Organization of stratification, turbulence, and veering in bottom Ekman layers

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[1] Detailed observations of the Ekman spiral in the stratified bottom boundary layer during a 3-month period in an upwelling season over the Oregon shelf suggest a systematic organization. Counter-clockwise veering in the bottom boundary layer is constrained to the weakly stratified layer below the pycnocline, and its height is nearly identical to the turbulent boundary layer height. Veering reaches $13\pm 4$ degrees near the bottom and exhibits a very weak dependence on the speed and direction of the interior flow and the thickness of the veering layer. A simple Ekman balance model with turbulent viscosity consistent with the law-of-the-wall parameterization modified to account for stratification at the top of the mixed layer is used to demonstrate the importance of stratification on the Ekman veering. The model agrees reasonably well with observations in the lower 60–70% of the bottom mixed layer, above which it diverges from the data due to the unaccounted physics in the interior. Neglect of stratification in an otherwise identical model results in far worse agreement with the data yielding veering in the bottom Ekman layer which is much smaller than measured, but distributed over a much thicker layer.


1. Introduction

[2] Friction in the bottom boundary layer (BBL) of the ocean causes ageostrophic transport perpendicular to the interior flow. This was first explored by Ekman, who arrived at a solution for the velocity profile by considering a momentum balance of three terms: the Coriolis force, the horizontal pressure gradient, and a vertical shear stress. This bottom transport is of particular interest on continental shelves, such as off Oregon, because it helps drive the upwelling of deep, nutrient-rich near-bottom water during southwards geostrophic flow (see Kundu [1976], Huyer and Smith [1978], and Perlin et al. [2005a] for a recent example on the Oregon Coast).

[3] Few observational tests of Ekman theory have been made in the bottom boundary layer because the velocities are small and measurements near the bottom are difficult to make. The constant viscosity BBL theory predicts that, as the bottom is reached, the direction of the current will have veered 45° counter-clockwise (N. Hemisphere) from the direction of the interior geostrophic velocity. Observations of oceanic bottom Ekman layers, however, indicate that the counter clockwise veering over the BBL is much less than $45^\circ$, typically reaching a maximum of only $20^\circ$ (Table 1).

[4] The length scale of the neutrally stratified bottom Ekman layer is often given by $D_E = \gamma u_b f$ with the proportionality constant $\gamma$ of order 1, where $f$ is the Coriolis parameter, $u_b$ is the friction (characteristic) velocity in the bottom boundary layer defined as $u_b = \sqrt{\tau_b / \rho}$, $\tau_b$ is the bottom stress, and $\rho$ is density. The length scale $D_E$ can be derived from the momentum balance [Cushman-Roisin, 1994] by assuming that the Ekman number (ratio of horizontal stress to Coriolis force) is of order one: $E \sim 1$, where $E$ is the eddy viscosity, and by assuming that $K_e \sim u_b D_E$, which comes from turbulence theory where $D_E$ is also the scale of the largest turbulent eddies. Previous work has used this scale to characterize the veering thickness, and have often found values of $\gamma$ near 0.4 (Table 1).

[5] In classic Ekman theory, the distributions of velocity and veering are calculated assuming a constant eddy viscosity near the boundary. However, turbulence theory and observations in unstratified boundary layers indicate that the eddy viscosity near a boundary is given by the law-of-the-wall in which the eddy viscosity is proportional to the distance from the boundary. Cushman-Roisin and Malačič [1997] (hereafter CRM) solve the bottom Ekman layer with the law-of-the-wall eddy viscosity and predict much smaller veering angles than the constant eddy viscosity solution. This parameterization has also been used by Ellison [1956] for the atmospheric boundary layer, by Madsen [1977] in a study of the surface Ekman layer, and by many others in studies of tidal currents in shallow well-mixed waters.

[6] A further complication to the classic Ekman theory is the presence of stratification, which suppresses velocity fluctuations and thus turbulence in the boundary layer,
and, in turn, influences velocity veering. Pollard et al. [1973] have argued that in the presence of stratification, the height of the Ekman layer is governed not by \( u_w f \), but by \( u_w \sqrt{N f} \) when \( N \geq f \), where \( N \) is the stratification of the layer adjacent to the mixed layer, representing the ambient stratification. Observational evidence of the modification of the Ekman layer in the presence of stratification can be found elsewhere. For example, Price and Sundermeyer [1999] showed that the Ekman spiral in the surface boundary layer was thinned by stratification and that the wind-driven transport was trapped mainly within the upper part of the Ekman layer. Weller and Plueddemann [1996] showed that the Ekman layer was concentrated above the pycnocline and was divided almost equally between the surface mixed layer and a weakly stratified layer immediately below, and that the diurnal cycle in mixed layer height affects the velocity structure inside the Ekman layer.

[7] We have recently suggested a modification to the law-of-the-wall (MLW) that is intended to account for the suppression of turbulence by stratification away from the boundary but within an unstratified boundary layer [Perlin et al., 2005b]. We have concluded that the MLW predicts significantly different velocity and turbulence dissipation profiles than the law-of-the-wall theory and agrees reasonably well with the measured data in the lower 60–70% of the bottom mixed layer (versus 20–30% for the law-of-the-wall). We have arrived at this conclusion by comparing the measured and modeled velocity and turbulence dissipation rates profiles. Here we extend this analysis to consider rotation and apply the MLW in the form of eddy viscosity formulation to investigate Ekman veering.

[8] Specifically, we demonstrate that in the presence of stratification, Ekman veering coincides with a clearly defined turbulent layer. Most of the Ekman transport occurs inside the well-mixed layer, but also extends into a weakly stratified layer above. Data from a broad, relatively flat region of the BBL on the Oregon Shelf are analyzed. The data, discussed in section 3, include 3 months of high-resolution measurements of velocity to within 2.75 m of the sea floor, supplemented by 50 h of intensive turbulence and density profiling measurements to within 2 cm of the sea floor. From these data we quantify the veering angle and the veering layer height and evaluate dependence of these properties on the interior current velocity and the level of turbulence in the BBL. The veering rate (change in veering angle/veering layer height) has been shown to be nearly inversely proportional to the veering layer height, so that the maximum veering angle has only a very weak dependence on interior flow velocity and thickness of the BBL. Our observations have been compared to solutions of the Ekman balance using both the law-of-the-wall (CRM) and the modified law-of-the-wall [Perlin et al., 2005b] parameterizations of eddy viscosity.

2. Definitions and Theory

[9] The BBL typically exhibits weaker stratification and higher turbulence than the interior. To characterize the state of the BBL, we define four length scales based on observational criteria (Figure 1). The bottom mixed layer (D) is defined as the distance from the bottom over which the potential density decreases by \( 6 \times 10^{-4} \) kg m\(^{-3} \) from its
value at the bottom. To avoid the effects of local overturns
in the estimation of \( D \), we also require the density difference
to remain below this threshold value for at least 1 m. Above
the mixed layer there often lies a weakly stratified layer. The
combined mixed and weakly stratified layers is termed
the remnant layer \( (D_r) \), and its thickness is defined as the
distance from the bottom over which the potential density
decreased by \( 3 \times 10^{-2} \) kg m\(^{-3} \). At least during upwelling
conditions, the top of the remnant layer marks the lower
boundary of the pycnocline. The turbulent bottom layer \( (D_x) \)
is defined as the height above the bottom at which the
turbulence dissipation rate decreases to \( 6 \times 10^{-9} \) m\(^{2} \) s\(^{-1} \).
For further details and discussion of mixed, remnant and
turbulent layer definitions, see Perlin et al. [2005a].

We define the veering layer \( (D_v) \) as the layer that contains a
systematic counter-clockwise rotation of the current vector
with depth (see section 4).

[10] Neglecting time dependence, nonlinear accelerations
and baroclinic pressure gradients, the horizontal momentum
balances in the bottom Ekman layer are given by:

\[
-\frac{f(v_e - v)}{\partial x} = \frac{\partial}{\partial z} \left( K_v \frac{\partial u_e}{\partial z} \right) \quad (1)
\]

\[
\tau = \frac{\partial}{\partial z} \left( K_v \frac{\partial v_e}{\partial z} \right). \quad (2)
\]

where the interior geostrophic flow \( v \) is in the direction of
the \( y \)-coordinate, \((u_e, v_e)\) are velocity components inside
the Ekman layer, the stress is parameterized by an eddy
viscosity \( K_v \), and \( z \) is the height above the bottom. The
boundary conditions are

\[
u_e \to 0, \quad v_e \to \bar{v}, \quad \text{as} \quad z \to \infty \quad (3)
\]

\[
u_e = 0, \quad v_e = 0, \quad \text{at} \quad z = z_0, \quad (4)
\]

where \( z_0 \) is a constant of integration (frequently termed a
roughness length) and is determined when the system of
equations (1)--(4) is solved given a specified \( K_v(z), v \), and
bottom stress.

[11] We consider two possibilities for the specification
of \( K_v(z) \). First we consider \( K_v \) given by the law-of-the-wall,

\[
K_v = u_u \ell, \quad \ell = nz, \quad (5)
\]

where \( \kappa \) is von Kármán’s constant (=0.4), and \( \ell \) is a
turbulence length scale which increases without bound as \( z \)
goes to infinity. The analytical solution of (1)--(5) is given
by CRM. Second, we use the modified law-of-the-wall,
where a stratified boundary layer has an outer length scale
\( h_d \) that limits the growth of \( \ell \). The MLW eddy viscosity is
given by:

\[
K_v = u_u \ell, \quad \ell = \kappa (1 - z/h_d), \quad (6)
\]

for \( z \leq h_d \) and \( K_v = 0 \) for \( z > h_d \). The value of \( h_d \) is chosen to
be the largest value such that \( \ell(z) \) \( \leq \ell_o(z) \) at all depths,
where the effect of stratification limiting the scale of
turbulence is represented by \( \ell_o = \sqrt{\varepsilon/N^2} \), the Ozmidov
scale, \( \varepsilon \) is the turbulence dissipation rate and \( N \) is the
buoyancy frequency at the top of the mixed layer [Perlin et
al., 2005b]. Since \( \ell_o \to \infty \) within the mixed layer \( D \), the
value of \( h_d \) is determined by the vertical profile of \( \ell_o \) just
above \( D \). Based on examination of many profiles on the
Oregon shelf, we find \( h_d \approx D^2/(D - 1) \) to be a good
approximation [Perlin et al., 2005b].

3. Overview of the Data

[12] The observations used in this study come from a
three-month long mooring deployment and a 50-h vertical
profiling time series over the Oregon shelf during the
summer upwelling season of 2001, as part of the Coastal
Ocean Advances in Shelf Transport (COAST) program
[Barth and Wheeler, 2005]. The mooring was deployed in
81 m of water (45° 0.01' N, 124° 7.00' W) directly offshore
(west) of Cascade Head [Boyd et al., 2002]. Currents were
observed with two acoustic Doppler profilers: an upward-
looking RDI 300 kHz ADCP 4 m above the bottom with 2 m
vertical resolution, and a downward-looking Nortek 2 MHz

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Aquadopp profiler 9 m above the bottom with 0.5 m vertical resolution. Due to side-lobe interference and acoustic spreading, there are no reliable velocity data within 2.75 m of the bottom.

The moored velocity data contain tidal and inertial oscillations. To isolate the low frequency characteristics of the BBL, daily averages of the 3-month data set were computed (Figure 2). Daily averages do not differ significantly from the 25-h low-passed filtered velocity (Chebychev, fourth order). Filtered velocity gives very similar results if used for further analysis. We opted to use daily averages for convenience. The flow has been predominantly southward (upwelling favorable), but with occasional relaxations, and even a large reversal in late June (Figure 2c). Data included in the analysis that follows come from only those daily averages during which there were no alongshore flow reversals. Since Ekman transport calculations are sensitive to erratic veering angles at low speeds, all profiles with a speed smaller than 0.05 m s\(^{-1}\) at 8 m above the bottom have been excluded from the analysis.

The turbulence profiler Chameleon was used to obtain a 50-h time series near the mooring, measuring turbulence dissipation rate, temperature and conductivity. A detailed description of Chameleon and the procedures used to process the data can be found in Moum et al. [1995]. Chameleon has been routinely run into the bottom, permitting profiles to within 2 cm of the seabed, which is a necessary condition for estimating bottom stress [Perlin et al., 2005b]. The time series starts during a period of high southward flow (Figure 3d) that has reversed 36 h into the observation. The mixed layer (\(D\)) was thicker (~20 m) during the strong flow conditions, thinning to less than 10 m as the flow slowed (Figure 3a). The turbulence was also larger during the high flow, and the thickness of the turbulent layer (\(D_\epsilon\)) roughly followed the mixed layer thickness (Figure 3b).

To relate the observations to the momentum balance (1)–(4), two quantities are needed: \(u_\ast\) and \(v_\ast\). \(u_\ast\) has been estimated directly from the profiles of turbulent dissipation \(\varepsilon(z)\) using the dissipation method \(u_\ast = (\varepsilon \kappa z)^{1/3}\). This method for computing the friction velocity is described by Dewey and

![Figure 2.](image-url)
Crawford [1988] and discussed by Perlin et al. [2005b] for this same data set. Direct estimates of $u_*$ cannot be made from the mooring data; however, two indirect estimates can be made based on the velocity time series. The first method is based on the quadratic drag approximation, $u_{*20} = \sqrt{C_D U_{20}^2}$, where $U_{20}$ is the velocity measured 20 m above the bottom. Perlin et al. [2005b] have found that for this shelf as a whole $C_D$ varies between $4.5 \times 10^{-4}$ and $1.6 \times 10^{-3}$,

Figure 3. Data from 50-h Chameleon time series collected near the mooring showing (a) 1-h averaged density (white lines mark mixed and remnant layer heights), and (b) 1-h averaged turbulence dissipation rate (white line marks turbulent layer height). Data from the mooring show (c) low-passed (second-order Butterworth 21h) eastward velocity (gray line marks veering layer height) and (d) low-passed (second-order Butterworth 21h) northward velocity (white line marks veering layer height). Vertical line marks the time of the profile shown in Figure 6.
with an average of $10^{-3}$. The second method assumes that $\tau_b \sim \tau_{y_b}$ and estimates $u_{\theta_f}$ from the Ekman transport normal to the interior geostrophic flow $\tau$ as $u_{\theta_f} \sim \sqrt{\tau_{y_b}/\rho} = \sqrt{\sqrt{\overline{\epsilon}} f}$, where $\sqrt{\overline{\epsilon}} = \int u, dz$. Note that since there are no data below 2.75 m, we calculate the integral using extrapolated $u_0(z)$ based on the law-of-the-wall. The two estimates of $u_{\theta_f}$ are well correlated ($r(u_{\theta_f}/u_{20}) = 0.8(0.65 - 0.89)$) with $u_{20} = 0.98(0.89 - 1.09)$ (Figure 4).

The interior geostrophic velocity $\tau$ is taken to be the velocity observed from the mooring at the top of the veering layer at height $D_v$ (defined in the next section) and hence $u_0(D_v) = 0$. Note that the direction of $\tau$ varies slowly as a function of time, although it is most often aligned in a nearly north-south direction.

### 4. Veering Layer

Owing to the Earth’s rotation, the current veers in a counter-clockwise direction as the bottom is approached from the geostrophic interior flow. The veering is defined to be contained within the veering layer $D_v$. The challenge is to distinguish between the veering which is induced by the presence of the bottom boundary and the veering that may be caused by other motions in the interior based on the velocity observations. To accomplish this, the veering layer thickness $D_v$ has been estimated as follows. The upper limit has been chosen at the height $h_{\text{max}} = 0.6 u_{20}$ above the bottom. The veering rate has been calculated as the change in the angle over a 5 meter interval from $h$ to $(h - 5)$ m. If at a height $h$ the veering rate exceeds the threshold value of $6/h$ (deg/m), we set $D_v = h$. If not, then the veering rate over 5 m is repeatedly calculated at successively deeper depths. The threshold has been set such that the veering rate at $z = D_v$ is large enough to result in at least 6 degrees of veering when extrapolated from $D_v$ to the sea floor. This definition was based on extensive analysis of the individual veering angle profiles. The threshold was defined to be small enough to identify veering layers with smaller than average rotation, but large enough to exclude random variation in direction of the velocity above the Ekman layer. Varying the threshold value from $1/h$ to $11/h$ (deg/m) leads to at most a 1 m root mean square (RMS) difference in the height of the veering layer (Figure 5), which is only a 20% difference for the thinnest veering layers.

An example of the veering of the velocity in a single profile is shown in Figure 6a. The interior current is slightly above $0.2 \text{ m s}^{-1}$ and flows 185 degrees clockwise from north. Counter-clockwise veering starts approximately at $D_v = 19$ m and reaches $\sim 12^\circ$ at 2.75 m above the bottom (below which we have no velocity measurements). The turbulence is high near the bottom (Figure 6b), and the water is well-mixed (Figure 6c). The veering, in this example, starts at the depth at which the turbulence is high.

It is only during the 50-h Chameleon time series that we can obtain estimates of $D_v$, $D_e$, and $D_b$. During this period, the veering height $D_v$ follows the turbulent layer thickness $D_r$ (Figure 7). Also, $D_r \approx 2D_e$, $D_r \approx 2/3D_v$, $D_v$ compares to the Ekman length scale for neutrally stratified boundary layers $u_{20}/f$ with coefficient of proportionality $\gamma \approx 0.26$, and the stratified Ekman length scale $u_{20}/\sqrt{f}$ with coefficient of proportionality $\beta \approx 1.6$ (Pollard et al. [1973] suggested $\beta = 2^{1/4} \approx 1.7$). In the case of the latter scaling, the stratification in the weakly stratified layer above $D$ has been used (if stratification over $D_v$ were applied, $\beta \approx 1.5$).

Stratification can suppress turbulence on a shorter than tidal time scale, which would affect the Ekman spiral, as has been shown by Price and Sundermeier [1999], and Weller and Plueddemann [1996]. It is extremely difficult however to decompose very noisy unfiltered data into flow related to internal tides and to Ekman dynamics in order to...
select $D_v$. We calculated $D_v$ from filtered velocity profiles. Hence, hourly $D_v$ values cannot be considered to be independent. Since the number of degrees of freedom for $D_v$ is very small, quantifying its correlation with other boundary layer scales is problematic.

We extend the analysis over the 3-month mooring record to examine the consistency of both the veering angle and the cross-axis velocity ($u_e$) with current direction, speed and veering layer thickness. Since there were no turbulence observations from which to obtain a direct estimate of $u_e^*$, the indirect estimates of bottom stress, $u_e^* t$ and $u_e^* 20$, have been used (see section 2). The veering thickness $D_v$ has been found to be highly correlated with both estimates of $u_e^*$ such that $r_{D_v/u_e^* 20} = 0.84(0.71 - 0.91)$ and $r_{D_v/u_e^* t} = 0.9(0.81 - 0.95)$ with linear regression $D_v \approx 0.36 u_e^* 20/f$ and mean ratio $D_v/(u_e^* t) \approx 0.34(0.31 - 0.37)$ and $D_v/(u_e^*/f) \approx 0.39(0.35 - 0.44)$ (Figure 8). The proportionality constant of 0.35 is slightly higher than 0.26 which appears to be a good fit over the 50 h period (Figure 7) when $u_e/f$ can be estimated directly from observed dissipation (see open circles in Figure 8 for the times corresponding to the time series).

To study the observed veering angles of profiles made at a variety of flow conditions and different values of $D_v$ it is useful to define the average veering angle by:

$$\alpha_{av}(z/D_v) = \left< \tan^{-1} \frac{u_e(z)}{v_e(D_v)} \right> \quad (7)$$

where the operator $<\cdots>$ is the arithmetic average of the angles at height $z$. This quantity is sensitive to erratic veering angles at low speeds, so flows with speed less than 0.05 ms$^{-1}$ measured 8 m above the bottom are not considered. This averaging is different from a true vector average. However, at speeds above 0.05 ms$^{-1}$, the methods do not differ significantly.

The average veering angle profile $\alpha_{av}(z/D_v)$ is remarkably constant for the duration of the 3-month time series (Figure 9). Since the daily average flow is nearly parallel to the coast, the observations can be divided into average flows that are either generally southward or northward with corresponding upslope or downslope Ekman flow. Shutdown time for our Ekman flow computed as in Garrett et al. [1993] is of the order of hours to a day, implying differences between the upslope and downslope Ekman flows. We have however demonstrated [Perlin et al., 2005a] that upwelling in the BBL could continue for a few days without being arrested, and that a simple 2D Ekman balance is roughly satisfied in the BBL at this location on the Oregon shelf. Our observations of Ekman veering confirm that finding: the difference in average veering angle between southward flow (upslope Ekman flow) and northward flow (downslope Ekman flow) is statistically insignificant (Figure 9a). Also, there are no appreciable differences in $\alpha_{av}$ between thick ($D_v > 10$ m) and thin ($D_v < 10$ m) veering layers (Figure 9b). Differences in $\alpha_{av}$ between high flow ($U_{20} > 15$ cm/s) and low flow

Figure 6. An example Chameleon profile on 8 August 2001, 23:00 showing (a) profiles of velocity magnitude (gray) and direction (black) (veering layer height is marked at 19.5 m), (b) turbulent dissipation rate (turbulent layer height is marked at 18.5 m), and (c) density profile (mixed layer height is marked at 13 m and remnant layer height is marked at 32 m).
(\(U_{20} < 15\) cm/s, Figure 9c) are statistically insignificant at the 95% confidence limit.

5. Comparison of Observations and Theory

[24] To illustrate the comparison between the observations and the model, a single observed velocity profile is plotted with the theoretical profiles using the viscosity parameterization of the law-of-the-wall (CRM) and the modified law-of-the-wall (MLW) (Figure 10). Both models do not include any fitting parameters. It has been assumed that the entire Ekman transport is contained below \(z = D_v\).

The values of \(u_*\) and \(h_d\) used in the model have been computed from the turbulence measurements [see Perlin et al., 2005b]. The value of \(h_d\) in this example is almost equal to \(D_v\); turbulent viscosity in the MLW model drops to zero at \(z = h_d\) and velocity profiles are not defined above that height. Velocity atop the veering layer is used in place of \(\bar{u}'\) in the model.

[25] The effect of the stratification in suppressing turbulence at the top of the BBL is clear in the MLW solution where the \(u\)-momentum is constrained below \(h_d\). The CRM solution is characterized by weaker \(u_*\) and smaller veering angles, but the \(u\)-velocity penetrates far above the seafloor. One outcome of the modified law-of-the-wall [Perlin et al., 2005b] has been the good agreement between the modeled and the measured profiles of \(\varepsilon(z)\) in the lower 60–70% of the mixed layer. This represents an improvement over the traditional law-of-the-wall which was only good in the lower 20–30% [Perlin et al., 2005b]. In agreement with that result, the velocity and the veering profiles compare with the data relatively well below 0.7\(D_v\), but disagree in the upper portion of the mixed layer. The MLW model is not viewed as a complete physical model, but rather as a test of the included physics (particularly, the effect of the stratification cap on the Ekman solution). Its applicability is limited by unresolved processes near the top of the mixed layer (like internal waves and advection). Controlled experiment is required to test the turbulent viscosity profile in the upper part of the bottom boundary layer, and our field data do not satisfy the necessary requirements for the controlled experiment. We have plotted an additional solution (marked MLW2), in which \(K_v\) does not drop to zero at \(h_d\) but remains at a small value \(K_v = 10^{-3} m^2 s^{-1}\) through the rest of the water column. This value is based on our estimation (from turbulence measurements) of mean \(K_v\) in the interior.

Non-zero value of \(K_v\) above \(h_d\) allows the veering to penetrate higher in the water column, which results in a thicker veering layer. This solution does not give much improvement over the original MLW and the tested parameterization of \(K_v\) in the interior does not reflect any real
physics. For further analysis we opted for a zero $K_v$ above $h_d$.  

[26] Note that all model solutions (CRM, MLW and MLW2) have the same transport perpendicular to $\bar{v}$ as required by the integral of (2), which in this example is not identical to the measured transport. Mean values of the friction velocities computed from the transport and from the turbulence data agree statistically and are highly correlated [Perlin et al., 2005a, Figure 14], while they could differ for a particular profile.

[27] We have then compared the entire 50-h Chameleon time series observations with the MLW model (Figure 11). We have not plotted the model solution above 0.7$D_v$ since we do not expect to see the agreement with the data in that region. During the 50-h time series the turbulence measurements have been used to estimate $u_\ast$ and $h_d$ directly. As previously discussed, for the purpose of data analysis we have assumed that the entire transport perpendicular to $\bar{v} = v(D_v)$ occurs below $z = D_v$. MLW provides a relatively good fit in the lower half of the boundary layer and predicts magnitudes of $u_e$ and veering angles.

Figure 8. Veering layer height from moored velocity is plotted as a function of friction velocity (a) $u_{\ast20}$ and (b) $u_{\ast t}$. Uncertainties for the veering layer height are based on 4° and 8° thresholds. Linear regression is shown by dashed line. 95% confidence limits are shown in brackets. Gray dots are the data from the 50-h time series (Figure 7).

Figure 9. Average veering angle profiles $\alpha_{ve}(z/D_v)$ (with 95% confidence limits) versus nondimensionalized height above the bottom. (a) Veering angle for northward (dark gray) and southward (light gray) flow; (b) veering angle for thick (>10 m, dark gray) and thin (<10 m, light gray) veering layers; (c) veering angle for fast current ($U_{20} > 15$ cm s$^{-1}$, dark gray) and slow current ($U_{20} < 15$ cm s$^{-1}$, light gray). The veering angles for thin veering layers and for slow velocities on (b) and (c) are truncated due to shadow zone of ADCP combined with normalization of $z$ by $D_v$. 

[28] Comparison of the moored observations and the MLW model for the entire 3-month time series has shown similar results (Figure 12). Here $u_{\ast20}$ has been estimated from a drag coefficient $C_D = 10^{-3}$ and observed $U_{20}$ [see Perlin et al., 2005b, Figure 4]. To apply the MLW model, we have assumed that $h_d = D_v$. Again, the MLW model
shows reasonable agreement with the data, slightly underpredicting the veering angle in the lower part of the profile.

While the gross characteristics of the observations are captured by MLW, some of the comparisons are obscured by the averaging in Figures 11 and 12. The maximum veering angle in the MLW model depends on $\nu$, so that the greater the interior current the stronger the veering. The data are too limited to demonstrate this relationship in any meaningful way.

6. Summary

Observations reveal a systematic counter-clockwise rotation with depth of the daily averaged current over three summer months on the Oregon shelf, except during periods of low speed or immediately following a reversal of alongshore velocity (Figure 9). The velocity profile near the bottom shows the maximum veering of $12^\circ$ (Figure 9). This is much less than the $45^\circ$ of the classic Ekman solution (constant $K_v$), but is in the range of the other BBL measurements (Table 1). The maximum veering angle has only a very weak dependence on the interior flow velocity and the thickness of the veering layer. The thickness of the veering layer $D_v$ does not appear to depend on the flow direction, regardless of whether the Ekman transport is upslope for southward interior flow, or downslope for northward interior flow (Figure 9).

For a 50-h segment during which our measurements of the velocity were supplemented by intensive turbulence and density profiles, we have found that the veering layer height is nearly equal to the turbulent boundary layer height. This is consistent with the result described by Perlin et al. [2005a], who have shown that the cross-shore movement of the fluid in the BBL is well-modeled by the measured friction velocity ($u_\tau$) and the momentum distributed over...
a turbulent layer height, $D_v$. This and the high near-bottom velocities presented here clearly demonstrate that the Ekman spiral is confined to the turbulent layer near the seafloor in a stratified ocean. We also expect that the thickness of the bottom turbulent layer depends on the time history of the turbulence, the background stratification, and the speed with which fluid mixed in the BBL is flushed either by geostrophic adjustment or by ambient currents. For instance, Moum et al. (2004) have found that the turbulence in the boundary layer transitions from being driven by stress during upwelling, to being driven by convection during downwelling. It seems unlikely that the thickness of the veering layer would scale in the same way for both regimes.

[32] The veering layer apparently is further constrained to a bottom Ekman layer on a continental shelf, where it requires the turbulent viscosity drop to zero. If the classic law-of-the-wall (CRM) is used for the turbulent viscosity, we find that the Ekman transport is distributed over a much thicker layer than in the observations, and the veering of the velocity near the bottom is underestimated (Figure 10).

[33] The Ekman lengths scales $u_{*}/f$ and $u_{*}/\sqrt{N}$ appear to relate to $D_v$ within accepted constants (Figure 7 and Pollard et al. [1973], respectively). Our measurements do not permit an evaluation of the influence of the latitude changes on the Ekman layer height. The model suggests dependence of layer height on rotation. However, the presence of stratification, presumably imposed by unrelated mesoscale processes, may lock the bottom Ekman layer in a thin boundary layer even at low latitudes. The analysis of the effect of rotation in the presence of stratification seems an important problem to address.

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