28°N, 130°W in February to its most northerly position at 38°N, 150°W in August and back again (ANON., 1961); the maximum pressure increases from winter to summer (~ 1020 mb in January to ~ 1025 mb in July) (Fig. 1). The atmospheric pressure over the extreme western United States is much higher in winter (~ 1020 mb in January) when the surface is cold, than it is in summer (~ 1015 mb in July) when the surface is warm (Fig. 1). This thermal effect on pressure is particularly marked in California's central valley, where the thermal Low reaches 1005 mb in July. Thus, the offshore pressure gradient along the west coast is strongest in summer. The strength of the pressure gradient across coastal waters varies with position along the coast, with season, and of course also with day-to-day variations in the weather.

Charts of mean monthly wind stress over the California Current based on ship reports (Fig. 2) show that the region of maximum equatorward wind stress migrates northward from ~ 25°N in January to ~ 39°N in July as expected from the pressure distribution (Fig. 1). The strongest equatorward winds are observed in July off northern California, where the center of the High is closest to the thermal Low (Fig. 1). South of about 40°N, winds are equatorward all year; north of this latitude, the wind stress is northward, i.e., unfavorable for coastal upwelling, in winter.

The alongshore wind stress, τ, causes a net offshore transport in the surface Ekman layer, τ/f, which in turn is equal to the amount of water that upwells along the coast. The Coriolis parameter, f, increases from 0.6×10^{-4} sec^{-1} at 24°N to 1.1×10^{-4} sec^{-1} at 50°N; thus favorable wind of the same magnitude will result in nearly twice as much upwelling off southern Baja California as off northern Washington. Using measured winds contained in ship reports, NELSON (1977) computed long-term mean wind stress for one-degree squares adjacent to the coast. Ekman transports computed from these data (Fig. 3) show that coastal upwelling occurs year-round as far north as San Francisco. South of San Francisco, the upwelling is strongest in April, but north of San Francisco the upwelling is strongest in July. The seasonal range is smallest in the Southern California Bight, and greatest off northern California. Coastal upwelling is relatively weak off central Oregon where the Coastal Upwelling Experiments (CUE-I and CUE-II) took place; it is much stronger in the vicinity of Pt. Arena, the site of the small-scale Coastal Dynamics Experiment (CODE) which is currently in progress. In the region covered by

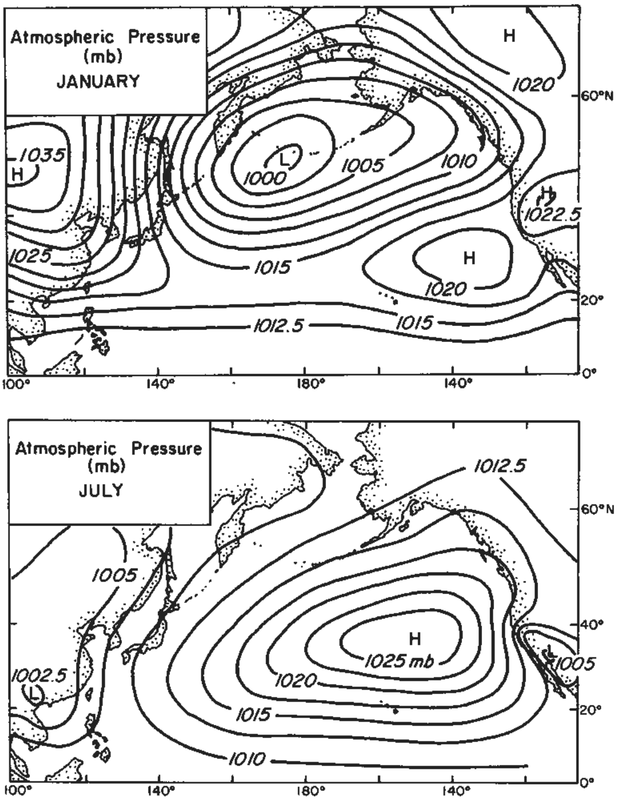

FIG. 1. Long-term mean atmospheric pressure at sea level for January and July (ANON., 1961).

the long-term, large-scale CalCOFI studies (24° to 38°N), upwelling is nearly independent of season, but strongly dependent on location.

Short-term fluctuations with periods of several days in the wind stress can cause upwelling, even when the mean wind stress for a given month is not favorable. Such "upwelling events" (e.g. HALPERN, 1976) are of considerable importance to the upwelling regime off Oregon. To determine whether the importance of these fluctuations varies along the coast we compared means and standard deviations of BAKUN's (1975) daily upwelling indices, which are offshore Ekman transports computed from the 6-hourly synoptic atmospheric pressure on a 3-degree grid. Although these indices overestimate the offshore Ekman transports along southern California, apparently because the coastal mountain range effectively isolates the thermal Low (BAKUN, 1973), this error presumably affects computations of daily values and longer-term means in the same way; if so, it would not affect the ratio between the mean and the standard deviation for each month. Thus, this ratio is a measure (albeit a crude one) of the relative importance of upwelling events at each latitude. Where the ratio is greater than one, the standard deviation of daily values is less than the monthly mean; the larger the ratio, the fewer periods of unfavorable wind would be expected. In July, the mean upwelling winds prevail along almost the entire coast (Fig. 4), unfavorable events seem to be important only in the extreme south (at 24°N) and the extreme north (at 48°N and 51°N). In April, the mean upwelling-favorable winds dominate along the southern coast up to San Francisco, but events dominate

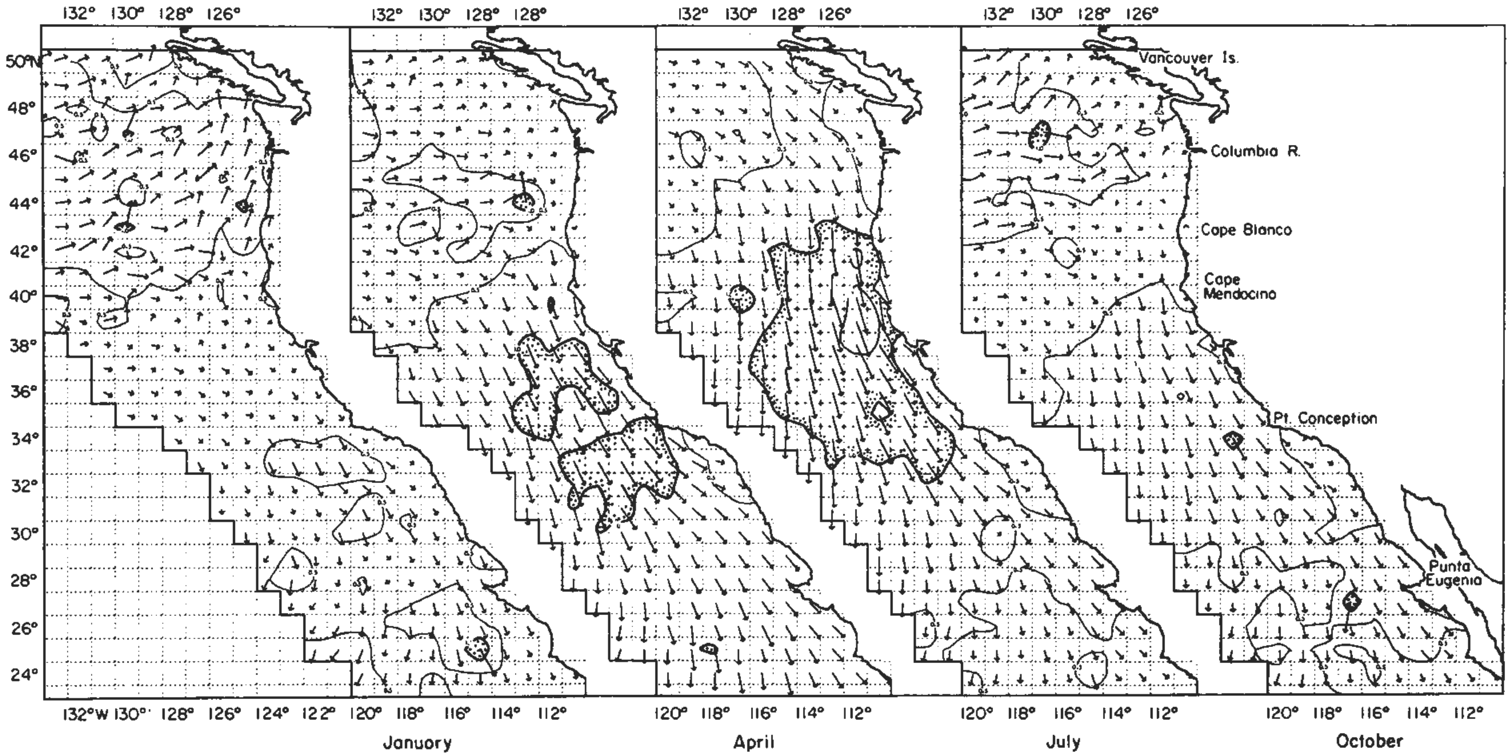

FIG. 2. Long-term (1854–1972) mean wind stress (dynes cm^{2}) by one-degree squares, from ship reports (NELSON, 1977). Contours correspond to constant magnitudes of 0.5, 1.0 and 1.5 dynes cm^{-2}; areas where the wind stress exceeds 1.0 dynes cm^{-2} are shaded.

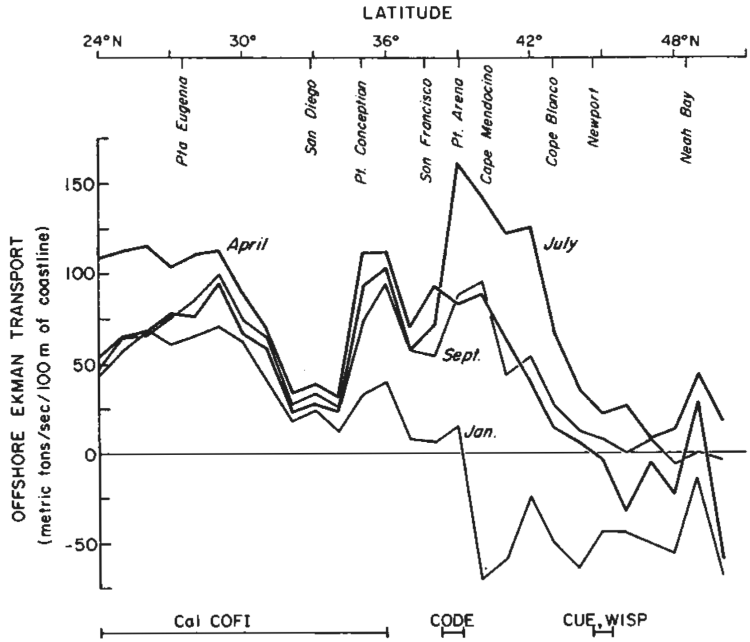

FIG. 3. Offshore Ekman transport computed from NELSON's (1977) long-term mean wind stress data for one-degree squares adjacent to the coast, for January, April, July and September. Bars indicate the alongshore extent of major upwelling studies.

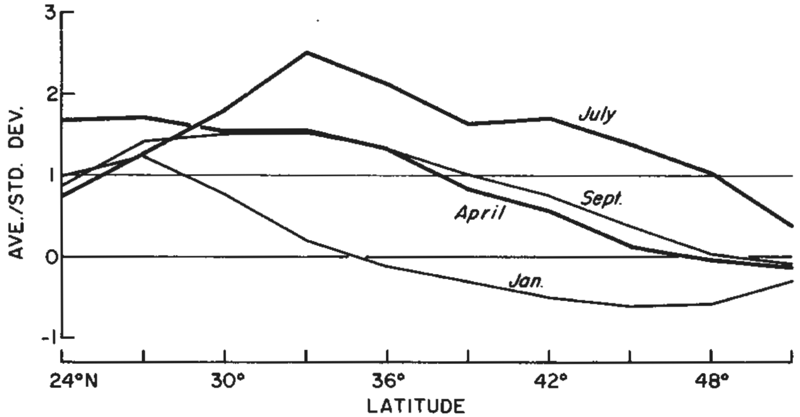

FIG. 4. The ratio of averages to standard deviations of daily means of six-hourly upwelling indices, for selected months, obtained from BAKUN's (1975) Tables 3 and 5. Values less than one indicate that short-term fluctuations have an amplitude greater than the mean.

off Oregon and Washington. In January, the ratio is about -0.5 off Oregon and Washington; this indicates that there are upwelling-favorable events there at this time of year, even though the mean wind stress has a northward component.

Open ocean upwelling can occur away from the coastal boundary if there is positive curl in the wind stress field, i.e., if there is net divergence in the Ekman transport field (SMITH, 1968).

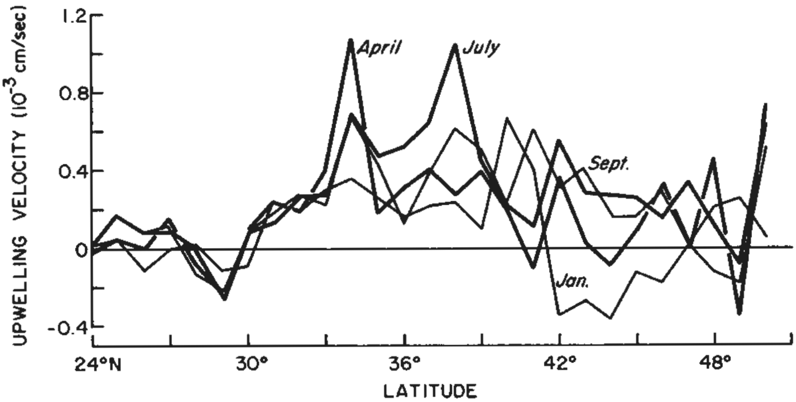

FIG. 5. The vertical upwelling velocity, *w*, due to curl of the mean wind stress field, for one-degree squares adjacent to the coast, calculated from NELSON's (1977, p. 19) wind stress curl data.

For example, if offshore winds are more strongly southward than coastal winds, there is stronger seaward Ekman transport offshore than near the coast, and there must be upwelling away from the coast as well as along the coast. Using finite-differences between one-degree squares, NELSON (1977) found persistent positive curl between 30° and 40°N (Fig. 5): south of 30°N, there is almost no wind stress curl along the coast, and hence the upwelling there is primarily coastal; between 30°N and 40°N, curl-induced upwelling reinforces the effects of coastal upwelling throughout the year. From 40°N to 48°N, the curl is favorable for upwelling in summer but not in winter, i.e., the curl-induced upwelling reinforces the seasonal coastal upwelling. Since NELSON's (1977) wind data has already been spatially averaged into one-degree squares, his estimates of wind stress curl and the associated vertical velocities are very small ($w < 10^{-3}$ cm $\sec^{-1}$). Gradients in the actual wind field are likely to be very much stronger near shore, particularly near large headlands such as Cape Mendocino, and along steep coastal mountain ranges. Unfortunately, wind data suitable for measuring the curl of the wind stress over near-shore waters are not generally available.

2.2. *Stratification*

The near-surface waters within the California Current are largely Subartic in character (TIBBY, 1941; REID, RODEN and WYLLIE, 1958): salinity generally increases with depth and there is a permanent halocline at a depth of about 150 m offshore. Temperature inversions are common (RODEN, 1964). There is a definite alongshore gradient in the water properties along the coast, with the percentage of Subartic water and the strength of the permanent halocline decreasing gradually from north to south (TIBBY, 1941; FLEMING, 1958).

The near surface stratification is seasonally influenced by river discharge. In winter, coastal currents are generally northward (HICKEY, 1979) and the effluent of rivers and coastal streams is concentrated near shore. In summer, there is virtually no discharge from coastal streams, negligible discharge from the rivers of California and southern Oregon (Fig. 6), and only two rivers dominate: the Columbia, which discharges directly into the Pacific at 46°15′N, and the Fraser, which discharges through the Straits of Georgia and Juan de Fuca into the Pacific at 48°30′N. Discharge from the Fraser is quite well mixed with deep salty water by strong tidal

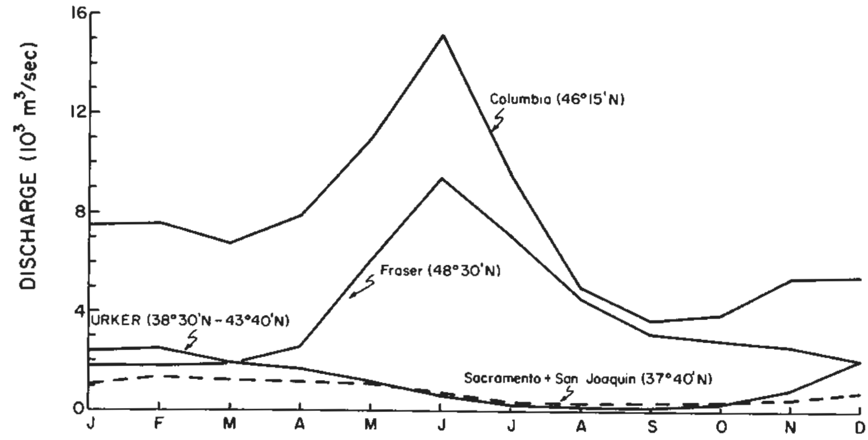

FIG. 6. Mean monthly streamflow data for major rivers discharging at different latitudes along the west coast. Data for the Umpqua (43°40'N), Rogue (42°30'N), Klamath (41°30'N), Eel (40°40'N) and Russian (30°30'N) were combined in the curve labelled URKER. (From ANON., 1978; BOURKE *et al.*, 1971; and JORGENSEN *et al.*, 1971).

currents within the straits before it enters the Pacific (HERLINVEAUX and TULLY, 1961), so its outflow is not readily distinguished from the surrounding coastal waters. The outflow of the Columbia, however, is still of very low salinity (< 20‰) when it reaches the sea. If the total April-through-August Columbia discharge were spread uniformly in a layer 1 m thick, it would cover an area of 129,000 km^2, say 250 km wide and 500 km long. If the typical speed of the southward coastal current were 10 km/day (half the value observed off central Oregon) a parcel of water leaving the Columbia on 1 April would be off San Diego before 31 August. The fresh water is diffused both horizontally and vertically after it enters the ocean, and the effluent takes the form of a shallow plume of low salinity water which extends offshore and southward in summer. The equivalent depth of fresh water calculated from the salinity distribution observed in July 1961 (BARNES, DUXBURY and MORSE, 1972) shows maximum values of 2 m just off the mouth of the Columbia and values of 0.5 m off Cape Blanco, more than 300 km to the south. Off central and northern Oregon, the Columbia River plume often lies very near shore [Fig. 7(a, b)] so that its inshore boundary coincides with the offshore boundary of the upwelled water. Figure 7(c) shows evidence (e.g. the 32.4 and 32.6‰ isohalines) that the plume extends beyond Cape Mendocino, and may still be discernible past San Francisco. Even off northern California, it may be difficult to distinguish between the "upwelling front" and the inshore boundary of the plume.

Except in the shallow (< 30 m) surface layer, the stratification does not vary much along the coast. Temperature and salinity both increase southwards – but the density is relatively constant, and so is its vertical gradient. The Rossby radius of deformation λ is a fundamental length scale of coastal upwelling in a stratified fluid; it can be estimated roughly using $\lambda = H\bar{N}/f$ (MONIN, KAMESKOVITCH and KURT, 1977, p. 152), where H is the water depth, f is the Coriolis parameter, and $\bar{N}$ is the vertically-averaged Brunt-Väisälä frequency which depends on the vertical density gradient. From hydrographic sections made across the shelf and slope at seven locations between 34°45'N and 43°20'N in February 1981 (FLEISCHBEIN *et al.*, 1981), we estimated the Rossby radius to be about 10 km over the mid-shelf, and about 40 km over the upper slope (near the 800 m isobath) at each latitude. Farther from shore, in water about

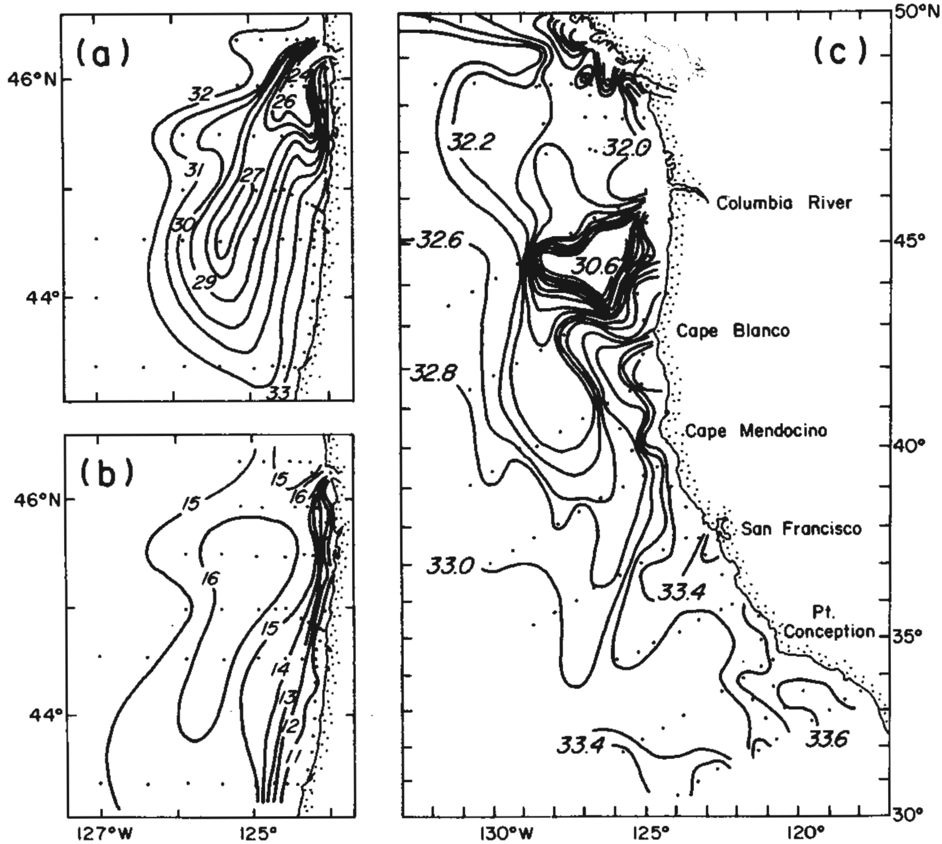

FIG. 7. Distributions of temperature and salinity showing the influence of the Columbia River plume: (a) surface salinity and (b) surface temperature off Oregon, 20 June–3 July 1968 (from PAK, BEARDSLEY and PARK, 1970); (c) salinity at 10 m off Washington, Oregon and California, 5–19 August 1950 (ANON., 1963).

3500 m deep, we estimated the Rossby radius to be about 80 km. More rigorous methods of computing the Rossby radius from the phase speed of the first dynamic modes are even more sensitive to total water depth (R. ROMEA, *pers. comm.*). Thus, the Rossby radius in this region varies more with bottom depth than with latitude.

2.3. *The consequences of upwelling*

Figure 8 shows maps of sea surface temperature for the California Current region for different months: relatively cold water adjacent to the coast is a likely consequence of coastal upwelling. In general, these maps show cold water along the coast wherever the mean monthly offshore Ekman transport is large: off northern Baja California in January; from Cape Mendocino to Pt. Conception and along most of Baja California in April; along the entire coast except the Southern California Bight and near the mouth of the Columbia in July; and from Cape Blanco to Pt. Conception and along northern Baja California in September. The major exception is the absence of cold coastal water along southern Baja California, in spite of significant offshore Ekman transport there throughout the year (Fig. 3). BAKUN and NELSON

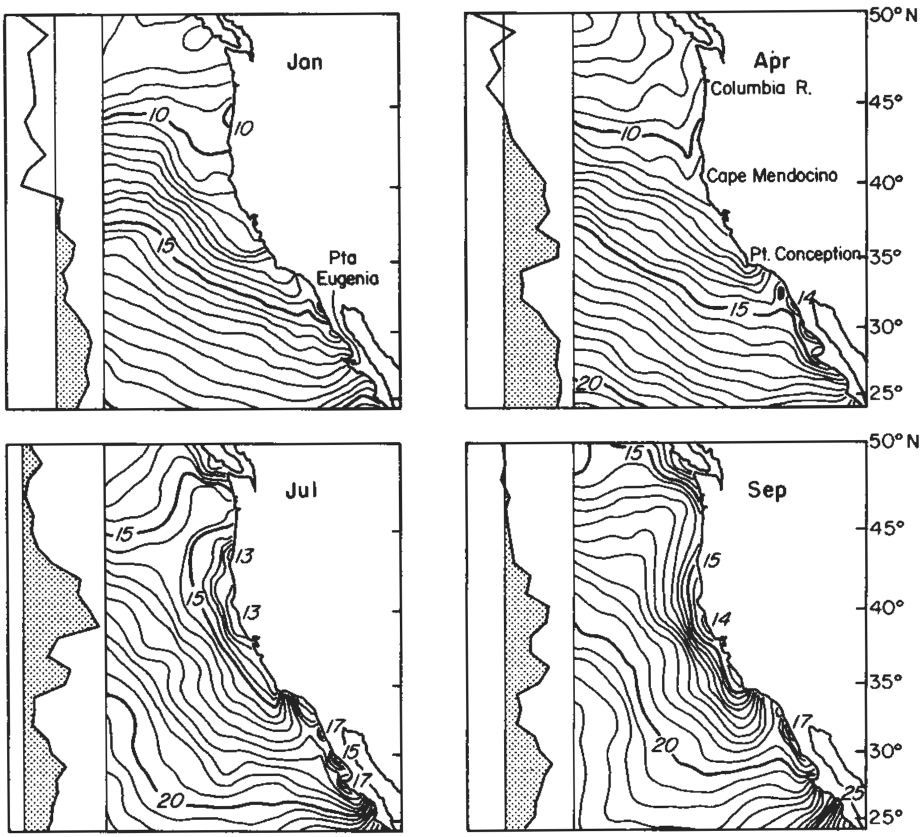

FIG. 8. Maps of mean monthly sea surface temperature in the California Current Region, 24° N to 50° N and 110°W to 135°W (from ROBINSON, 1976). Shown with each map is the corresponding profile of the Ekman transport from Fig. 3 (offshore transport is shaded).

(1977) note that the very low cloud cover south of Pta. Eugenia (28°N) would lead to rapid surface heating along southern Baja California, reducing the effect of upwelling on sea surface temperature there. More detailed maps (LYNN, 1967) do show the presence of a narrow (< 50 km) band of relatively cold coastal water south of Pta. Eugenia from April through July. Similarly, the absence of cold water off the Columbia River in summer is explained as the result of rapid warming of the thin Columbia River plume (BAKUN, MCLAIN and MAYO, 1974); again, more detailed maps [e.g. Fig. 7(b)] show a very narrow band of cold water is present along northern Oregon in summer. The September map (Fig. 8) shows cold coastal waters along Washington and northern Oregon, even though the monthly offshore Ekman transport seems negligible there; this is probably the result of shorter period upwelling events.

With nearly constant stratification along the coast, we might expect the depth from which water upwells to be greatest where the wind stress is strongest. If we again take relatively cold water adjacent to the coast as an indication of coastal upwelling, it appears that water upwells from depths greater than 30 m along the entire coast (Fig. 9) and from depths greater than 60 m between Cape Mendocino and Pt. Conception and along Baja California. Water appears to upwell from depths exceeding 90 m near Pt. Arena (39°N), Pt. Conception (35°N), and

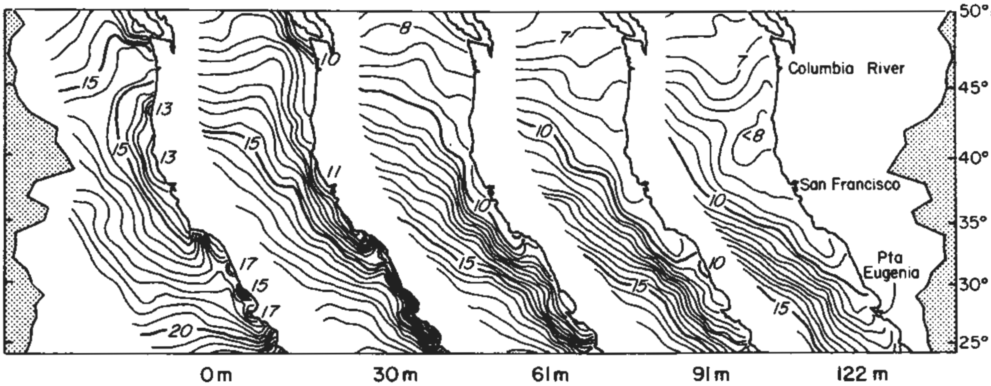

FIG. 9. Maps of mean July temperature at different depths (from ROBINSON, 1976). The meridional profile of the offshore Ekman transport is shown on both sides of the figure.

along northern Baja California. Thus there does seem to be some correspondence between the depth from which water upwells and the magnitude of the offshore Ekman transport.

Both Figs 8 and 9 suggest that the effects of coastal upwelling can extend more than a hundred kilometers out to sea. Simple theories of time-dependent coastal upwelling (e.g., ALLEN, 1973) indicate that the width of the upwelling zone is given by the Rossby radius of deformation, which is about 10–20 km over the shelf along the entire coast. Where strong upwelling persists, water properties associated with upwelling can be carried beyond this narrow upwelling zone by advection and eddy diffusion (DE SZOEKE and RICHMAN, 1981). The zone influenced by upwelling seems to be widest where the offshore Ekman transport is greatest: it may be narrowest off northern Oregon where the Columbia River plume provides an offshore limit.

Coastal upwelling causes a redistribution in the density field as lighter surface water is replaced along the coast by heavier, originally subsurface, water. Hence, upwelling can affect the pressure field, the sea surface topography, and the velocity field. Off Washington, Oregon and northern California, coastal sea level is significantly correlated with alongshore wind stress, on both the several-day time scale (OSMER and HUYER, 1978) and on the monthly anomaly time scale (ENFIELD and ALLEN, 1980); upwelling-favorable winds are associated with low sea level. Coastal sea level adjusted for the "inverted barometer effect" agrees well with near-shore steric height, on both several-day (LAFOND, 1939; HUYER, SOBEY and SMITH, 1979) and longer time scales (REID and MANTYLA, 1976; CHELTON, 1980). Sea level is also correlated with the strength of coastal currents on time scales longer than a day. Although the exact sequence and the dynamics are not fully understood, it is clear that coastal upwelling affects the density field and hence the steric height and the alongshore currents in the northern part of the California Current system.

3. UPWELLING OFF OREGON

Coastal upwelling off Oregon is a seasonal phenomenon, with significant but relatively weak mean offshore Ekman transport in summer and mean onshore Ekman transport in winter (Fig. 3). The wind stress is highly variable throughout the year, and upwelling-favorable events of several days duration occur even in winter when the mean winds are unfavorable; such events can be particularly important in spring and fall when the mean wind stress is near zero. This

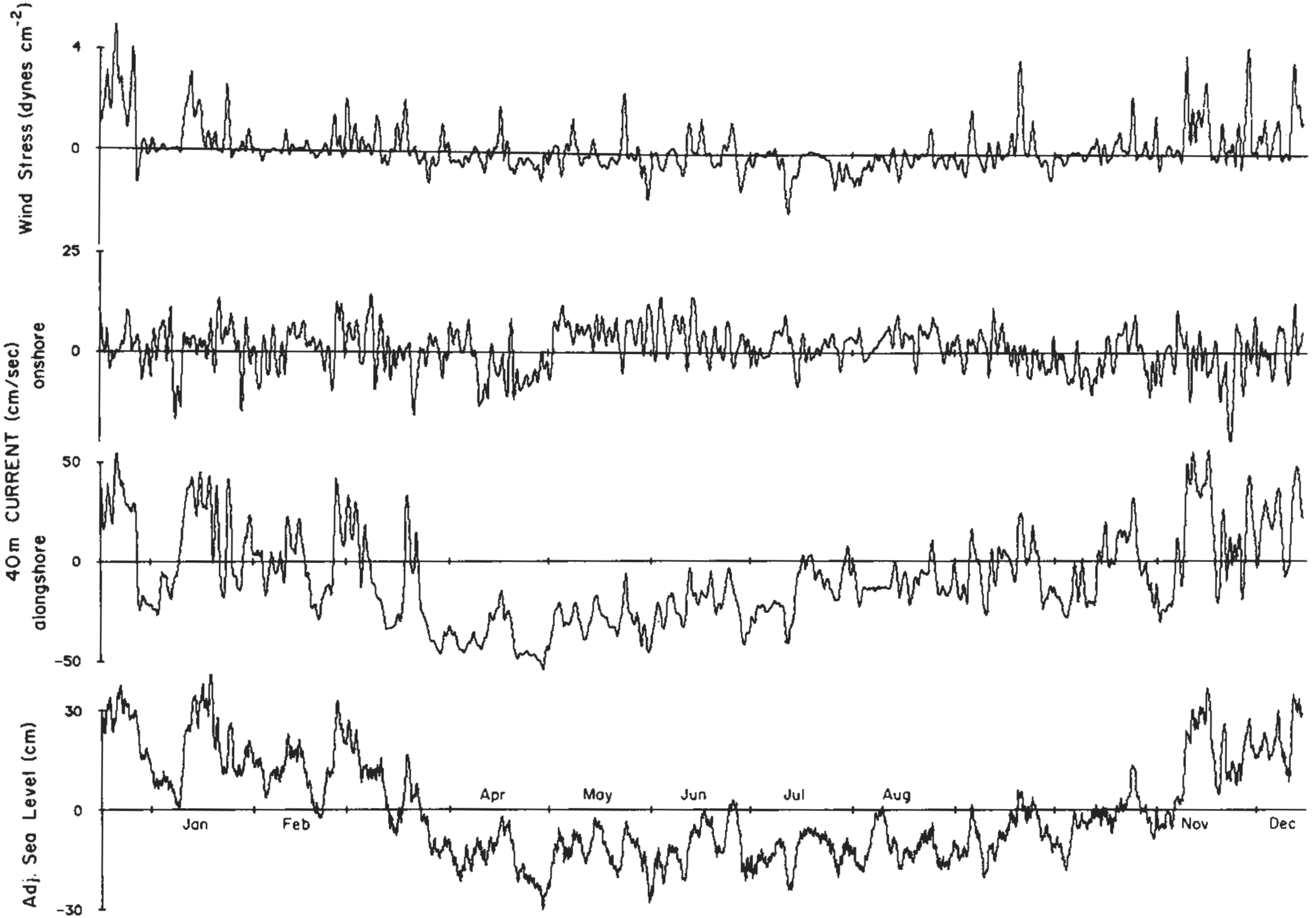

FIG. 10. Low-passed ($f < 0.6$ cpd) time series of northward wind stress at Newport, onshore (toward 101.6°T) and alongshore (toward 21.6°T) current at 40 m at Poinsettia (44°45′N, 124°17′W), and adjusted sea level at Newport, 17 December 1972 to 12 December 1973.

time variability is reflected in the onshore and alongshore components of the current and in coastal sea level (Fig. 10), and in the hydrographic regime. There is also variability on even shorter time scales: the diurnal sea breeze in the wind; diurnal and semi-diurnal tidal and inertial currents; and semi-diurnal and shorter period oscillations in the density field. As we shall see below, processes occurring at the different frequencies are not necessarily independent, i.e., some longer-term effects appear to be the result of shorter-period forcing.

3.1. *Summer*

The intensive Coastal Upwelling Experiments (CUE-I and CUE-II) conducted during the summers of 1972 and 1973 have resulted in a detailed understanding of the summer upwelling regime off central Oregon. These studies were largely restricted to the continental shelf which is about 30 km wide, and generally between 50 and 200 m deep. Schematics of the structure of this upwelling regime are shown in Fig. 11. The surface Ekman layer, in which the wind-driven offshore transport occurs, is less than 20 m deep (HALPERN, 1976; SMITH, 1981); it contains the near-shore edge of the Columbia River plume, which is advected back and forth with fluctuations in the wind stress (HUYER, SMITH and PILLSBURY, 1974). The bottom Ekman layer is about 10–15 m thick (KUNDU, 1976): thus, except very near shore, most of the water column is occupied by a "geostrophic interior" (KUNDU, 1977), in which friction is negligible and in which the slowly-varying currents are very nearly balanced by varying pressure gradients.

The density field off Oregon has a shallow seasonal pycnocline associated with the Columbia River plume, and a deeper permanent pycnocline (σ_t: 25.5–26.0) associated with the permanent halocline (32.5–33.8‰) that is ubiquitous in the eastern Subartic Pacific. In summer, isopycnals from depths as great as 150 m slope upwards toward the coast and intersect the surface (Fig. 11): maximum horizontal density gradients ("fronts") occur where the pycnoclines

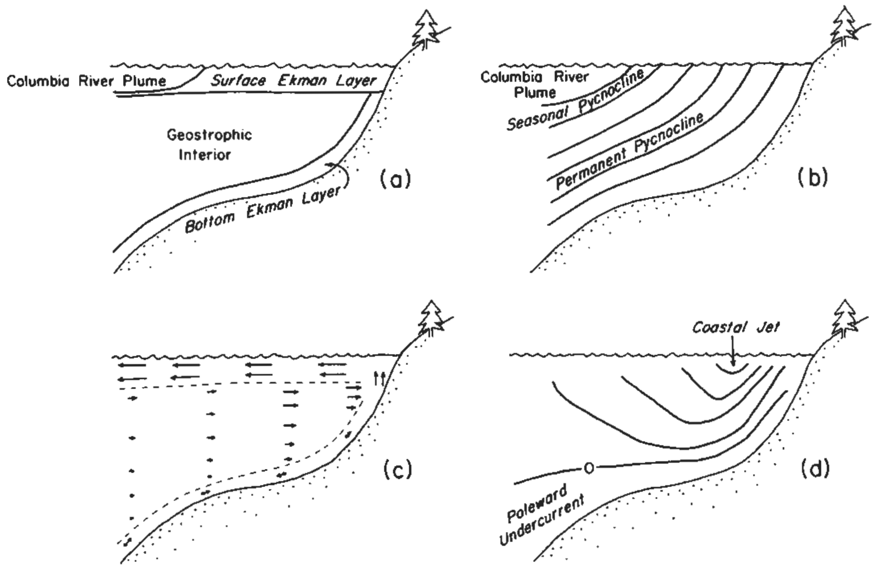

FIG. 11. Schematics of the coastal upwelling regime over the Oregon shelf at the height of the upwelling season. (a) Kinematic zones (b) density structure (c) onshore/offshore circulation (d) alongshore currents.

intersect the surface. During strong upwelling, the Columbia River plume is advected far offshore, and the "upwelling front" corresponds to the permanent pycnocline. At other times, the upwelling front and the edge of the plume are indistinguishable. The density structure within the surface Ekman layer (the upper 20 m or so) changes rapidly as a result of changes in the wind stress (HUYER, SMITH and PILLSBURY, 1974; HOLLADAY and O'BRIEN, 1975) but the upward sloping isopycnals in the "geostrophic interior" persist through the upwelling season (HUYER, 1977).

The dominant features of the alongshore velocity field (Fig. 11) are a southward "coastal jet" at the surface and a poleward undercurrent along the bottom (MOOERS, COLLINS and SMITH, 1976; HUYER, PILLSBURY and SMITH, 1975a). Vertical gradients of the alongshore currents are in approximate geostrophic balance with the offshore density gradients: there is strong vertical shear in the sense that near-surface currents are more strongly southward than deeper currents. The southward coastal jet has its maximum (~ 25 cm sec^{-1}) about 15–20 km from shore (KUNDU and ALLEN, 1976) at the upwelling front where there is a maximum in the offshore density gradient. The vertical shear is relatively constant ($\sim 4 \times 10^{-3}$ sec^{-1}) during the upwelling season in spite of significant fluctuations in the wind stress (SMITH, 1974; KUNDU, ALLEN and SMITH, 1975). The poleward undercurrent has a velocity of about 5 cm sec^{-1} (HUYER, PILLSBURY and SMITH, 1975a). Superimposed on this pattern are fluctuations that are very nearly barotropic (SMITH, 1974), are coherent over alongshore distances of 200 km or more (HUYER *et al.*, 1975b) and with local sea level and wind stress (SMITH, 1974), and have an offshore decay scale of about 15–20 km (KUNDU and ALLEN, 1975; HUYER, SMITH and SOBEY, 1978).

The onshore–offshore circulation is relatively weak and therefore more difficult to measure and analyze. Because of its potential importance to the nutrient and plankton distributions, conceptual models of the upwelling circulation were developed even before the intensive CUE experiments. On the basis of 25 hr mean current profiles obtained at several locations in September 1966 (COLLINS *et al.*, 1968) and a persistent temperature inversion near the upward-sloping pycnocline, MOOERS (1970) postulated a two-celled circulation pattern with offshore flow both at the surface and along the bottom of the permanent pycnocline (SMITH, MOOERS and ENFIELD, 1971; MOOERS, COLLINS and SMITH, 1976). This model seemed to be consistent with the distributions of optical parameters (PAK, BEARDSLEY and SMITH, 1970) and phytoplankton (SMALL and MENZIES, 1981).

Sparse observations from vertically profiling current meters also seemed to support the two-celled circulation pattern (JOHNSON, VAN LEER and MOOERS, 1976; JOHNSON, 1977; JOHNSON and JOHNSON, 1979). However, these observations were of short duration (one 64 hr period in August 1973 and two 24 hr periods in August 1974), and unfortunately did not coincide with strong southward winds (daily mean wind speeds were less than 5 m sec^{-1}). The secondary maxima in the 64 hr mean velocity profile had values less than 2 cm sec^{-1} in August 1973 (JOHNSON *et al.*, 1976). In August 1974, there were only four six-hourly profiles for each 24 hr period (JOHNSON, 1977), and the mean profiles are probably aliased by tidal and inertial oscillations: secondary maxima are less than 5 cm sec^{-1} while standard deviations were ~ 10 cm sec^{-1} (JOHNSON and JOHNSON, 1979). Thus, the evidence from profiling current meters for a two-celled upwelling circulation is rather weak.

The evidence from fixed-level moored current meters indicates that the onshore–offshore circulation consists of only one cell (Fig. 11), with the offshore flow in the surface Ekman layer and the onshore return flow in the geostrophic interior; the flow in the bottom Ekman layer has a weak offshore component when the near-bottom alongshore flow is poleward

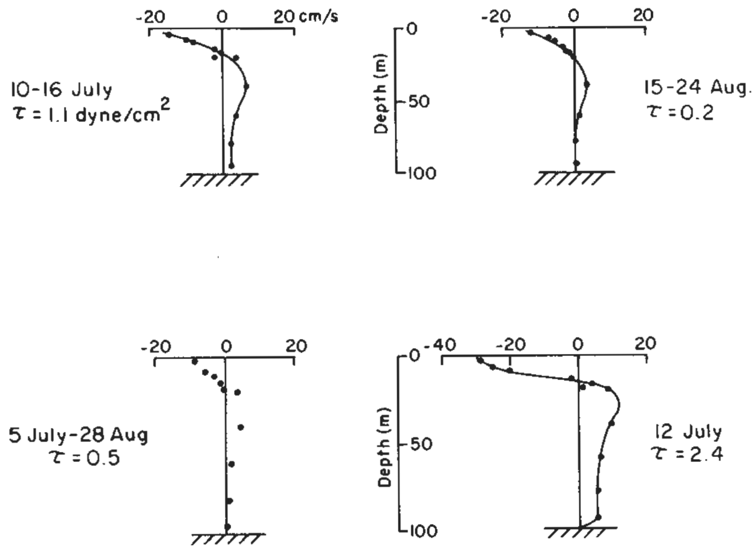

FIG. 12. Profiles of the mean onshore velocity over the 100 m isobath at 45°15′N, with the corresponding mean southward wind stress, for different periods during July and August 1973. Adapted from KUNDU (1977), SMITH (1981) and HUYER (1976).

(i.e., when the undercurrent is present), and a weak onshore component otherwise. The strongest evidence for this pattern comes from current meters moored during July and August 1973, when there was good vertical coverage throughout the water column over the 100 m isobath (HALPERN, 1976; SMITH, 1981). Mean vertical profiles from this mid-shelf array for a six-day period of strong upwelling (KUNDU, 1977), a ten-day period of weaker upwelling (KUNDU, 1977), and the full 55-day period (BRYDEN, 1978; SMITH, 1981) all show the same structure (Fig. 12), with offshore flow in a thin surface layer, onshore flow in a thicker interior layer and rather weak flow near the bottom. This same structure is also observed on 12 July 1973 (Fig. 12), at the height of the strongest upwelling event of the season (HALPERN, 1976; HUYER, 1976). The onshore component of the first empirical orthogonal mode of the low-frequency (< 0.6 cpd) current fluctuations also has the same structure (SMITH, 1981). Simultaneous current measurements over the inner shelf and outer shelf did not include surface layer observations but subsurface observations are consistent with the one-celled pattern during both weak and strong southward winds, and in the mean (Fig. 13).

The observed complex vertical distributions of temperature, chlorophyll and optical parameters (SMALL and MENZIES, 1981; KITCHEN *et al.*, 1975; KITCHEN, ZANEVELD and PAK, 1978) seem at first to contradict such a simple one-celled onshore/offshore circulation pattern. If these complexities cannot be attributed to the onshore/offshore circulation pattern, they must originate in some other fashion. There are two likely candidates: advection in the alongshore direction; and transient, higher-frequency motions over the shelf. Alongshore advection apparently causes the persistent relative minimum in the vertical temperature distribution over the continental shelf (HUYER and SMITH, 1974), but there are also transient maxima in the vertical temperature distribution. These warm anomalies are apparently caused by transient, localized phenomena, in the presence of a mean flow with significant vertical shear. For example, the surface mixing might briefly penetrate deeper than usual at some location; the resulting deep, relatively warm water might then be overlain by a layer of colder but lighter

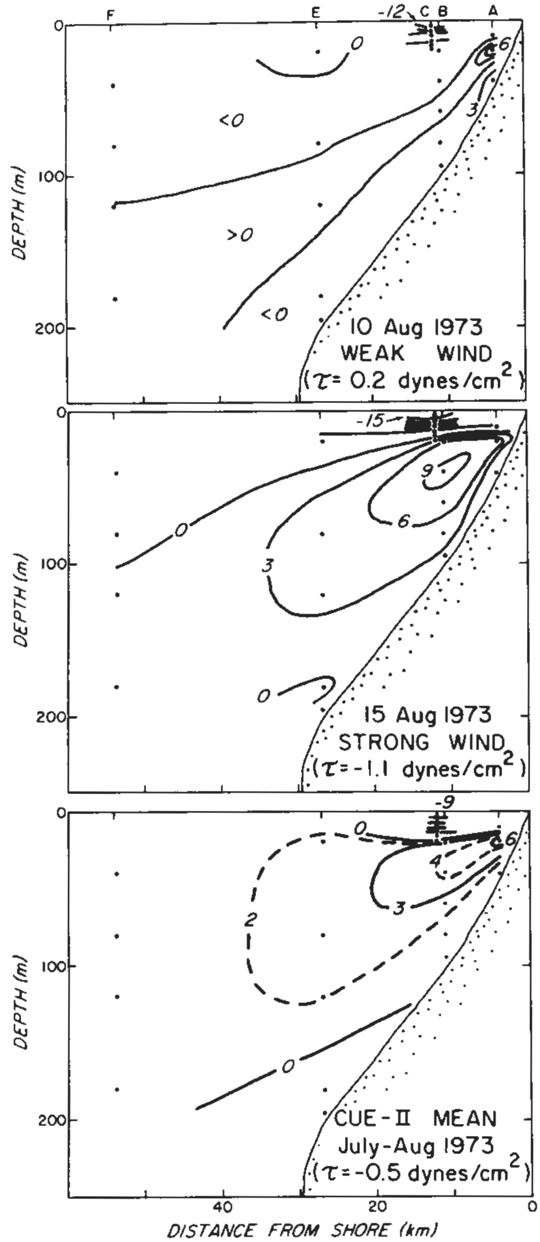

FIG. 13. Distribution of the low-passed onshore velocity over the Oregon shelf at 45°15′N, on 10 and 15 August 1973 (from HUYER, 1976) and the distribution of the mean onshore velocity (plotted from values given by KUNDU and ALLEN, 1976).

water of lower salinity. There may be mixing and stirring of waters well below the surface due to the presence of strong vertical shear and even breaking of internal waves (WANG and MOOERS, 1977). HAYES and HALPERN (1976) showed that the energy of tidal, inertial and shorter-period internal waves varied greatly during a three-week period in July 1973, and

WANG and MOOERS (1977) showed that the temperature–salinity relationship of shelf waters changed significantly during the same period. Thus there is evidence that high-frequency motions cause internal mixing and probably account for some of the complexities in the temperature field. These phenomena would also cause complex structure in the vertical distributions of chlorophyll and optical properties.

Vertical velocities in the upwelling zone are frequently estimated from isopycnal displacements during particular upwelling events, but this method always yields underestimates because mixing and diffusion tend to counteract the density changes due to upwelling. Nevertheless, these estimates range as high as 2×10^{-2} cm sec^{-1} during summer upwelling events (HUYER, 1974; HALPERN, 1973, 1976). BRYDEN (1978) used the principles of conservation of heat and mass to estimate the mean vertical velocity over the 56 days of the CUE-II current observations; he found the mean to be 1.7×10^{-2} cm sec^{-1}, i.e., almost as large as the largest isopycnal-displacement estimates. JOHNSON (1978) calculated vertical velocities from the horizontal flow divergence during and after a weak upwelling event in August 1974, and also obtained values of about 2×10^{-2} cm sec^{-1}. During strong upwelling events, maximum vertical velocities are probably considerably larger.

The magnitude of the vertical velocity decays approximately exponentially with distance from shore with an offshore length scale of about 15 km, which is roughly equal to the local Rossby radius of deformation (HALPERN, 1976). The alongshore current fluctuations and the coastal jet also appear to have an offshore decay scale of 15–20 km, i.e. roughly the Rossby radius, but this may be coincidental.

Many of the effects of upwelling, e.g. the modified temperature distribution, nutrient distributions, etc., may extend well beyond the zone of active upwelling. For example, if cold water is upwelled and advected offshore, the offshore temperature gradient will depend on the rates of heating and mixing as well as on the rate of upwelling (DE SZOEKE and RICHMAN, 1981). Off central Oregon, the offshore limit of water influenced by upwelling seems to be set by the location of the inner boundary of the Columbia River plume. Figure 14 shows surface distributions of several variables observed in the late stage of a strong upwelling event: the inner boundary of the plume (as indicated by the 32‰ isohaline) lies roughly along the 100 m isobath, near the outer edge of the shelf; the highest chlorophyll concentrations lie near this boundary, and coldest temperatures and highest nutrients lie well inside. Later in summer the plume usually lies farther from shore, and the zone of low temperature, high salinity and high nutrients is correspondingly wider (HUYER, 1974; TOMLINSON *et al.*, 1973).

Another interesting feature of Fig. 14 is the tendency of contours of the different variables to lie roughly along isobaths. KUNDU and ALLEN (1976) have shown that the velocity fluctuations are also aligned roughly parallel to the local isobath.

The summer upwelling regime off Oregon seems to be governed by simple Ekman dynamics to a good first approximation, but there are significant complications. The mean offshore transport in the upper layer agrees with the mean offshore Ekman transport computed from the wind stress, and low frequency variations in the upper layer offshore transport also agree with variations in the computed Ekman transport (SMITH, 1981). However, the mean onshore transport in the lower layer exceeds the mean offshore transport in the upper layer by about a factor of three, implying a significant alongshore gradient in the alongshore velocity (positive $\partial \bar{v}/\partial y$), whose existence is supported by the observation that the mean southward flow along the 100 m isobath increases by 2.5 cm sec^{-1} between moorings 50 km apart (SMITH, 1981). The presence of a mean positive $\partial v/\partial y$ may in part be due to large scale dynamics: off Oregon, the summer surface current is about 20 cm sec^{-1} southwards, but somewhere up north (say at 52°N) there is no

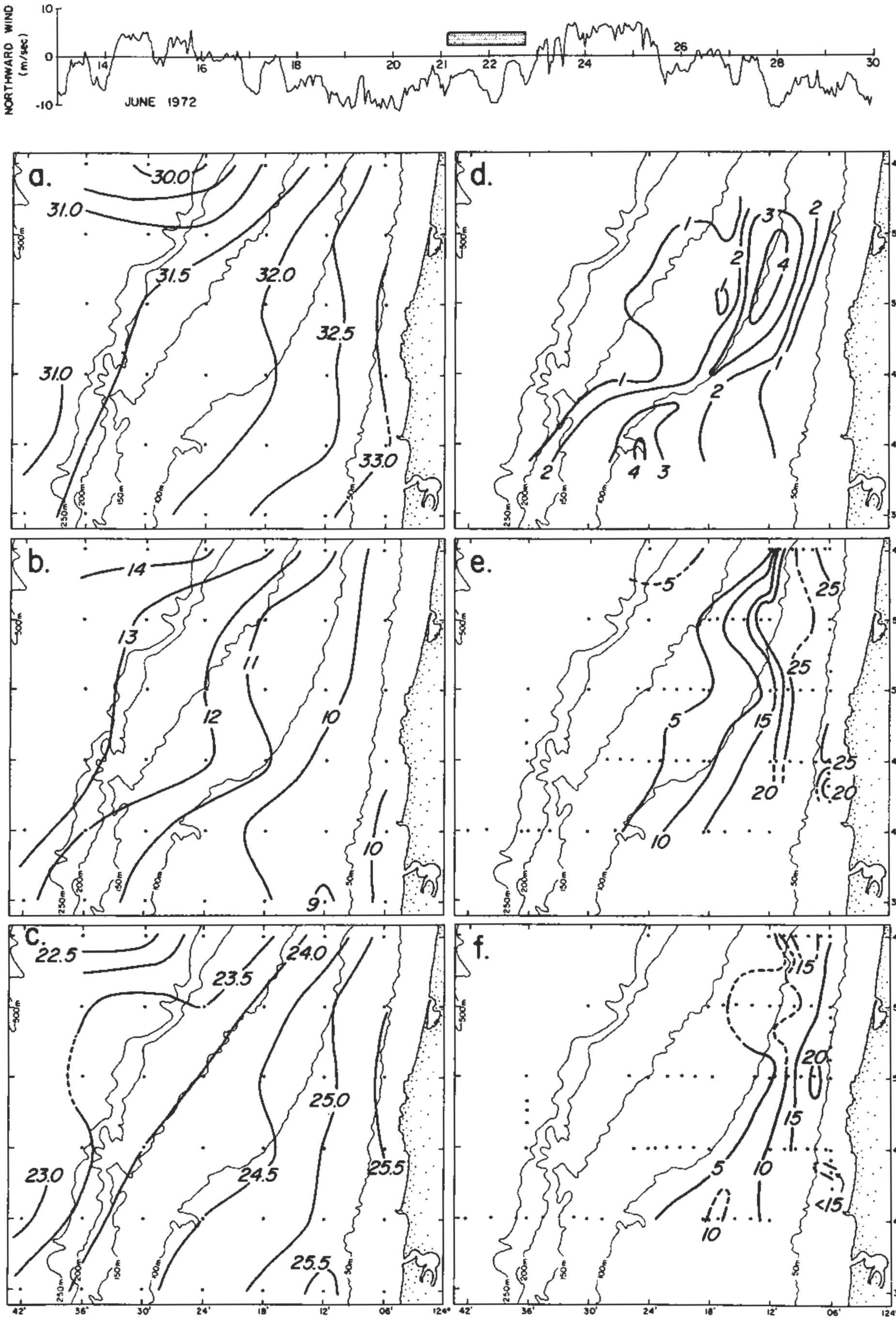

FIG. 14. Surface maps of (a) salinity, (b) temperature, (c) sigma-t, (d) chlorophyll, (e) nitrate and (f) silicate on 21–22 June 1972, during upwelling-favorable winds (shown above maps). The chlorophyll map is from SMALL and MENZIES (1980), and the nutrient maps are from TOMLINSON *et al.* (1973).

mean southward coastal current; if this change is gradual, it corresponds to a mean dv/dy of about 3 cm/sec per 100 km. This estimate is in surprisingly good agreement with SMITH's observation and also in qualitative agreement with the 6 cm sec^{-1} difference over 200 km observed between southern Washington and central Oregon during later summer 1972 (HUYER *et al.*, 1975b).

SMITH (1981) found that although upper layer and lower layer observed mass transports were both significantly correlated with the calculated Ekman transport (at the 99% level), they were not correlated with each other even at the 90% significance level. This may be due to the presence of continental shelf waves (CUTCHIN and SMITH, 1973; HSIEH, 1982a, b) even though the coastal sea level gradient does not balance the alongshore acceleration and onshore transport terms in the alongshore momentum equation (ALLEN and SMITH, 1981): second mode shelf waves, which may have dominated (HSIEH, 1982a), have only a very small signature in coastal sea level. Another possibility is that the onshore flow is subject to variations in the baroclinic alongshore pressure gradient, as suggested by ALLEN and SMITH (1981): the complex sea surface temperature maps of HOLLADAY and O'BRIEN (1975) strongly suggest that such gradients occur in this region.

3.2. *Other seasons*

Periods of upwelling-favorable winds occur occasionally in winter, but their effects have not been observed in detail. The oceanographic regime over the shelf is quite different in winter than summer: there is no mean southward surface current and no mean vertical shear, and isopycnals are approximately level (HUYER *et al.*, 1975a; HUYER, 1977). Even the low frequency current fluctuations have a different structure in winter: they have a larger offshore decay scale, and a fluctuating vertical shear with associated density fluctuations (HUYER, SMITH and SOBEY, 1978). The WISP program designed to observe the transition between these regimes was conducted in late winter and spring 1975: CTD sections were made about once per month, and three current meter arrays (equipped also with temperature, conductivity and pressure sensors) were moored across the shelf at 45°N. A period of upwelling-favorable winds began on 25 March and continued intermittently through 31 March. During this upwelling event, the nearly level or downward sloping isopycnals rose rapidly near shore (Fig. 15) to their characteristic upward-sloping summer position. The maximum rate of isopycnal displacement during this event (40 m in 12 hr) corresponds to a vertical velocity of 9×10^{-2} cm sec^{-1}. The offshore Ekman transport seemed to be confined to the upper 25 m or so, and the compensating onshore transport occurred in the bottom Ekman layer, consistent with the strong southward flow observed throughout the water column (HUYER, SOBEY and SMITH, 1979). Onshore flow in the bottom Ekman layer and strong southward flow at depth was simultaneously observed over the mid-shelf at 46°48′N (SMITH and LONG, 1976). This return flow through the bottom Ekman layer is in marked contrast to the summer situation, when the return flow is through the geostropic interior. Unlike the summer upwelling events, this one began with approximately level isopycnals, as do most models of time-dependent upwelling (e.g. O'BRIEN and HURLBURT, 1972; THOMPSON and O'BRIEN, 1973; ALLEN, 1973). The time scales of the density changes and of the development of the coastal jet during this event are consistent with those predicted by these models. This single upwelling event apparently caused the transition from the typical winter regime with level isopycnals and no mean vertical shear to the typical spring/summer regime with sloping isopycnals, a mean southward surface current and a strong mean vertical shear (HUYER, SOBEY and SMITH, 1979). It

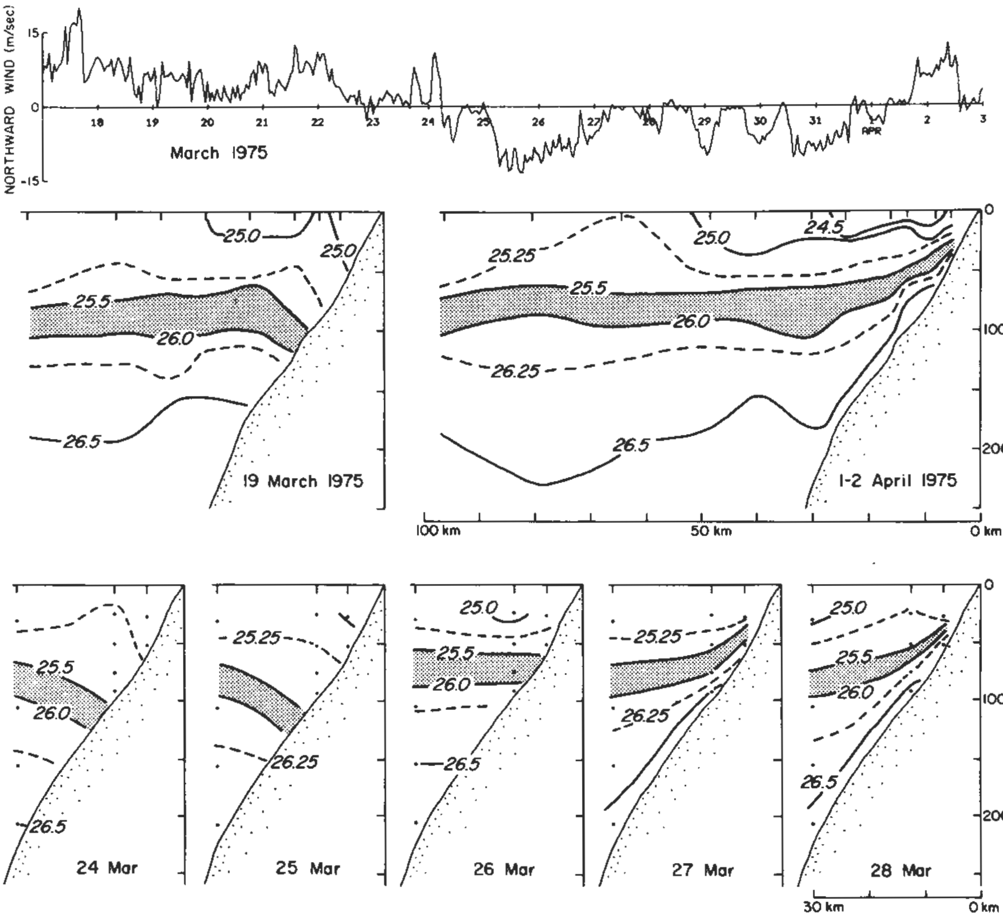

FIG. 15. Distributions of sigma-t along 45°N, before, during and after an upwelling event that began on March 1975. The 19 March and 1–2 April sections are based on CTD observations; the daily sections for 24–28 March show the distribution of sigma-t at 0000 GMT calculated from hourly data recorded by ten current meters equipped with temperature and conductivity sensors.

apparently also caused changes in the structure of the low-frequency current fluctuations: decreasing their offshore length scale and decreasing the fluctuating shear (HUYER, SMITH and SOBEY, 1978), and changing the dominant mode of continental shelf waves from the first to the third mode (HSIEH, 1982b). Thus a single upwelling event of several days duration can cause a seasonal change in the oceanographic regime which persists for several months.

4. UPWELLING OFF SOUTHERN CALIFORNIA AND BAJA CALIFORNIA

The California Cooperative Fisheries Investigations (CalCOFI) program has conducted large-scale, long-term studies of the waters of California and Baja California since 1949 (e.g. REID, RODEN and WYLLIE, 1958). Most of the observations are between Pt. Conception at 35°N and Cabo San Lazaro at 25°N; sections are 40 n mi. apart and they extend offshore about 200 n mi. The station spacing along sections is relatively coarse: even in the near-shore zone,

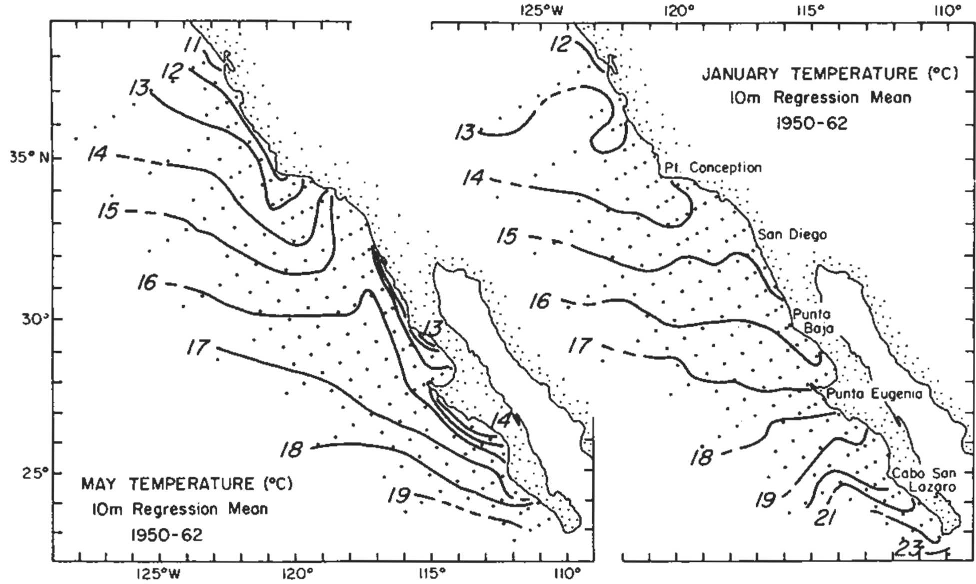

FIG. 16. Distribution of the mean temperature at 10 m for May and January in the CalCOFI region (from LYNN, 1967).

they are generally 20 n mi. apart. Nevertheless, the effects of coastal upwelling are obvious in most of the mean monthly distributions of temperature and salinity at 10 m (LYNN, 1967). In May, when local upwelling-favorable winds are strongest (NELSON, 1977), there are strong temperature gradients with cold coastal water wherever the coastline is parallel to the predominately southeastward winds: San Francisco to Pt. Conception, San Diego to Pta. Baja (at 30°N), and Pta. Eugenia to Cabo San Lazaro (Fig. 16). In each of these regions, the temperature gradients are strongest within about 20 n mi. of the coast i.e., the offshore length scale is similar to that observed off Oregon in summer. Although winds along Baja California are favorable for upwelling throughout the year (Figs 2, 3), there is no evidence of coastal upwelling in the January temperature distribution except between San Diego and Pta. Baja (Fig. 16). The absence of a temperature signal south of Pta. Eugenia (28°S) may be due partly to the increased solar insolation there (BAKUN and NELSON, 1977).

The large-scale temperature distributions (Fig. 16) show no effects of coastal upwelling in the northern part of the Southern California Bight, where winds are relatively weak (Fig. 3). More detailed study of this area (TSUCHIYA, 1980) shows a narrow zone of southward surface flow along the coast from March to May, when upwelling is strongest. There is little day-to-day variation in sea surface temperatures in the northern part of the Southern California Bight, but upwelling events do occur along the coast south of Balboa at about 33°40′N (DORMAN and PALMER, 1981).

Small-scale, short-term studies of the coastal upwelling regime off Baja California show striking similarities with the Oregon summer regime. In June 1976 at 31°N, BARTON and ARGOTE (1980) observed isopycnals sloping generally upward toward the coast over the continental shelf; strong vertical shear in the alongshore flow, with strong (~ 50 cm sec^{-1}) southward flow at the surface and weak (~ 5 cm sec^{-1}) southward flow at the bottom over the mid-shelf; and offshore flow in the surface layer. At 27°N, south of Pta. Eugenia, WALSH *et al.*

(1977) observed southeastward surface currents in late March and early April 1973, with a mean vertical shear in the alongshore component; they observed offshore flow in the surface layer, and onshore flow at mid-depth and near the bottom; they also observed relatively cold surface water (< 15°C) very near shore, and strong surface temperature gradients within 30 km of the coast. In both locations, winds were upwelling-favorable during all or almost all of the study period; variations in wind stress resulted in depth variations of isotherms and isopycnals over the shelf; and some poleward flow was observed near the bottom (WALSH *et al.*, 1977; BARTON and ARGOTE, 1980).

The long-term nature of the CalCOFI observations has made it possible to study interannual variations in this region (CHELTON, 1980, 1981). Nonseasonal changes in the strength of the California Current appear to be correlated with anomalies of coastal sea level, but not with local wind stress; the sea level anomalies in turn are highly correlated with sea surface temperature in the eastern tropical Pacific (CHELTON, 1981). These equatorial anomalies may not extend to the northern part of the California Current system: at San Francisco and farther north, sea level anomalies are significantly correlated with local wind stress (ENFIELD and ALLEN, 1980).

5. SUBJECTS FOR CURRENT AND FUTURE RESEARCH

There are still major gaps in our knowledge and understanding of coastal upwelling along the west coast of North America. Perhaps the most obvious is the geographical one – there have been very few studies of the region between San Francisco at 38°N and Newport at 44°40′N. This is a stretch of more than 700 km of coastline that includes several significant promontories (Cape Blanco, Cape Mendocino, Pt. Arena and Pt. Reyes), a change of coastline orientation from north–south to northwest–southeast, and some major irregularities in the shelf-slope topography. All of these features may locally enhance vertical velocities in the near-shore zone (YOSHIDA, 1967; PEFFLEY and O'BRIEN, 1976) and thus contribute to the strength of coastal upwelling. This stretch of coastline also includes the region of the strongest upwelling-favorable winds (Figs 2, 3), and some areas of locally strong wind stress curl (Fig. 5). Hence, the structure and dynamics of the coastal upwelling regime may be considerably more complex in this region than off central Oregon, where the coastline and bathymetry are relatively simple and upwelling-favorable winds are moderate. This geographical gap will be partly filled by field work now in progress: the Coastal Ocean Dynamics Program (CODE) is conducting an intensive study of the continental shelf between Pt. Arena and Pt. Reyes (BEARDSLEY, 1980), and a large-scale shelf experiment (ALLEN, HUYER and SMITH, 1981) includes CTD sections and current measurements off Coos Bay (43°13′N), Crescent City (41°54′N) and Half Moon Bay (37°25′N).

The extreme southern and northern ends of the California Current also deserve further study. Although upwelling favorable winds persist throughout the year off southern Baja California, there is no noticeable effect on the surface temperature distribution during much of the year. Is this entirely due to the reduced cloud cover there? Or is it because the warm surface layer is so deep that upwelling does not reduce the surface temperature? Or is there such a strong poleward undercurrent that the upwelling is not sufficient to overcome the downward isopycnal slopes associated with poleward geostrophic flow? At the northern extreme, off northern British Columbia, the summer upwelling season must consist of only a few upwelling events. Does the regime there resemble the Oregon summer regime, or the Oregon winter regime? The very complex bathymetry in this region may result in a radically different upwelling regime.

There are considerable discrepancies between the time and space scales of the studies that have been conducted in the northern and southern parts of the California Current system. Little is known about interannual variations off Oregon and northern California, or about day-to-day variations off Baja California. Not much is known about the near-shore structure of current fluctuations along Baja California, nor about the larger-scale structure of the California Current off Oregon and Washington. There is speculation that the narrow coastal jet observed off Oregon gradually widens and becomes the broader California Current observed off southern California, but this hypothesis remains to be verified or rejected by future work.

We do not fully understand the interaction of processes with different time scales, and the effect of these interactions on the upwelling regime. In particular, is it not clear how or why a single upwelling event can induce a persistent seasonal transition in the flow pattern. We do not know whether the resultant change is limited to the continental shelf, nor whether it extends farther offshore along some stretches of coastline. We are not sure how the slowly varying seasonal cycle of the wind stress along the coast affects coastal upwelling at different locations. Ongoing research programs may answer some of these questions, but others will remain for future research.

6. SUMMARY

A great deal has been learned about coastal upwelling in the California Current system since it was reviewed by WOOSTER and REID (1963). Many of the largest scale features were already known then, but the new analyses of the large-scale winds by BAKUN (1973, 1975) and NELSON (1977) have provided considerably more detail on alongshore variations within the system. ROBINSON's (1976) new atlas of the North Pacific clearly shows the seasonal influence of upwelling on both surface and subsurface temperatures along the coast.

Most of the new knowledge is in the local structure and dynamics of the active coastal upwelling zone, limited to a coastal strip only a few tens of kilometers wide. This zone has been studied intensively off central Oregon, where the coastline and shelf topography are relatively simple, and seasonal upwelling-favorable winds are only moderate. We are not yet certain to what extent the observations there apply to other parts of the west coast. It appears that the width of the active coastal upwelling zone is roughly equal to the Rossby radius, which is about 10–25 km along the entire coast from 50°N to 25°N. Within this narrow coastal zone, the shelf waters off Oregon in summer can be divided into a surface Ekman layer, which contains the offshore Ekman transport, a quasi-geostrophic interior which contains the onshore return flow, and a thin bottom Ekman layer. Within the geostrophic interior, there is persistent vertical shear, with the strength of the southward flow decreasing with increasing depth; the shear is balanced by the offshore density gradient associated with upward sloping isopycnals. This kind of shear has been observed off Baja California as well as off Oregon. Fluctuations in the wind stress during the upwelling season off Oregon cause variations in the strength of the onshore–offshore circulation, but they do not much affect the alongshore velocity shear or the isopycnal slopes in the geostrophic interior. Instead, they result in quasi-barotropic alongshore current fluctuations with most of the characteristics of continental shelf waves. A single upwelling event in early spring seems to cause the transition from the typical winter to the typical summer oceanographic regime. Off Oregon, the local wind seems to account for most of the variability of the alongshore currents. In the southern part of the California Current system, the influence of local winds seems to be less important.

Current research projects which include intensive observations in the region of very strong upwelling-favorable winds off northern California will certainly shed new light on the structure and dynamics of coastal upwelling in the California Current System.

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