Measurements of Near-Surface Wind Stress Over an Upwelling Region Near the Oregon Coast

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ABSTRACT

This paper describes the temporal and spatial variations of the wind stress (computed from the square of the wind velocity vector) and wind-stress curl recorded during July and August 1973 at two moored buoy stations, one 13 km and the other 120 km from the Oregon coast, along 45°15'N. Some facets of the relationship between wind stress and the physical oceanography over the continental shelf off Oregon are discussed.

1. Introduction

Throughout July and August, 1973, near-surface wind measurements were made from two buoys, one installed 4 July at 103 m depth about 13 km from the Oregon coast (buoy B: 45°15.8'N, 124°07.8'W), and the other on 5 July at 2575 m depth about 120 km from the coast (buoy H: 45°14.5'N, 125°29.6'W; see Fig. 1). The wind recorders and their calibrations and the mooring configurations have been described in Halpern *et al.* (1974). Visual observations of the pitch and roll motions of the surface float suggested that the tilt angles of the wind sensors were usually no greater than about 20° from the vertical. According to Pond (1968), this would produce errors of up to 10% in determining the wind stress. Since the period 12–15 July was the only time during which surface waves were higher than 1 m for even 1–2% of the time (Halpern *et al.*, 1975), the wind sensors were practically never in the shadow of a wave crest.

At the offshore site the wind recorder measured 14 sequential samples for 74 s every 15 min. Each sample, consisting of a speed, a compass direction, and a vane direction, was measured at 5.27 s intervals. Between each 74 s recording period the instrument was in a "shutdown" mode; wind fluctuations with periods between 74 s and 30 min were aliased. At the onshore site the wind recorder operated continuously and recorded integrated values of the east and north component speed at 3.75 min intervals. Summaries of the wind observations have been presented elsewhere (Halpern *et al.*, 1974) as standard statistics, progressive vector diagrams, time-series plots of speed, direction, east and north components, and Cartesian-component and rotary-component spectra.

2. Results

Vector-averaged hourly values of the east (positive x direction) and north (positive y direction) wind-stress components (τ_0^x, τ_0^y) were computed. We used the bulk aerodynamic parameterization

$$\frac{(\tau_0^x,\tau_0^y)}{\rho_a}\equiv(u_*^2,v_*^2)\equiv C_z(|\mathbf{W}_z|U_z,|\mathbf{W}_z|V_z),$$

where ρ_a, C_z, u_* and v_*, and $\mathbf{W}_z$ represent the density of air (1.25×10^{-3} g cm^{-3}), the drag coefficient, the east and north components of the friction velocity, and the mean wind velocity vector (with east and north speeds U_z and V_z) relative to the surface current. Since the ratio of the surface-current speed to the wind speed was nearly always less than 0.02, the velocity of the surface water was neglected. C_z and $\mathbf{W}_z$ are specified for the same height z. As Pond (1975) indicates, $|\mathbf{W}_z|$, U_z and V_z should each represent a wind speed averaged over one to at most a few hours. We referenced our observations to the 10 m level and used a constant drag coefficient C_{10} of 1.3×10^{-3} (Kraus, 1972). Assuming hydrostatically neutral conditions and the logarithmic wind profile, the wind speed at the 10 m level is related to the wind observations made at 2 m above the sea surface by

$$(U_{10},V_{10})-(U_2,V_2)=1.69\frac{(u_*,v_*)}{\kappa}.$$

The value of von Kármán's constant κ is 0.4, although Businger *et al.* (1971) have suggested 0.35. Substitution of $(u_*,v_*)=\sqrt{C_{10}}(U_{10},V_{10})$ yields

$$(U_{10},V_{10})=1.17(U_2,V_2),$$

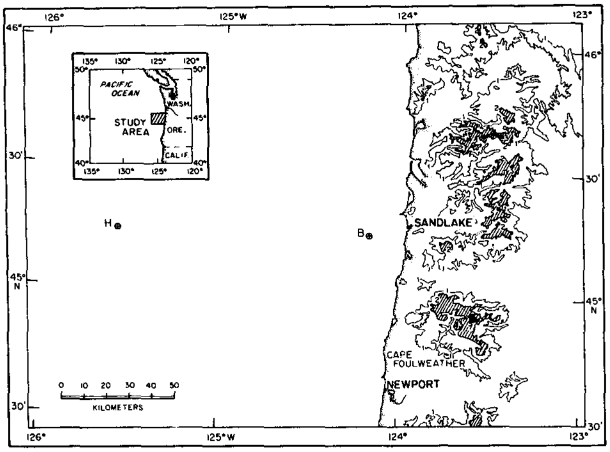

FIG. 1. Location diagram showing positions of wind measurements. Topographic height contours are in thousands of feet. Dark areas represent heights greater than 3000 ft, and diagonal areas are heights between 2000 and 3000 ft.

where U_{10} and V_{10} were hourly vector-averaged values of the 10 m level winds. Our estimates of wind stress contained a minimum uncertainty of 20–30% because we neglected to account for the variations of C_{10} with wind speed (Smith and Banke, 1975) and with atmospheric stability (Katsaros, 1969).

Some standard statistics of the wind stress computed from the vector-averaged hourly values are listed in Table 1. At both stations the winds varied from calm to moderate, reaching a maximum hourly averaged stress value at 0000 GMT 12 July of about -5 dyn cm^{-2} at the inshore station and at the same time nearly -3 dyn cm^{-2} at the offshore station, a maximum difference of about 2 dyn cm^{-2} (100 km)$^{-1}$. At the onshore station the mean values of the east-west and north-south components give little evidence of net zonal flow, only of meridional motion. Apparently the Pacific Coast mountains adjacent to the central Oregon coastline prevented the development of energetic eastward surface motion (Halpern, 1974a).

The mean values of the alongshore component of the wind stress decreased toward the coast (Table 1). This trend, which was opposite to the coastal wind profile postulated by Bang and Andrews (1974) for the upwelling region off South Africa, seemed to continue at least up to the coastline, if wind speeds and directions measured by the National Weather Service at Newport were representative of the winds at Sand Lake (Halpern, 1974b). The zonal gradient of the mean southward wind stress between the onshore station and the coastline was equal to about 14×10^{-8} dyn cm^{-3}; between the two buoys it was 2×10^{-8} dyn cm^{-3}. Presumably, frictional retardation of the winds by the land produced the positive mean gradient.

The energetic diurnal-period fluctuations characteristic of Oregon coastal winds (Burt *et al.*, 1974; Halpern, 1974a) were removed from the data since our interest was directed toward the low-frequency variations. At the onshore station the ratio of the variance of the fluctuations of the wind stress with frequencies greater than $(40)^{-1}$ cycles per hour (cph) to the total variance was 0.24; at the offshore station this ratio was 0.17. The low-pass $[f_c=(40)^{-1}$ cph$]$ filtered time series of the wind stress at each station (Fig. 2) were obtained by Helms's (1967) "Four T" method. The Fourier coefficients were computed by Brenner's (1967) fast Fourier transform algorithm, subtracting the harmonics with frequencies >0.025 cph and retransforming the residual coefficients.

Because of the colatitudinal positions of the two buoys, the vertical component of the wind-stress curl,

$$\mathrm{curl}_z\boldsymbol{\tau}_0=\frac{\partial\tau_0^y}{\partial x}-\frac{\partial\tau_0^x}{\partial y},$$

TABLE 1. Standard statistics of the vector-averaged hourly values of the wind stress computed from wind observations recorded at stations B and H for the 52-day period 6 July–26 August, 1973 ($N=1248$). Units are dyn cm^{-2}.

	Mean		Standard deviation		Minimum		Maximum	
	B	H	B	H	B	H	B	H
$\lvert\tau_0\rvert$	0.66	0.84	0.74	0.62	0.00	0.00	4.91	2.96
τ_0^x	0.01	0.09	0.15	0.21	−1.19	−1.06	0.85	1.28
τ_0^y	−0.50	−0.72	0.84	0.72	−4.84	−2.96	2.71	1.51

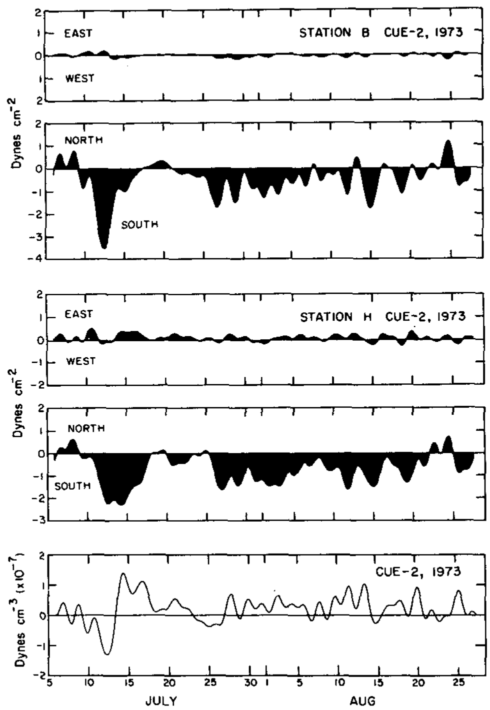

FIG. 2. Zonal and meridional components of low-pass filtered wind stress measured at stations B and H during CUE-2 (upper four sections), and zonal gradient of low-pass filtered meridional component of wind stress (approximately equal to wind-stress curl) between stations B and H (bottom section).

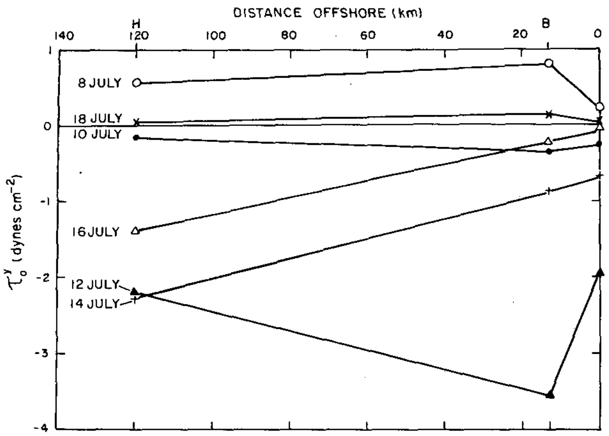

FIG. 3. Zonal gradient of low-pass filtered meridional components of wind stress occurring at 1200 GMT on alternate days during the interval 8–18 July in the region between the coastline and 120 km offshore. Wind stress at the coastline (Sand Lake) is assumed to be equal to the stress measured at Newport, Ore.

was approximately equal to the difference of the north-south wind-stress components; i.e.,

$$\mathrm{curl}_z\boldsymbol{\tau}_0 \sim \frac{\partial \tau_0^y}{\partial x} = \frac{\tau_0^y(\mathrm{B}) - \tau_0^y(\mathrm{H})}{L_{\mathrm{BH}}},$$

where $(\partial/\partial x, \partial/\partial y)$ are partial derivatives in the eastward (x) and northward (y) directions, $\tau_0^y(\mathrm{B})$ and $\tau_0^y(\mathrm{H})$ are the stresses at stations B and H, and L_{BH} the distance between the two stations. The wind-stress curl (Fig. 2) was computed from the difference of the low-pass north-south component time series. The summertime or seasonal mean value (0.21×10^{-7} dyn cm^{-3}) of the curl was positive and half as great as the standard deviation (0.44×10^{-7} dyn cm^{-3}). At the onshore station the strong southward wind stress associated with the 12 July storm occurred for a relatively short period of about 1.5 days, whereas at the offshore station high values were measured for a much longer period. For about 4 days prior to the occurrence of the maximum wind stress at the onshore station the curl was negative $[\tau_0^y(\mathrm{H}) > \tau_0^y(\mathrm{B})]$, reaching its maximum value of -1.3×10^{-7} dyn cm^{-3} on 12 July. The curl rapidly changed to positive values, reaching a maximum of about 1.3×10^{-7} dyn cm^{-3} on 14 July. Thus, within $\sim$50 h the wind-stress curl varied by 2.6×10^{-7} dyn cm^{-3}.

3. Discussion

The coastal wind regime off Oregon is complex, containing a myriad of time and spatial scales. During the winter months the winds blow predominantly from the south and southeast; during July and August the dominant wind direction is toward the south. At times during July and August the coastal winds reverse direction, shifting from predominantly southward to northward for periods of about 5 days. Often a coastal upwelling event is generated when the wind returns to the southward direction (Halpern, 1974c; Huyer *et al.*, 1974).

The low-frequency temporal variations were large (Fig. 2), especially at the onshore station between 10 and 15 July when the stress values were higher by a factor of 2 than the values at other times during the summer. This 5-day interval occurred between periods of northward stress, producing ideal conditions for the onset and relaxation of a coastal upwelling event (Halpern, 1976). Within the 50 h interval beginning at 0000 GMT 12 July when the maximum stress occurred, the stress at the onshore station varied more than 2.5 dyn cm^{-2}. The maximum magnitude of 3.5 dyn cm^{-2} and the continually varying low-pass filtered wind stress at station B contrasted markedly with constant wind-stress values of 0.5 or 1.0 dyn cm^{-2} used in recent theoretical models (Allen, 1973; McNider and O'Brien, 1973) to describe the mesoscale response of the ocean to wind events over a coastal upwelling region.

Assuming Newport wind observations to be representative of the winds occurring at Sand Lake, a number of 3-point zonal profiles of meridional wind stress (Fig. 3) were prepared at 2-day intervals for the 8–18 July period, which encompassed the strongest upwelling event of the summer. Our estimate of the spatial structure of the wind stress, which must be considered preliminary because of the inadequate resolution, clearly showed it to be a non-uniform function of offshore distance. This is in marked contrast to the simple wind stress distributions used in theoretical models of coastal upwelling (McNider and O'Brien, 1973; O'Brien and Hurlburt, 1972). The significant spatial variations of τ_0^y increase the difficulty of determining the mass balance between the Ekman (1905) transport and the near-surface offshore transport derived from direct current meter measurements. Halpern (1976) examined this correspondence using the measurements at the onshore station.

The vertical motion near the surface of the ocean is generated by the wind-stress curl, in addition to the upwelling or downwelling in the upper layers of the ocean near a coastline caused by a spatially homogeneous wind stress. Stommel (1965) has shown that for linearized stationary motion without horizontal viscous stresses, the horizontal divergence of the transport within the Ekman layer, $\mathrm{div}_H\mathbf{M}_e$, is given by

$$\mathrm{div}_H\mathbf{M}_e = \frac{\mathrm{curl}_z\boldsymbol{\tau}_0 - \beta M_e^y}{f},$$

where β is the northward rate of increase of the Coriolis parameter, and M_e^y the north-south component of the mass transport per unit width of Ekman layer. With zero vertical motion at the sea surface, $\mathrm{div}_H\mathbf{M}_e$

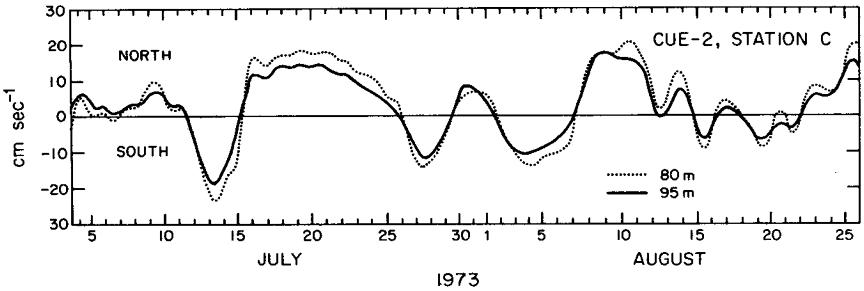

FIG. 4. Low-pass filtered north-south component of currents measured at depths of 80 and 95 m at a subsurface mooring located about 0.5 km from station B. Curves redrawn from Fig. 5 of Kundu *et al.* (1975).

represents the vertical velocity at the bottom of the Ekman layer. At the onshore site the neglect of horizontal viscous stresses seems reasonable according to Kundu and Allen's (1976) study of current-meter data obtained at a number of places on the continental shelf during CUE-1 and CUE-2. The largest values of M_e^y occurred during the 12 July onset of upwelling when $M_e^y \sim -3\times10^4$g cm^{-1} s^{-1} (Halpern, 1976). At 45°N latitude $\beta = 1.6\times10^{-13}$ cm^{-1} s^{-1}. An upper bound of $-\beta M_e^y \sim 5\times10^{-9}$ dyn cm^{-3} was nearly an order of magnitude less than the 52-day mean wind-stress curl. Thus, on the average, $\beta M_e^y \ll \mathrm{curl}_z \tau_0$. An order of magnitude calculation of the mean upward vertical velocity at the bottom of the Ekman layer [e.g., at about 20 m depth (Halpern, 1976; Smith *et al.*, 1971)] shows it to be about 10^{-4} cm s^{-1} offshore of Station B and 10^{-3} cm s^{-1} inshore of this site. On a shorter time scale, such as the period 12–16 July, the maximum offshore wind-stress curl (1.3×10^{-7} dyn cm^{-3}) may have produced an upward motion of magnitude 1×10^{-3} cm s^{-1} for a few days. However, the maximum nearshore wind curl (10^{-6} dyn cm^{-3}; Fig. 3) could have been sufficient to account for the 10^{-2} cm s^{-1} vertical velocity inferred (Halpern, 1976) on 13 July from a time series of hydrographic data.

A poleward undercurrent, observed in many upwelling regions (Gunther, 1936; Hart and Currie, 1960; Smith, 1974), becomes significantly enhanced by a positive wind-stress curl in the upwelling zone (Pedlosky, 1974; Hurlburt and Thompson, 1973). There seemed to be a significant correlation between periods of positive curl (e.g., 14–23 July and 7–14 August) and northward current above the bottom (Fig. 4). Periods of negative curl (e.g., 10–15 July and 24–27 July) were correlated with the absence of a poleward undercurrent, when in fact there was southward flow throughout the water column. Inspection of the wind-stress curves at the onshore station (Fig. 2), the wind-stress curl (Fig. 2), and near-bottom current V measured adjacent to Station B (Fig. 4) reveals a stronger correlation between $\mathrm{curl}_z \tau_0$ and V than between τ_0^y and V. We may hypothesize an observed relationship between wind-stress curl and the direction of the meridional component of the current near the bottom of the continental shelf, but the hypothesis must be speculative because wind measurements from the two buoys are insufficient to determine a realistic wind-stress profile.

In conclusion, these data provided information on the temporal and spatial representativeness of point wind-stress measurements, which are necessary to determine initialization procedures and errors of predictive numerical models. Simultaneous observations at a larger number of offshore locations are required to accurately determine the wind profile in the coastal zone.

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