NOTES AND CORRESPONDENCE

A Comparison of Some Shallow Wind-Driven Currents[1]

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ABSTRACT

Four sets of current measurements made in water depths ranging between 28 and 38 m over periods ranging from three to five weeks are examined and compared. The response of the water column to wind forcing is examined by computing regression coefficients between the surface wind stress and two different parameterizations of bottom stress in terms of measured currents. Coefficients computed for the different data sets vary by as much as a factor of 4. While such variations might be due to instrumental differences, it seems more likely that the assumed dynamical balance between surface and bottom stress is incomplete, i.e., other forces such as the alongshore pressure gradient are quantitatively important even when the water depth is comparable to the turbulent Ekman layer thickness.

Several field experiments have recently been undertaken to examine the response of shallow (30 m depth) waters to surface wind stress. Scott and Csanady (1976) and Beardsley *et al.* (1977a) have measured such currents off the southern coast of Long Island. Hunter *et al.* (1977) maintained a vertical array of current meters on the continental shelf off Chesapeake Bay, and Winant *et al.* (1978) measured currents at several depths off the southern California coast as Tropical Storm Doreen passed near the site in August 1977. These data sets are all several weeks long and mostly include two or three events characterized by notable alongshore winds. While a large number of experiments have been conducted to study wind-driven currents in water 50 m or more deep on different continental shelves, we will limit our attention here to these four sets of current measurements made at shallower depths near 30 m to see if some degree of universality can be deduced from the different measurements.

Imagine an x-y Cartesian coordinate system oriented so that $+x$ is alongshore and $+y$ is offshore. The vertically integrated x or alongshore momentum equation can then be written in the form

$$\tau_{xs} - \tau_{xb} = h(\rho g\partial\zeta/\partial x - \rho f\langle v\rangle + \rho\langle Du/Dt\rangle), \quad (1)$$

where τ_{xs} and τ_{xb} are the alongshore components of the surface and bottom stresses, ρ the water density, h the total water depth, $\rho g\partial\zeta/\partial x$ the effective alongshore subsurface pressure (SSP) gradient, $f\langle v\rangle$ the Coriolis acceleration associated with the depth-averaged or net offshore velocity $\langle v\rangle$ and $\langle Du/Dt\rangle$, the depth-averaged total fluid acceleration in the alongshore direction. In the limit of very shallow water (i.e., as h approaches 0) the right-hand side of (1) must vanish and the essential dynamics must reduce to a balance between the alongshore surface wind stress and bottom stress components. If the current flowing over the bottom is steady on time scales much smaller than characteristic wind-forcing time scales, the bottom stress may be parameterized by the well-established quadratic drag law

$$\tau_{xb} = \rho C_D|u|u, \quad (2)$$

where C_D is a dimensionless drag coefficient and u the x or alongshore current component. On the other hand, a linear relation between current and bottom stress like

$$\tau_{xb} = \rho C_D Q u = \rho r u \quad (3)$$

may be more appropriate when tidal and other background currents are comparable or larger than the mean or wind-driven current components. The parameter Q in (3) represents a measure of the mean rms tidal or other background currents.

We will now compare the four data sets on the assumption that the surface and bottom stress components are equal. Because tidal currents and mean wind-driven currents are often of similar magnitude, it is difficult to decide *a priori* which parameterization of bottom stress (2) or (3) is more realistic.

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Thus we will compare the data using both parameterizations. The first [Eq. (2)] suggests a linear relation between $\rho|u|u$ and the wind stress of the form

$$\rho|u|u = a + b\tau_{xs}, \tag{4}$$

while the other parameterization [Eq. (3)] suggests

$$\rho u = c + d\tau_{xs}. \tag{5}$$

Estimates of C_D in (2) and r in (3) may be computed from the parameters b and d, which in turn can be determined by regression analysis. Since currents and bottom stresses are not perfectly correlated with surface stresses (i.e., other motions exist which are not forced by local winds), it is important to compute regression slopes using surface stress as the *independent* variable. The resulting slope estimates are then not significantly dependent on data characterized by weak wind stress. The four data sets examined here consist of either daily or semi-daily averages of currents and wind stress. In order to obtain comparable results in each case, all data sets were reduced to daily averages. To roughly estimate the effect of this temporal filtering on computed correlations and regression estimates, both the semi-daily and daily averaged data presented by Hunter *et al.* (1977) and Winant *et al.* (1978) were analyzed. The results for each data set differed by less than 10%.

The results of the correlation and regression analysis on the four daily averaged data sets are given in Table 1. Within each data set, the regression estimates for bottom drag (C_D or r) are notably larger for measurements made closest to the bottom. This results from the lower velocities observed near the bottom. In a turbulent bottom boundary layer characterized by a bottom roughness height of order 20 cm and a quadratic drag law coefficient $C_D = 1.1 \times 10^{-3}$ at the top of the boundary layer, the mean velocity is reduced to half the free stream value at a height of about 3 m above the bottom (see Wimbush and Munk, 1970).

Between data sets, the correlation coefficients ρ_1 and ρ_2 characterizing the Winant *et al.* (1978) California data are notably lower than those obtained in the mid-Atlantic Bight. This results from the relatively short amount of time during which significant wind stress occurred compared to the total length of those observations. The correlation coefficients ρ_1 are larger than ρ_2 because the currents induced by local wind stress on the California shelf are larger in magnitude than the background motion.

The regression estimates C_D and r computed at similar heights through the water column in the different experiments vary by as much as a factor of 4. These differences are not noticeably dependent on which regression estimate is considered, nor do the correlation coefficients ρ_1 and ρ_2 differ significantly for the three mid-Atlantic Bight data sets. We are thus unable to determine which form of the bottom stress parameterization [(2) or (3)] is more accurate.

The rather large spread in the regression estimates may be explained by any combination of the following three causes: experimental differences in measuring currents and wind speed, geomorphological differences in the experimental site, and additional dynamics eliminated in the assumed $\tau_{xs} = \tau_{xb}$ bal-

TABLE 1. Summary of the four shallow water studies discussed in text including site location and data collection interval, vertical location of current meter, and the correlation coefficients $\rho_{1,2}$ and regression estimates of C_D and r.

Investigation	Location	Period over which data was collected	Local water depth (m)	Height of current meter above bottom (m)	Comparison of τ_{xs} and $\rho\|u\|u$: Correlation ρ_1	Comparison of τ_{xs} and $\rho\|u\|u$: Regression C_D	Comparison of τ_{xs} and ρu: Correlation ρ_2	Comparison of τ_{xs} and ρu: Regression r (cm s^{-1})
Hunter *et al.* (1977)	36°45′N, 75°10′W (off Chesapeake Bay)	16 Jan–12 Feb, 1974	38.0	6.6	0.82	2.6×10^{-3}	0.85	5.8×10^{-2}
				19.2	0.85	1.2×10^{-3}	0.86	3.9×10^{-2}
				31.2	0.83	1.25×10^{-3}	0.86	4.0×10^{-2}
Scott and Csanady (1976)	40°44′N, 72°25′W (11 km S of Long Island coast)	5 Sep–28 Sep, 1975	32.5	2.8	0.87	11.5×10^{-3}	0.85	17.9×10^{-2}
				16.0	0.88	2.2×10^{-3}	0.85	7.6×10^{-2}
				26.3	0.93	2.1×10^{-3}	0.87	6.7×10^{-2}
Beardsley *et al.* (1977*a*)	40°47′N, 72°30′W (6 km S of Long Island coast)	5 Feb–21 Mar, 1976	27.8	2.8	0.74	14.3×10^{-3}	0.76	20.8×10^{-2}
				12.1	0.76	3.7×10^{-3}	0.76	9.5×10^{-2}
				20.4	0.71	4.5×10^{-3}	0.74	10.4×10^{-2}
				24.3	0.74	3.8×10^{-3}	0.75	9.5×10^{-2}
Winant *et al.* (1978)	32°33′N, 117°13′W (2 km off Southern California coast)	3 Aug–19 Sep, 1977	30.0	3.0	0.33	5.9×10^{-3}	0.24	10.5×10^{-2}
				10.0	0.39	2.1×10^{-3}	0.29	5.1×10^{-2}
				20.0	0.36	2.0×10^{-3}	0.27	5.0×10^{-2}
				28.0	0.55	1.1×10^{-3}	0.42	3.0×10^{-2}

ance. Current meter intercomparison tests like CMICE76 (Beardsley *et al.*, 1977a) and others suggest that current measurement errors in experiments like those considered here are probably less than 20%. Wind speed measurements are notably more accurate, although point stress estimates (using observed wind velocities and an assumed drag coefficient) may not be truly representative of the wind stress over a large coastal area. While the differences in C_D and r might be due to these measurement difficulties, it seems more likely that the assumed dynamical balance $\tau_{xs} = \tau_{xb}$ is incomplete, i.e., the mean and time-dependent body forces such as the alongshore pressure gradient and Coriolis and inertial forces are quantitatively important even in the relatively shallow waters studied here. Three of the four experiments were conducted on the broad continental shelf in the mid-Atlantic Bight, while the Winant *et al.* (1978) data were obtained on the very narrow shelf off southern California. The regional response to wind events includes sea surface adjustment and generation of local SSP gradients which, because they may be partially coherent with the wind, can modify local wind-driven currents. Beardsley *et al.* (1977b) report measurements which imply that the mean and wind-generated, time-dependent, alongshore surface slopes are of the order of 10^{-7} along the New Jersey-Long Island coast. Substitution of $\partial\zeta/\partial x = 10^{-7}$ into (1) yields an effective net stress of $\rho g h \partial\zeta/\partial x = 0.3$ dyn cm^{-2} for 30 m deep water column. A similar net stress is associated with a depth-averaged offshore current of $\langle v \rangle = 1.0$ cm s^{-1} or an alongshore acceleration of $\Delta\langle u \rangle = 8.6$ cm s^{-1} over a 1-day period. The observed thickness of the neutral (one-dimensional) turbulent Ekman layer over a level surface is $h_e \approx 0.3u_*/f$, where u_* is the friction velocity defined by $(\tau_b/\rho)^{1/2}$ [see Tennekes (1973) for discussion]. The inner constant stress or logarithmic layer adjacent to the bottom boundary is much thinner, of order $0.1h_e$. Csanady (1978) and Jones and Kenney (1977) recently demonstrated that a similar logarithmic layer exists in the turbulent surface boundary layer. While the scale thickness of the surface logarithmic layer is not precisely known, it seems reasonable to assume that the surface and bottom boundary layers behave in a roughly similar fashion. If these scaling relationships remain valid when a lateral boundary is present and when the water depth h is comparable to h_e, the body forces then become negligible in (1) when the two logarithmic layers fill the fluid column, i.e., when $h \approx 0.2h_e$. For a midlatitude storm characterized by a 10 dyn cm^{-2} stress, the thickness of the combined constant stress layers is roughly $0.2h_e = 20$ m, which is somewhat less than the local water depth at any of the experimental sites considered here. The rough agreement of the regression estimates shown in Table 1 indicates a crude balance between τ_{xs} and τ_{xb}, however, which may be adequate for engineering estimates of shallow coastal currents driven by local wind forcing.

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