

SOME EFFECTS OF HEATING ON THE WIND-DRIVEN VELOCITY IN THE UPPER OCEAN

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ABSTRACT

The relationship between the mean wind stress and the mean, wind-driven, upper ocean velocity is analyzed using measurements made by a surface mooring in the open Sargasso Sea. Mean values are estimated by ensemble averaging daily averages which have been rotated to a common wind direction. Velocity is referenced to 50 m depth to suppress tides and eddies. The resulting mean velocity is strongly surface trapped, e-folding over about 12 m, and rotates slightly to the right (cum sole) with increasing depth. This vertical structure of the mean velocity spiral is a result of diurnal cycling, and can be simulated well by a one-dimensional model driven with the estimated wind stress and diurnally varying surface heat flux. Volume transport is within about 10% of that expected from a steady Ekman balance, and is nearly 90° to the right of the wind.

INTRODUCTION

The direct, local effect of a steady wind stress, τ , upon a homogenous ocean was calculated by V. W. Ekman at the turn of the century. Ekman assumed that the horizontal momentum flux from the surface wind stress was transmitted through the water column by a Fickian diffusion process with constant effective viscosity, A , and was balanced by the Coriolis acceleration. The velocity profile which satisfies these dynamics has an elegant spiral shape, called the Ekman spiral (Pond and Pickard, 1983), whose depth integral is the Ekman transport,

$$M = \tau \times k / \rho f ,$$

where k is the vertical unit vector, f is the Coriolis parameter, and ρ is the density of sea water. The magnitude and direction of M depend only upon there being a balance between wind stress and Coriolis acceleration, and not at all upon the effective viscosity or any other detail of the vertical mixing. As such, the Ekman transport is on much firmer theoretical ground than the Ekman spiral itself. However, the dynamics of the spiral, and specifically the parameter dependence of its trapping depth, determines where the Ekman transport occurs in the water column,

and thus determines which waters make up the Ekman transport. A sound observational and theoretical understanding of the mean velocity spiral and the accompanying transport is thus an integral part of the larger problem of the wind-driven ocean circulation. Here we give a progress report on our effort to understand these phenomena based upon an analysis of *in situ* velocity observations and numerical modelling.

Most previous attempts to make *in situ* observations of the mean wind-driven velocity and Ekman transport have been frustrated by a difficult signal-to-noise ratio with which we must also contend; the mean wind-driven velocity is small ($O(0.05 \text{ m s}^{-1})$) compared to the orbital motion of surface waves (which are about an order of magnitude larger in this case), and compared to the currents due to tides, inertial motions and nearly geostrophic eddies (which are about five times larger). To observe such a small mean velocity in the upper ocean requires both a current meter which rejects virtually all surface wave motion, and a record length sufficient to average out tides and eddies.

LOTUS FIELD OBSERVATIONS AND THEIR ANALYSIS

For this study we have a five month-long time series of upper ocean velocity and wind made from a surface mooring as part of the Long Term Upper Ocean Study (LOTUS) carried out in the western Sargasso Sea (Briscoe and Weller, 1984). Ocean velocity measurements were made by Vector Measuring Current Meters (VMCMs) which are able to reject nearly all surface gravity wave motions (Weller and Davis, 1980). VMCMs were set at depths of 5, 10, 15, 25, 50, 75, 100 m (and deeper) to give reasonably good resolution in the upper ocean. Meteorological measurements were made with calibrated, state of the art instruments, and surface fluxes were estimated from bulk transfer formulae (Deser et al., 1983; Large and Pond, 1981). We use the longest continuous record of velocity obtained during the LOTUS project, the LOTUS 3 record made from 14 May 1982 to 20 October 1982 (160 days). This period saw mainly summer conditions of strong daytime heating, and light winds (Stramma et al., 1986).

Recent field studies have shown that even the small stratification that occurs as part of the diurnal cycle ($O(0.2 \text{ C})$) (Price et al., 1986) can block the downward penetration of momentum flux from the surface. Thus, the effect of direct wind-driving may be confined to a comparatively thin surface layer during midday, but reaches deeper during the evening when cooling and wind-mixing erode away the diurnal stratification. This insight into the effects of diurnal heating (or stratification) leads to a fairly straightforward analysis method applied here in order to detect the mean velocity spiral and the Ekman transport.

First, given that the directly wind-driven velocity is going to be much more strongly surface-trapped than is the “noise” velocity due to tides and eddies, then it is appropriate and generally necessary to analyze a velocity vertical anomaly rather than the absolute velocity itself. We calculate a vertical anomaly by subtracting the velocity at a reference level, 50 m, from the upper ocean velocity (Davis et al., 1981a; Price et al., 1986). This reference depth was within the seasonal thermocline for all but the last few weeks of the LOTUS 3 record, and we see no evidence of any important direct wind-driving at that level or below. Moreover, the use of other plausible reference depths, 25 or 75 m, gives nearly the same result for the mean velocity spiral.

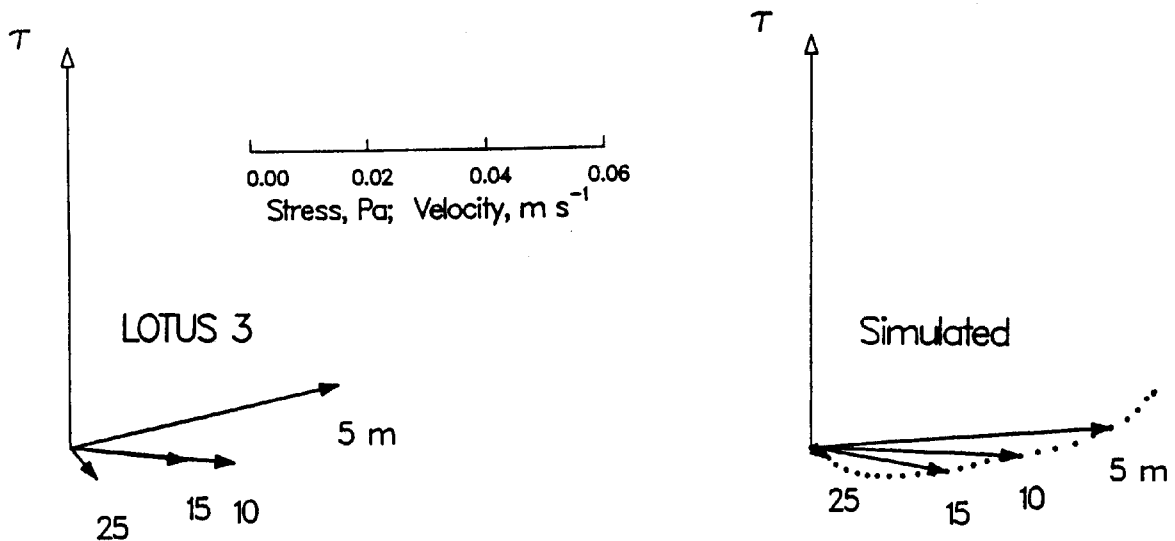
Further suppression of the non-wind-driven velocity is possible only by long-term averaging, and here we introduce a way to coherently average the data to enhance the signal-to-noise ratio. The aim is to account for the effects of varying wind direction. Wind stress and velocity data are first averaged over a day, and then an ensemble average is formed by rotating the daily averages of wind and velocity to a common wind direction (“up” in Figure 1). (The choice of a daily averaging interval for the record pieces is not crucial for the ensemble average, and essentially the same result comes from averaging over pieces of, say, two days, or over pieces of one or several inertial periods.) The coherently averaged wind stress has an amplitude of 0.068 Pa, while the simple time-mean average of wind stress during LOTUS 3 was 0.015 Pa, or about a fourth as large. Thus the coherent ensemble averaging applied here serves to enhance considerably the amplitude of the mean wind stress and Ekman transport detectable in this data set. The ensemble average is interpreted as if the wind had held a constant amplitude and direction over the observation period.

MEAN VELOCITY SPIRAL AND VOLUME TRANSPORT

The mean velocity at the four VMCM depths above 50 m is shown in Figure 1 (left), and uncertainties on the mean velocity at each depth are listed in Table I as standard errors. The integral time scale of the daily values was found to be surprisingly short, only 1.5 days, (most variability is contributed by tides and inertial motions rather than geostrophic eddies), and hence the number of effective degrees of freedom used to estimate standard error was taken to be $160/(2 \cdot 1.5) = 53$. Mean values at 15 m and above are thus fairly well defined, but the 25 m value is not distinguishable from zero.

The mean velocity spiral is strongly surface trapped in the LOTUS 3 data set. The 5 m velocity has an amplitude of about 0.04 m s^{-1} , and is well off to the right of the wind, about 78° . Velocity amplitude decays fairly rapidly with depth, e-folding over about a 12 m scale. The velocity vector rotates only about 20° over the e-

MEAN STRESS AND VELOCITY



TRANSPORT

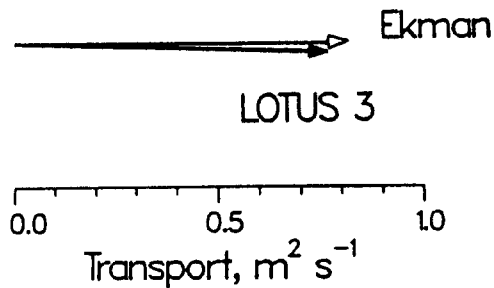


Figure 1. (Left) Mean velocity spiral from the LOTUS 3 data set. The mean has been estimated by an ensemble average over daily averages which have been rotated to a common wind direction (arbitrarily "up" in this figure). Uncertainties on the mean values are listed in Table I. (Lower) The observed transport is plotted (solid arrow head) along with the Ekman transport, (open arrow head). (Right) Mean velocity spiral simulated by the numerical model. The vectors correspond to the depths sampled by LOTUS 3, and the dots are at 1 m intervals.

TABLE I

STATISTICS ON MEAN VELOCITY AND TRANSPORT

Mean Velocity, m s⁻¹

<u>Depth, m</u>	<u>Cross-Wind</u>	<u>Down-Wind</u>
5	0.046 ± 0.012 ^a	0.010 ± 0.007
10	0.028 ± 0.007	-0.003 ± 0.004
15	0.020 ± 0.007	-0.002 ± 0.005
25	0.004 ± 0.004	-0.005 ± 0.004
50 ^b		
75	0.006 ± 0.003	-0.002 ± 0.004
100	0.011 ± 0.006	-0.007 ± 0.006

Transport, m² s⁻¹

	<u>Cross-Wind</u>	<u>Down-Wind</u>
Observed	0.76	-0.02
Ekman ^c	0.82	0

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- a. Uncertainty is standard error. 90% confidence limits are larger by a factor 1.7, and 95% confidence limits are larger by a factor 2.0.
- b. The simple time – mean velocity at 50 m was 0.181 m s⁻¹ westward, and 0.007 m s⁻¹ northward.
- c. Computed from a mean stress of 0.068 Pa.

folding depth so that the spiral found here has a rather flat (and not particularly elegant) shape when compared to the classical Ekman spiral. This has been noted before for other mean, wind-driven velocity spirals (Price et al., 1986, and Weller, 1981). (Elegant or not, the mean velocity does have an unmistakable spiral shape which makes analogy with diffusion models compelling. We do not pursue diffusion models in detail here, but simply note that an effective viscosity which decreases from a surface value of about $400 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ to zero at 30 m depth gives a spiral much like the one observed.)

Volume transport was calculated by a trapezoidal rule integration from 50 m to the surface and is plotted in Figure 1 (lower) along with the theoretical Ekman transport. Perhaps the most striking result is that the estimated volume transport closely approximates the theoretical Ekman transport in both magnitude and direction. We believe that this is the first reasonably sound, quantitative verification of the steady Ekman balance made from direct velocity observations in the upper ocean.

DIURNAL VARIABILITY OF VELOCITY SHEAR

There is an important and in some ways quite dramatic diurnal variability in the wind-driven velocity which is part of the diurnal cycle process of the upper ocean (Kondo et al., 1979; Price et al., 1986; Woods and Strass, 1986). The dynamics of this diurnal variability are central to understanding the surface trapping of the Ekman transport. Diurnal variability is, of course, lost in forming the ensemble average over the full data set, but can be seen by forming ensemble averages for the nighttime, 20 LST to 8 LST (Local Solar Time) and the daytime, 8 LST to 20 LST, Figure 2. This clearly shows that most of the shear in the upper ocean is supported by the stable stratification of the diurnal thermal cycle. Upper ocean shear is greatly reduced during the early morning when cooling by heat loss and wind-mixing have caused the mixed-layer to deepen well below its midday value.

SIMULATIONS

A consistent picture of the mean velocity spiral and the diurnal variability emerges from numerical simulations made with a simple one-dimensional numerical upper ocean model (Price et al., 1986). This model has a surface mixed-layer within which the effective viscosity is infinite and whose depth is controlled by a bulk Richardson number, a transition layer within which a gradient Richardson number controls mixing, and zero diffusion or mixing in the fluid below. The model was forced with the 160-day series of wind stress and surface heat flux from LOTUS 3. With this wind and buoyancy forcing, the depth of the mixed-layer goes through diurnal excursions from typically about 5 m near noon, to about 30 m before sunrise.

DIURNAL VARIABILITY of VELOCITY

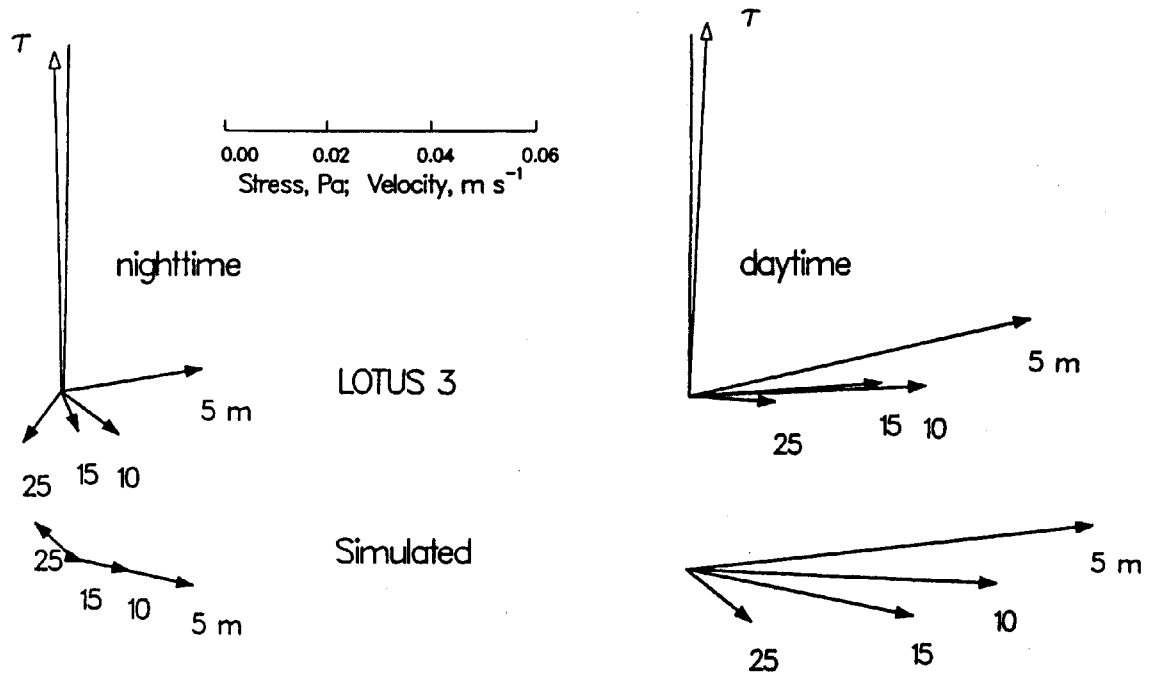


Figure 2. The ensemble average velocity during nighttime (left) and during the daytime. The wind stress for the corresponding time of day is plotted as a vector, with the mean value shown as a solid line. Note that there is very little diurnal variability of wind stress. This figure shows that there is much larger near-surface velocity and overall vertical shear during the day than at night.

Wind-driven velocity was analyzed exactly as it was from the LOTUS data, Figure 1. In the model solution there is, trivially, an exact Ekman balance (over any stable averaging period) since only wind stress and Coriolis forces act on the water column, and hence the transport balance is not plotted.

The important, nontrivial model results are the amplitude of the diurnal cycle, Figure 2, and the overall shape and trapping depth of the mean velocity spiral, Figure 1. The night-to-day variation of the velocity profile is fairly realistic, Figure 2, suggesting that the model responded to surface heating in a realistic way and the trapping depth of the simulated velocity profile is close to that actually observed, e-folding in about 15 m. We know of nothing that makes these results unique to this model, and expect that other upper ocean models (e.g., Mellor and Durbin, 1975; Kondo et al., 1979; Davis et al., 1981b) would perform as well as this one given the same wind stress and diurnally varying heat flux.

DISCUSSION

The principal result of this analysis is that the LOTUS 3 record was found to have a long-term mean upper ocean volume transport consistent with a steady Ekman balance, and that this transport was found to be fairly strongly surface trapped. The model results, together with the observed diurnal variability of the velocity and shear, show that this surface trapping is a consequence of surface heating and the diurnal cycling process. The upper ocean Ekman transport is thus found to be a much more dynamic phenomenon, and with a much richer parameter dependence than might have been supposed from the classical theory. It is heartening to see that modern observational tools like the LOTUS surface buoy and VMCM instruments can measure the upper ocean with enough fidelity and detail, and over sufficiently long periods, to at last provide a solid descriptive basis for developing new and better theories for the upper ocean.

ACKNOWLEDGMENTS

We are grateful to the Office of Naval Research for their support of the LOTUS project through contract N00014-76-C-0197, NR 083-400, and for support of J.F.P., R.A.W, and R.R.S. through contract N00014-84-C-0134, NR 083-400 with the Woods Hole Oceanographic Institution. We thank Ms. Nancy Pennington and Ms. Christina Light for their assistance with data processing and analysis.

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