The Subinertial Momentum Balance of the North Atlantic Subtropical Convergence Zone

CRAIG M. LEE
Department of Physical Oceanography, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

CHARLES C. ERIKSEN
School of Oceanography, University of Washington, Seattle, Washington

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ABSTRACT

The upper-ocean response to forcing by pressure gradients and wind stress is examined using observations from the Frontal Air–Sea Interaction Experiment. A moored array acquired time series of winds, upper-ocean currents, temperatures, and salinities between winter and late spring in a region of the Sargasso Sea known for the presence of upper-ocean fronts. These fronts have timescales of 10 days and dominate current variance, while winds varied with the 4-day timescale of passing weather systems. Employing a frequency domain regression model, it is found that geostrophy accounts for most of the low-frequency (>100 h) current variance in the seasonal pycnocline, but wind-forced shear becomes important nearer the surface. In particular, currents oriented in the typical NE–SW alongfront direction display geostrophic balance, while those perpendicular to them do not.

Wind forcing can produce geostrophic currents indirectly through Ekman pumping, and knowledge of the geostrophic shear is required to distinguish between this and currents driven directly by the wind through turbulent shear stress. Previous investigations rely on the assumption that no wind-driven stress penetrates below the mixed layer to remove the wind-coherent geostrophic flow. Baroclinic pressure gradients are calculated using estimates of density across the moored array. A linear regression model uses the pressure gradient record to explicitly remove the geostrophic shear and isolate the directly wind-driven acceleration at timescales longer than 10 days. The resulting response satisfies the Ekman transport relation, penetrates well into the stratified fluid, spirals to the right, and decays with depth.

1. Introduction

Ekman transport (Ekman 1905) is an important component of the long timescale circulation, which has proven difficult to verify observationally. The small size of directly wind-driven currents relative to other sources of upper-ocean variability, such as fronts and mesoscale eddies, hampers attempts to quantify the transport. Recent observational studies of subinertial wind-driven upper-ocean dynamics (Davis et al. 1981a,b; Price et al. 1987; Weller et al. 1991) focus on a one-dimensional model balancing acceleration and turbulent momentum flux. These investigations attempt to statistically isolate the response to wind-forced stress convergence from other sources of upper-ocean shear. However, at long timescales, Ekman pumping drives wind-coherent currents indirectly through pressure gradients, making it difficult to identify the Ekman transport. Knowledge of the geostrophic flow can be an important tool in distinguishing between direct wind-forcing through turbulent momentum flux and indirect driving through Ekman pumping. The goal of this study is to provide a complete description of the processes important to subinertial upper-ocean dynamics in the North Atlantic subtropical convergence zone.

We use observations from the Frontal Air–Sea Interaction Experiment (FASINEX) to explore the roles of geostrophy and wind-driving in the upper 200 m of the water column. FASINEX took place in a region marked by the presence of strong upper-ocean fronts. One of the goals of FASINEX was to examine the influence of strong horizontal gradients on upper-ocean dynamics. Semigeostrophic frontogenesis theories (Hoskins and Bretherton 1972; Hoskins 1975; MacVean and Woods 1990) invoke geostrophic balance for alongfront currents, but retain a full set of dynamics for acrossfront flows. Eriksson et al. (1991) qualitatively examine geostrophy during FASINEX, and Rudnick and Weller (1993a) find vertical integrals of 48-h lowpass velocity shear and horizontal temperature gradient to be orthogonal and well correlated, consistent with geostrophic flow at long timescales. Here we test the applicability of the thermal wind balance as a function of depth and frequency and examine the dynamics of the North Atlantic STCZ for front-related anisotropy.

Corresponding author address: Dr. Craig M. Lee, Department of Physical Oceanography, MS-21, Woods Hole Oceanographic Institution, Woods Hole, MA 02543.

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Two approaches have been used in previous attempts to directly examine wind-driven upper-ocean currents. Davis et al. (1981b), Price et al. (1987), Schudlich and Price (1995, manuscript submitted to J. Phys. Oceanogr.) and Weller et al. (1991) use moored observations of winds and ocean currents to test a one-dimensional momentum balance. Long records allow a statistical separation of wind-coherent currents, but the resulting response is too large and penetrates too deep to agree with theories of directly wind-driven currents (Weller et al. 1991). To correct for this, they assume that directly wind-driven currents are more surface intensified than geostrophic currents and remove the response slightly deeper than the average mixed layer depth from that shallower. The mixed layer referenced response agrees with directly wind-driven dynamics, but the results are sensitive to the choice of reference depth. This method assumes turbulent momentum flux does not penetrate too deeply and precludes testing the vertical extent of the response. Weller et al. speculate that the strong, deep-penetrating response they remove may be due to geostrophic currents that are coherent with local winds.

Chereskin and Roemmich (1991) and Wijffels et al. (1994) provide an alternative approach for investigating wind-driven upper-ocean dynamics. They calculate geostrophic shear using CTD data taken along a zonal transect and obtain observations of the total upper-ocean shear using an Acoustic Doppler Current Profiler (ADCP). Differencing the two provides an estimate of the ageostrophic shear, which they integrate in the vertical to estimate the meridional Ekman transport. The primary assumption is that ageostrophic flow vanishes far below the base of the mixed layer. The results compare favorably with transport estimates obtained using shipboard and climatological winds. Geostrophic currents are explicitly removed and the results appear insensitive to the choice of the deep reference level.

We examine the upper-ocean response to wind-forcing by extending the univariate statistical models used in previous studies (Davis et al. 1981b; Price et al. 1987; Schudlich and Price 1996, manuscript submitted to J. Phys. Oceanogr.; Weller et al. 1991) to incorporate direct estimates of the geostrophic shear. The FASINEX moored array allows us to estimate baroclinic pressure gradients and explicitly remove the geostrophic currents. This is critical for identifying the response to wind-driven turbulent momentum flux, involving assumptions similar to those made by Chereskin and Roemmich (1991). The present analysis extends the results of Weller et al. (1991) by considering pressure gradient forcing and horizontal momentum advection. We identify directly wind-driven currents penetrating well below the mixed layer base and demonstrate that the anomalous wind-coherent flow reported previously is a pressure-coupled response.

The FASINEX mooring data are described in section 2. Section 3 begins with a careful examination of the terms of the horizontal momentum equations, followed by a study of the geostrophic balance in the upper ocean. After a brief review of Ekman theory and Weller et al.'s results, we quantify the geostrophic currents and isolate the upper-ocean response to wind-driven turbulent momentum flux. We discuss our results and examine their implications in section 4.

2. Data

We use moored observations made near 27°N, 70°W between January and June 1986 (Weller et al. 1990a,b). The FASINEX moored array was set to resolve the scales of frontal variability in the North Atlantic subtropical convergence zone (STCZ) and facilitate the study of air–sea interaction in a region of strong horizontal gradients. Typical horizontal separation is 20 km, with a maximum separation of about 60 km (Fig. 1). Three profiling current meters (PCMs), F3, F5, and F7, provide time series of velocity, temperature, conductivity, and pressure at 5-m intervals between 40 and 180 m. The PCMs performed 20-minute ascent profiles at 4-hour intervals for the duration of the experiment. Three surface moorings, F4, F6, and F8 sampled velocity and temperature at depths of 10, 20, 30, 40, 80, 120, and 160 m using a combination of vector-measuring (VMCM) and vector-averaging (VACM) current meters. Instruments at 20 and 120 m (F4), 40 m (F6), and 120 m (F8) failed to return us-

![Fig. 1. Plan view of the FASINEX moored array. Squares mark Profiling Current Meters (F3, F5, and F7) and circles mark surface moorings (F2, F4, F6, F8, and F10). Separations between moorings are marked in km (after Weller et al. 1991). This study focuses on observations from moorings F3, F4, F5, F6, F7, and F8.](image-url)
able records. We calculate wind stress from meteorological data and sea surface temperatures collected by vector-averaging wind recorders (VAWRs) mounted on the surface buoys. Hourly-averaged surface mooring data are further averaged into 4-hour bins corresponding to PCM profile times when performing calculations involving the entire array. Moorings F2 and F10 provide additional data used solely to reconstruct the results of Weller et al. (1991).

Differences in the velocity response characteristics between instrument types required correcting for the various sensors. Following Weller et al. (1990b), we scale VACM velocities by 0.9 to correct for the over-response of these instruments when deployed from surface moorings. PCM tow tank calibrations (J. Dahlen 1989, personal communication) and potential flow calculations indicate that the spherical housing of the instrument accelerates flow over the velocity sensors. We apply a corrective gain of 0.96 to PCM velocity records.

3. Analysis and results

We examine elements of the subinertial momentum balance with the goal of identifying the processes important to upper-ocean dynamics in the subtropical convergence zone. The horizontal momentum equations for the upper ocean may be written as

\[
\begin{align*}
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - f v &= -\frac{1}{\rho_0} \frac{\partial p}{\partial x} + \frac{1}{\rho_0} \frac{\partial \tau^x}{\partial x} \quad (1) \\
\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + f u &= -\frac{1}{\rho_0} \frac{\partial p}{\partial y} + \frac{1}{\rho_0} \frac{\partial \tau^y}{\partial y} \quad (2)
\end{align*}
\]

(a) (b) (c) (d) (e) (f)

Here \((u, v, w)\) are eastward, northward, and upward velocities; \(f\) the inertial frequency, \(\rho_0\) the mean density, \(p\) the perturbation pressure, and \((\tau^x, \tau^y)\) are eastward and northward stress. We begin by examining various terms of (1) and (2) and testing the quasigeostrophic balance \((a + d = e)\) in the presence of strong upper-ocean fronts. A review of the wind-driven balance between geocentric acceleration and the vertical convergence of stress \((a + d = f)\) follows. Previous investigators (Davis et al. 1981b; Price et al. 1987; Chereskin and Roemmich 1991; Weller et al. 1991; Wijffels et al. 1994) have explored this balance and find strong currents coherent with local wind stress, but extending far beyond the penetration depth of the wind-driven turbulent momentum flux. Weller et al. (1991) suggest that these currents may be driven indirectly by the wind through pressure gradient forcing, that is, Ekman pumping. The primary focus of this analysis is an examination of the full linear momentum balance. Here geocentric acceleration \((a + d)\) responds to both pressure gradient forcing \((e)\) and vertical convergence of stress \((f)\). We isolate both pressure and directly wind-driven flows, as well as an Ekman pumping response driven indirectly by the wind through pressure gradients. One of the objectives of FASINEX was to examine the influence of the strong horizontal gradients associated with the fronts on the upper-ocean response to forcing. The moored array offers a unique opportunity to investigate how nonlinear interactions modify the directly wind-driven response by adding horizontal momentum advection \((b)\) to the linear momentum balance. We neglect the vertical advection of momentum \((c)\).

a. Analysis methods: Elements of the horizontal momentum equations

We calculate surface wind stress for each surface mooring using the bulk formulas of Large and Pond (1981) and VAWR meteorological records. Surface stress shows virtually no horizontal variation over the span of the array larger than the estimated 12% observational error (Weller et al. 1990a). A plane fit to surface mooring winds provides an estimate of stress over the PCMs. The 2–4 day timescales of passing weather systems dominate wind stress variance (Fig. 2a). These systems moved eastward and northeastward to the north of the array, causing winds at the FASINEX site to rotate clockwise as they passed through. Rotary power spectra thus display a prominent peak in the clockwise weather band, with dominance of clockwise over anticlockwise energy on timescales longer than 2 days (Fig. 3a).

Time series of density profiles allow us to estimate the effects of pressure gradient forcing and illustrate changes in the background stratification. We calculate the density at each PCM mooring using profiles of conductivity, temperature, and pressure. Combining observed temperatures and daily-averaged PCM temperature–salinity curves, we estimate salinity and density at each surface mooring. Extrapolating the observed \(T–S\) curve to include near-surface temperatures introduces large errors, thus restricting density estimates to depths below 40 m. Rapidly rising and diving isopycnals mark frontal passages in early March, April, and May (Fig. 4). The average
mixed layer depth was approximately 40 m, though during frontal passages the mixed layer extended as deep as 100 m. Below 100 m, a sharp increase in stratification marks the top of the seasonal pycnocline. Through spring, stratification in the seasonal pycnocline weakens and that between 40 and 100 m strengthens.

We calculate geocentric acceleration relative to the deepest observation at each mooring as

$$A_{G0}(z, t) = \left( \frac{\partial}{\partial t} + i\omega \right)(U(z, t) - U(z_0, t)), \quad (3)$$

where $U = u + i\omega$ is the complex velocity, $z_0$ is the depth of the deepest instrument (-160 m for surface moorings and -180 m for PCM moorings) and centered first differences estimate the time derivative. Vector series of area-averaged geocentric acceleration (Fig. 2c) exhibit remarkable similarity to pressure-driven acceleration, closely mirroring the signature of the frontal jets. Free inertial oscillations have no geocentric acceleration and give rise to the spectral well around $\omega = -f$ (Fig. 3b), where $f = 6.621 \times 10^{-2}$ s$^{-1}$ and negative rotary frequencies indicate clockwise motions. Using geocentric acceleration filters out energetic free inertial oscillations.
and enhances our ability to identify the directly wind-driven flow.

Horizontal gradients of velocity and density are needed to calculate the momentum advection and pressure gradient terms in (1) and (2). Motions with inadequately resolved lateral scales will alias gradient estimates. Following Rudnick and Davis (1988), we use a truncated series of empirical orthogonal functions to describe the resolved portion of the velocity and density variance. The temporal mean and first three (four) EOFs form filtered velocity (density) fields retaining 84% (90%) of the original variance. The moored array was set to resolve frontal temperature scales. Other FARSINEX investigators (Pollard and Regier 1992) and a simple scaling of the thermal wind relation suggest that velocity should contain more small-scale variance than density. We use least squares plane fits to the filtered fields to obtain gradient estimates over the entire array.

Vertically integrating density gradients upward from the deepest observation at each mooring provides an estimate of the baroclinic pressure gradient,

\[ -\frac{1}{\rho_0} \nabla p(z, t) = -\frac{1}{\rho_0} (\nabla p(z, t) - \nabla p(z_0, t)) \]

\[ = \frac{R}{\rho_0} \int_{z_0}^{z} \nabla p(z', t) dz'. \]  \hspace{1cm} (4)

Here \( \rho \) is density, \( g \) gravity, \( \nabla = \partial/\partial x + i \partial/\partial y \) the horizontal gradient operator expressed in complex form; the hydrostatic balance has been assumed. Sharp density gradients accelerate strong, surface-intensified geostrophic flows, and intense pressure gradients (Fig. 2b) accompany the passage of fronts (Fig. 4). Pressure gradients are most energetic at the 10+ day timescale dominated by the fronts and show no preference for accelerating clockwise or anticlockwise flow.

We estimate horizontal momentum advection [term b in (1) and (2)] relative to the deepest observation at each mooring using the filtered velocity gradients. During frontal passages, Rossby numbers reach a few tenths. Horizontal momentum advection is largest at these times (Fig. 2d) and might be expected to play a role in the momentum balance. With typical frontal vertical shears of 0.005 s\(^{-1}\) and vertical velocities of 0.0002 m s\(^{-1}\) (Rudnick and Weller 1993a), vertical momentum advection scales as \( 1 \times 10^{-6} \) m s\(^{-2}\), a fifth the typical horizontal momentum advection accompanying the fronts. For comparison, in the center of the

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**Fig. 3.** Rotary power spectra of F4 (a) wind stress and (b) geocentric acceleration at 40 and 160 m. Solid (dashed) lines indicate anticlockwise (clockwise) motions. Spectra for the 160-m record are offset down one decade. A vertical line marks the inertial frequency in (b), and 95% confidence intervals are plotted beneath the spectra. Clockwise energy dominates the wind stress record at timescales longer than 2 days.
Gulf Stream estimates of rms vertical velocity \((0.0008 \text{ m s}^{-1})\) and vertical shear \((0.002 \text{ s}^{-1})\) (Bower and Rossby 1989) yield a similar value for vertical momentum advection. In the analysis that follows, vertical momentum advection may be comfortably neglected.

**b. Geostrophy**

Using PCM observations, we examine the degree to which geostrophy holds in a surface-intensified front. FASINEX fronts have typical mixed layer temperature changes of \(1^\circ\text{C}\) over 20 km and peak relative vorticities of \(\pm f/3\) (Eriksen et al. 1991). Although high Rossby numbers imply ageostrophic dynamics, semigeostrophic theory (Hoskins and Bretherton 1972; Hoskins 1975) suggests that, in the absence of curvature, geostrophy should be maintained for the alongfront current component.

Consider the "time-dependent" thermal wind relation,

\[
\left( \frac{\partial}{\partial t} + \mathbf{i} f \right) \left( \frac{\partial u}{\partial z} + \mathbf{i} \frac{\partial v}{\partial z} \right) = \frac{g}{\rho_0} \left( \frac{\partial p}{\partial x} + \mathbf{i} \frac{\partial p}{\partial y} \right),
\]

where we calculate time derivatives and vertical shears of horizontal velocity using centered first differences of area-averaged PCM velocities and approximate horizontal density gradients from plane fits to PCM densities. From (5), we derive the frequency domain regression model

\[
\frac{\partial}{\partial z} A_C(z, \omega) = \frac{\partial}{\partial z} \left( \frac{\partial}{\partial t} + \mathbf{i} f \right) U(z, \omega)
\]

\[
= H_{n}(z, \omega) \frac{g}{\rho_0} \nabla \rho(z, \omega) + n(z, \omega)
\]

to examine the validity of the quasigeostrophic balance. Here \(n(z, \omega)\) is uncorrelated noise and \(\omega\) is rotary frequency with negative (positive) frequency indicating clockwise (anticlockwise) motion. The complex transfer function
Fig. 5. Rotary cross spectra of $\partial A_\phi/\partial z$ and $(\eta \rho \nabla \rho$ from PCM observations. Coherence squared (skill), admittance magnitude and phase are plotted for negative (clockwise motions, left) and positive (anticlockwise motion, right) frequencies. Thick solid, dashed, and dotted lines correspond to depths of 50, 100, and 150 m, respectively. The thin solid line specifies the smallest skill significantly different from zero at 95% confidence. Skill and admittance magnitude of 1.0 and zero phase indicate perfect thermal wind balance. Observations agree best with time-dependent thermal wind at low frequencies in the seasonal pycnocline (100 and 150 m).

$$H_{nm}(z, \omega) = \frac{\langle \nabla \rho^* A_{\phi} \rangle}{\langle \nabla \rho^* \nabla \rho \rangle} \tag{7}$$

is unitless and describes the response of the shears of geocentric acceleration to the horizontal density gradients, where angle brackets indicate an ensemble average over frequency bands and data segments and asterisks denote complex conjugation. The squared coherence, or skill, is the fraction of the density gradient variance described by the velocity shears. If the dynamics were entirely quasigeostrophic, skill and transfer function magnitude would be 1.0 and transfer function phase would be 0.0.

The time-dependent thermal wind balance holds best at low frequencies within the seasonal pycnocline. We find significant coherence for $-0.009 \text{ cph} < \omega < 0.009 \text{ cph}$, with skill generally increasing with increasing depth and decreasing frequency (Fig. 5). At 50 m, observed skill is only significantly different from zero at 95% confidence for $-0.002 \text{ cph} < \omega < 0.005 \text{ cph}$. Transfer function phase is essentially zero, while magnitude ranges from 0.75 to 0.9. Skill is lowest near the surface as wind-forced turbulent momentum flux contributes variance that geostrophy cannot explain. Thermal wind accounts for approximately 70% of the 100-m density gradient variance for $-0.002 \text{ cph} < \omega < 0.002 \text{ cph}$, but fails at higher frequencies. Within the seasonal pycnocline, we find significantly nonzero coherence for $-0.005 \text{ cph} < \omega < 0.009 \text{ cph}$. Skills range between 0.9 and 0.35, with transfer function magnitudes between 1.0 and 0.4 and essentially zero phase.

The presence of strong upper-ocean fronts with a preferred NE–SW orientation (Erikson et al. 1991) suggests that there should be horizontal anisotropy in the geostrophic balance. Principal axes for spectra of vertical shears of horizontal velocity and of density gradients (not shown) indicate a preference for flows aligned NE–SW, but significant variance resides in flows perpendicular to this orientation. We test the directionality of the thermal wind relation using a rectilinear model in a rotated coordinate frame,

$$f \frac{\partial}{\partial z} u'(z, \omega) = H_{nm}(z, \omega) \frac{\rho}{\rho_0} \frac{\partial}{\partial y} \rho(z, \omega) + n(z, \omega). \tag{8}$$

Here primes indicate an anticlockwise coordinate rotation by some angle $\theta$ (Fig. 6), and we test one component of the thermal wind balance to find a rotation for which it is optimally satisfied. A significant change in skill as a function of $\theta$ indicates anisotropy for motions with timescales longer than 10 days (Fig. 7). A distinct maxima in skill occurs when $x'$ is aligned with the preferred NE–SW alongfront direction noted by

Fig. 6. Definition sketch for testing thermal wind in a rotated coordinate frame. Anticlockwise rotation of the standard east–north coordinates $(x, y)$ by an angle $\theta$ defines the new coordinate system $(x', y')$. 
Eriksen et al. (1991). Conversely, when $x'$ is oriented acrossfront, we find significantly nonzero skill only in the seasonal pycnocline. Higher frequency motions show a similar pattern, but with greatly reduced skill. Skills less than unity result from a combination of ageostrophic dynamics and failure to properly resolve the scales of variability in the flow. Both observations and theory indicate that crossfront length scales should be shorter than alongfront scales. As array separations are comparable to the crossfront scales observed in SeaSoar/ADCP surveys of FASINEX fronts (Eriksen et al. 1991), alongfront gradients should be well resolved. These results imply that flows aligned NE–SW are in geostrophic balance, while those perpendicular to them are not.

c. **Ekman theory**

We begin our analysis of upper-ocean response to wind-forcing by reviewing the classic Ekman balance (Ekman 1905). Consider the one-dimensional horizontal momentum equation for an upper-ocean driven solely by stress divergence,

$$
\left( \frac{\partial}{\partial t} + if \right) U(z, t) = \frac{1}{\rho_0} \frac{\partial}{\partial z} \tau(z, t),
$$

where $\tau = \tau^x + i\tau^y$. For an ocean driven by surface wind stress $\tau_0$ with harmonic time dependence $\propto \exp(i\omega t)$, vertically integrating (9) yields the Ekman transport relation,

$$
\rho_0 \int_{-\infty}^{\infty} U(z, \omega) \, dz = -i \frac{\tau_0(\omega)}{f} \frac{\omega}{\rho_0}.
$$

For $\omega > -f$, this simply predicts depth-integrated transport to the right (left) of the applied wind stress in the Northern (Southern) Hemisphere. This is a robust result, independent of the parameterization of stress divergence.

Parameterizing mixing with a constant eddy viscosity (Ekman 1905; Faller and Kaylor 1969) yields

$$
\tilde{A}_G(z, \omega) = i(\omega + f)U(z, \omega) = K_z \frac{\partial^2}{\partial z^2} U(z, \omega),
$$

where $\tilde{A}_G$ is the model geocentric acceleration and $K_z$ is an eddy viscosity. The well-known solution to (11) is

$$
\tilde{A}_G(z, \omega) = A_0(\omega) \exp \left[ (1 \pm i) \left( \frac{|f + \omega|}{2K_z} \right)^{1/2} z \right]
$$

$$
A_0(\omega) = (1 \pm i) \frac{\tau_0(\omega)}{\rho_0} \left( \frac{|f + \omega|}{2K_z} \right)^{1/2}.
$$

Geocentric acceleration has the same sense of turning and decay with depth as velocity in the classic Ekman solution (Fig. 8), but surface acceleration is $\pi/4$ to the

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Fig. 8. Hodograph of the Ekman acceleration spiral (12) scaled by $D_{\rho} \rho_0 / \rho_0$. Hollow-tipped arrows mark the solution for $0 > z > -10D_\rho$ plotted at depth intervals of $D_\rho/2$ with the size of the heads decreasing with depth. The solid arrow is a unit vector in the direction of the real axis. The hodograph turns to the right and decays with depth, and may be interpreted as the acceleration response to a wind stress in the direction of the scale vector.
left of the surface stress. The e-folding depth $D_E = (2K_s/|\omega + f|)^{1/2}$ varies with frequency, with the deepest penetration found near the resonance at $\omega = -f$ and shoaling away from this frequency. The details of this smooth spiral depend on the parameterization of the turbulent flux, and we offer this solution as a starting point for discussing the observations rather than as a dynamical explanation.

d. The wind-driven momentum balance

Weller et al. (1991) examine the balance between geocentric acceleration and wind stress during FASINEX. Their results form the starting point of our analysis, and we briefly revisit them here to motivate the work which follows. Weller et al. (1991) choose to model the directly wind-driven acceleration as

$$\left( \frac{\partial}{\partial t} + if \right) U(z, \omega) = H_E(z, \omega) \frac{\tau_0(\omega)}{\rho_0} + n(z, \omega).$$

(14)

We define the wind-coherent acceleration as $\tilde{A}_G(z, \omega) = H_E(z, \omega) \tau_0(\omega)/\rho_0$, where $H_E(z, \omega)$ illustrates the vertical structure of stress convergence. If (11) completely described the dynamics, $H_E(z, \omega) = \rho_0 \tilde{A}_G(z, \omega)/\tau_0(\omega)$ from (12) and the response would be as pictured in Fig. 8. The success of this analysis rests on the degree to which the true dynamics are linear and on the pressure gradient forcing being incoherent with the local wind stress. Pressure gradients drive stronger currents than do turbulent momentum fluxes. Any coherence between wind and pressure will alias portions of the larger pressure-driven flow into the wind-driven response, swamping the signal of interest.

We reproduce the results of Weller et al. (1991) by calculating the transfer function between wind stress and geocentric acceleration averaged over surface moorings F2, F4, F6, F8, and F10 (Fig. 9). Hodographs of $H_E(z, \omega)$ illustrate the response in three frequency bands; $-0.028 \text{ cph} < \omega < -0.042 \text{ cph}$, corresponding to the clockwise weather band, $-0.042 \text{ cph} < \omega < 0.042 \text{ cph}$, the band dominated by frontal variability, and $0.042 \text{ cph} < \omega < 0.028 \text{ cph}$, the anticlockwise weather band. We plot one band in each column, where
the top two rows represent $H_{E}(z, \omega)$ at depths from 10–40 m and 80–160 m. The solid-tipped scale arrow indicates 0.02 m$^{-1}$ in the direction of the real axis and gray circles mark the size of a standard error. The transfer functions may be thought of as the upper-ocean response to a surface stress applied in the direction of the scale vector.

Substituting $A_{C}(z, \omega)$ into (9) and integrating vertically from a depth $z_{0}$ (below the influence of the turbulent momentum flux) to the surface provides a test of the Ekman transport relation,

$$
\int_{z_{0}}^{0} H_{E}(\zeta, \omega) d\zeta = 1. \tag{15}
$$

Satisfaction of (15) indicates that the model accounts for all the observed surface stress. The Ekman transport relation (15) is independent of the parameterization of mixing and provides a better measure of the model’s success than any specifics of the vertical structure.

To remove the pressure-coupled response isolated by the model, Weller et al. (1991) assume that currents at 80 m are entirely pressure-driven and that geostrophic currents above 80 m are vertically uniform. Subtracting the 80-m response from that above yields a mixed-layer-referenced transfer function, which is integrated from 45 m to the surface to test the Ekman transport relation (Fig. 9). The bottom row of hodographs represent the vertical integral of $H_{E}(z, \omega)$ calculated as described above, where a response identical to the solid scale vector indicates perfect satisfaction of the Ekman transport relation.

Both clockwise and anticlockwise weather band responses generally turn rightward and decay with depth, nearly vanishing by 80 m. Mixed-layer-referenced transfer functions are nearly identical to their unreferenced counterparts. Vertically integrated, mixed-layer-referenced transfer functions agree reasonably well with the Ekman transport relation (15). In contrast, the low-frequency response is very energetic and shows no clear sign of decay or spiraling with depth. The unreferenced transfer function is inconsistent with a directly wind-driven response. Nevertheless, the vertical integral of the mixed-layer-referenced transfer function is within two standard errors of that predicted by theory. Although Rudnick and Weller (1993b) demonstrate that mooring motion can generate a deep, wind-coherent response, this cannot explain the magnitude and phase of the low-frequency transfer function. We expect mooring motion to be significant only at high frequencies and to cause clockwise phase shift in the transfer function for $\omega > -f$, the opposite of that observed.

Previous investigators have found strong, deep currents coherent with local wind stress. During MILE, Davis et al. (1981b) identified a strong response similar in magnitude and phase to the low-frequency response reported here, but attributed it to lack of statistical reliability in their relatively short time series. Pressure coupled, wind-coherent currents presumably interfere with Price et al.’s (1987) efforts to isolate the directly wind-driven response, as they remove the flow at 50 m from that above before performing their analysis. Weller et al. (1991) found wind-coherent currents at 1000 m during FASTEX and speculated that these might be pressure-coupled flows driven indirectly by the wind through Ekman pumping. In the sections that follow, we present evidence to support this hypothesis.

e. The linear momentum balance

Difficulties in observing the Ekman balance may be attributed to the weakness of directly wind-driven currents relative to those driven by pressure gradients and inertial oscillations and to problems associated with distinguishing between wind-driving through turbulent shear stress and indirect driving by pressure gradient forcing. Ekman pumping can couple pressure gradients with both local and remote wind stress. Lack of pressure gradient information forced previous investigators to make significant assumptions concerning the geostrophic flow and the penetration depth of direct wind-forcing to filter the pressure-coupled response. The FASTEX moored array allows us to estimate low-frequency pressure gradient acceleration and relax these assumptions by considering a more complete set of dynamics. We restrict this analysis to $-0.0042 \text{ cph} < \omega < 0.0042 \text{ cph}$, where the one-dimensional momentum balance (Weller et al. 1991) saw limited success and where section 3b suggests that we can reliably identify the geostrophic flow.

Consider the linearized horizontal momentum equation

$$
\left( \frac{\partial}{\partial t} + if \right) U(z, t) = - \frac{1}{\rho_{0}} \nabla p(z, t) + \frac{1}{\rho_{0}} \frac{\partial}{\partial z} \tau(z, t). \tag{16}
$$

We examine (16) using the two-input regression model

$$
A_{\omega_{0}}(z, \omega) = \frac{1}{\rho_{0}} \left[ -H_{p}(z, \omega) \nabla p_{m}(z, \omega) \right.
$$

$$
+ H_{p}(z, \omega) \tau_{0}(\omega) \big] \big] + n(z, \omega), \tag{17}
$$

where the transfer functions representing pressure gradient forcing and wind forcing are

$$
H_{p}(z, \omega) = \frac{\langle \nabla p_{m} \ast A_{\omega_{0}} \rangle \left( 1 - \frac{\langle \tau_{0} \ast A_{\omega_{0}} \rangle \langle \nabla p_{m} \ast \tau_{0} \rangle \rangle \langle \nabla p_{m} \ast \nabla p_{m} \rangle (1 - \gamma_{12}) \right)}{\langle \nabla p_{m} \ast \nabla p_{m} \rangle (1 - \gamma_{12})} \tag{18}
$$
and

\[
H_p(z, \omega) = \frac{\langle \tau_0^* A_{G0} \rangle \left( 1 - \frac{\langle \tau_0^* \nabla p_0^* \rangle \langle \nabla p_0^* A_{G0} \rangle}{\langle \nabla p_0^* \nabla p_0^* \rangle \langle \tau_0^* A_{G0} \rangle} \right)}{\langle \tau_0^* A_{G0} \rangle [1 - \gamma^2_{12}]} \quad (19)
\]

Here \( \gamma^2_{12} = |\langle \nabla^2 p_0^* \rangle|^2 / (\langle \nabla^2 p_0^* \rangle \langle \tau_0^* \rangle) \), and we estimate geostrophic acceleration, baroclinic pressure gradient, and wind stress as in section 3a. We assume that the flow at \( z_0 \) is entirely pressure forced, \( \nabla^2 \) is constant above 40 m, and the dynamics are sufficiently linear for (16) to hold. For all moorings, \( z_0 \) is in the seasonal thermocline, far beneath the maximum mixed layer depth and isolated from surface-forced turbulent momentum flux. Assuming constant density gradients above 40 m is equivalent to an assumption of constant geostrophic shear. An advantage of this formulation is that we calculate transfer functions between geostrophic acceleration and the parts of the two forcing fields not coherent with each other. Thus, \( H_p(z, \omega) \) (unitless) is the acceleration response to baroclinic pressure gradient forcing not coherent with the wind, and \( H_r(z, \omega) (m^{-1}) \) is the response to wind forcing not coherent with the pressure gradients. This acts to filter the coupled response. Complex transfer functions \( H_p(z, \omega) \) and \( H_r(z, \omega) \) describe the response to forcing by pressure gradients and by wind-driven stress divergence.

We calculate the transfer functions at moorings F3, F4, F5, F6, F7, and F8 and form averages at each depth to estimate the response (Fig. 10). Here the leftmost column is \( H_p(z, \omega) \) and the center column is \( H_r(z, \omega) \), with the three rows corresponding to depth groupings. For the pressure-driven response, the solid-tipped scale arrow indicates 1.0 (unitless) in the direction of the real axis. Coincidence of \( H_p(z, \omega) \) and the scale arrow in-
icates perfect geostrophic balance. Interpretation of the wind-driven response is similar to that described in section 3c, with slight changes in scale.

As expected from section 3b, we find good agreement with geostrophy, although transfer function magnitudes are slightly less than unity. In sharp contrast with the earlier model forced solely by wind stress, the low-frequency wind-driven response exhibits surface intensification, a distinctive rightward-turning spiral, and decay with depth. Vertically integrating $H_r(z, \omega)$ from 180 m to the surface demonstrates that the response is in good agreement with Ekman transport (rightmost column, Fig. 10). Including pressure gradient forcing in (17) allows us to explicitly separate the flow driven by stress convergence from the much larger response forced by Ekman pumping. These results are not sensitive to the choice of $z_0$, provided it is far below the influence of direct wind forcing.

Ekman pumping drives a longwave response to vertical velocities imposed by curl in the surface wind field. The observed response may be the result of forcing by local winds or a signal propagating from a remote forcing site (Brink 1989). Atmospheric fronts had wavelengths large compared to the footprint of the moored array, making direct tests of dynamical balances using local wind stress curl problematic. Instead, we identify acceleration coherent with both local winds and pressure gradients as driven by Ekman pumping, under the assumption that the wind field and its curl will be well correlated. We may express this response as

$$H_{EP}(z, \omega) = H_E(z, \omega) - H_r(z, \omega),$$

(20)

where we calculate $H_E(z, \omega)$ from (19). The pressure-coupled response is much larger than that due to stress divergence and displays surface intensification and uniform phase with respect to the wind over the upper 200 m (Fig. 11). It is this large, wind-coherent response that has complicated past attempts to isolate directly wind-driven, upper-ocean currents. Attempts to reconcile the phase of $H_{EP}(z, \omega)$ with that expected from forced Rossby wave dynamics have been unsuccessful. Further progress would require using gridded wind products to examine the phase relationship between local wind stress and local and remote curl.

f. Nonlinear acceleration

Peaks in horizontal momentum advection corresponding to frontal passages (Fig. 2d) suggest that these terms may play a significant role in the upper-ocean momentum balance. We include horizontal advection in our model by reformulating the acceleration term on the rhs of (17) as

$$A_{NL0}(z, \omega) = A_{G0}(z, \omega) + [\mathbf{u} \cdot \nabla \mu + i(\mathbf{u} \cdot \nabla \nu)]_{\omega,0}$$

$$- [\mathbf{u} \cdot \nabla \mu + i(\mathbf{u} \cdot \nabla \nu)]_{\omega,0}. \quad (21)$$

Here $\mathbf{u}(z, \omega) = (u(z, \omega), v(z, \omega))$, $\nabla = (\partial/\partial x, \partial/\partial y)$ and we calculate horizontal momentum advection relative to the deepest observations as described in section 3a. Substituting $A_{NL0}(z, \omega)$ for $A_{G0}(z, \omega)$ in (17), we estimate the transfer functions as before (Fig. 12).

The pressure-driven response remains virtually unchanged, indicating that accelerations due to horizontal momentum advection are small relative to geostrophic accelerations. Likewise, the directly wind-driven response at any individual depth is not altered by a margin larger than the error. However, adding horizontal momentum advection brings the depth integrated transport into better agreement with theory and the transport is significantly different from that calculated without advection. Rudnick and Weller (1993b) obtain a similar result at super- and near-inertial frequencies. Both the results presented here and those of Rudnick and Weller (1993b) demonstrate that horizontal gradients in the background flow affect the wind-driven response. The effects of horizontal momentum advection are small but significant in the long timescale, average sense of this analysis. During frontal passages, ageostrophic crossfront currents acting on large horizontal gradients will enhance the role of the advective terms.
4. Discussion and conclusions

Geostrophy holds for timescales longer than approximately 100 hours (Fig. 5), with the highest skill levels residing in the seasonal pycnocline. Near the surface, wind forcing contributes significant ageostrophic shear, resulting in the reduced skill levels observed. A simple scaling argument explains the decrease in skill at higher frequencies. Advectional forcing implies $\omega = c_p \cdot K$, where $c_p$ is phase speed and $K$ is the horizontal wavevector. A phase speed of 0.2 m s$^{-1}$ (Eriksen et al. 1991) and a horizontal resolution of 40 km yields a timescale of 56 hours. This places a lower bound on the timescales for which we can resolve horizontal density gradients.

Fronts passing through the FASINEX moored array not only display a preferred alignment (Eriksen et al. 1991), but alongfront currents exhibit geostrophic dynamics. The low-frequency geostrophic balance exhibits anisotropy, with currents aligned in the NE–SW alongfront direction in thermal wind balance while those perpendicular to them are not. Alongfront alignment as indicated by the thermal wind balance is consistent with both the principal axes of density gradient fluctuations as well as the typical frontal orientation previously reported by Eriksen et al. (1991).

At low frequencies, identifying the geostrophic flow is critical to our ability to distinguish between currents driven directly by the wind through convergence of turbulent momentum flux and those forced indirectly by Ekman pumping. Comparing $H_{sp}(z, \omega)$ (Fig. 11) to $H_{s}(z, \omega)$ (Fig. 12) demonstrates that the response driven indirectly by the wind through pressure gradient forcing is much larger than that driven directly by stress convergence. The difficulty lies in separating the two wind-coherent responses. Currents driven by Ekman pumping are in geostrophic balance. Previous investigators (Davis et al. 1981b; Price et al. 1987; Weller et al. 1991) exploit this by referencing currents to the flow just below the mean mixed layer base, assuming that the pressure-driven response is barotropic and that wind-driven turbulent momentum flux penetrates no deeper than the base of the average mixed layer. In contrast, we use estimates of baroclinic pressure gradients to identify the geostrophic flow and filter the
Ekman pumping response. We assume that the effects of wind-driven turbulent momentum flux vanish at a deep reference level within the seasonal pycnocline, well below the mixed layer base, and that geostrophic shear is constant above 40 m, our shallowest density measurements. Our analysis is insensitive to the choice of reference level, provided it is far below the mixed layer. This insensitivity and satisfaction of the Ekman transport relation (Figs. 10 and 12) support the validity of our assumptions. By employing a model that includes pressure gradient forcing, we relax assumptions about the penetration depth of wind-driven mixing and demonstrate that the anomalous low-frequency response reported in previous studies is indeed forced by pressure gradients.

Linear wave theory predicts the influence of Ekman pumping to shoal as $\omega \rightarrow f$ (Philander 1978), and the pressure-coupled response may play a much smaller role at short timescales. In this case, a one-dimensional model (14) should suffice to isolate the directly wind-driven response and pressure gradient forcing may be safely neglected. Indeed, repeating the analysis presented in section 3e for the clockwise and anticlockwise weather bands produced no change in the directly wind-driven response from that found by Weller et al. (1991). This is not surprising, as weather-band transfer functions from the one-dimensional model (Fig. 9) required little correction to satisfy the Ekman transport relation.

The observed response to direct wind-forcing extends through the mixed layer and into the stratified fluid below. Variations in stratification modulate penetration depth, and the average penetration depth described by the transfer function may not accurately reflect the instantaneous extent of the wind-driven shear stress. However, even in the average sense of the transfer function, direct wind-forcing penetrates to depths below the well-mixed region and often within the seasonal pycnocline. After normalizing by mixed layer depth, Chereskin and Roemmich (1991) and Wijffels et al. (1994) find Ekman transport divided into a slab mixed layer overlaying a sheared transition layer of similar vertical extent. One-third of the transport occurs in the stratified region below the mixed layer. Clearly, the specifics of these results depend on how the mixed layer is defined (Wijffels et al. 1994), but the implication is that the effects of direct wind-driving penetrate into the stratified fluid. This sub-mixed-layer response accounts for observed transports falling short of theoretical values in studies that assume no turbulent flux through the mixed layer base.

In contrast, during the Long Term Upper Ocean Study (LOTUS), Price et al. (1987) find 95% of the Ekman transport trapped within the 25-m deep summertime mixed layer. LOTUS was marked by fair summer weather, a clear diurnal cycle of stratification and the passage of energetic eddies, while FASINEX took place from winter to late spring in a highly advective regime dominated by upper-ocean fronts (Rudnick and Weller 1993a). The diurnal cycle of solar heating controls stratification during LOTUS, but advective effects and seasonal differences in atmospheric forcing make it less dominant during FASINEX. This may explain the different degrees of surface trapping observed.

The low-frequency wind-driven response spirals to the right and decays with depth. The presence of a spiral in our results does not imply that an eddy diffusivity is the proper way to parameterize mixing, as other mechanisms can produce sheared current profiles in the mean. Price et al. (1987) attribute a similar spiral observed during LOTUS to the diurnal cycle of stratification and wind stress (Price et al. 1986). Using the model described by (9)–(13), a nonlinear least squares fit of a constant eddy viscosity to the observed response (Fig. 10) yields $K_e = 3.6 \times 10^{-2} (\pm 1.3 \times 10^{-2})$ m$^2$ s$^{-1}$. The implied $e$-folding scale of 33 m reflects the average mixed layer depth and is similar to eddy diffusivities estimated for LOTUS (Price et al. 1987).

We attribute the large, deep-penetrating response coherent with both wind stress and pressure gradient to Ekman pumping associated with wind stress curl. Brink (1989) reports predominantly nonlocal coherence between deep oceanic response during FASINEX and basinwide curl using NOAA ATOLL, Bakun, and FNOC gridded wind products. At the timescales considered (shorter than 160 days), only barotropic Rossby waves are free to propagate and generate nonlocal response, while baroclinic modes are evanescent and contribute only to local coherence. Our results clearly show a surface-intensified response, which maintains a constant phase over the depth range of the observations (Fig. 11). This may be the signature of a locally forced response, unable to propagate and decaying away from the surface. Brink (1989) also finds surface intensification but notes that, over sloping topography, the equivalent of the barotropic mode can become surface intensified. Thus, given possible coherence between local and remote winds, remote forcing offers an alternative explanation for our observations.

Horizontal advection of a geostrophic frontal jet by ageostrophic, directly wind-driven currents may also produce a pressure-driven response that is coherent with local wind. To examine the size of this effect, we consider an Ekman current of 0.01 m s$^{-1}$ directed along the frontal density gradient, acting over the typical 4-day timescale of passing weather systems. Assuming the Ekman flow extends no deeper than 100 m, an upper bound on the pressure gradient acceleration due to advection of the frontal jet by directly wind-driven currents is $O(1 \times 10^{-6}$ m s$^{-2})$. In contrast, typical values for the transfer function of the pressure-coupled response, $H_{pp}$ (Fig. 11) and the surface stress (Fig. 2) yield accelerations of $O(1 \times 10^{-5}$ m s$^{-2})$. Thus, this mechanism should not play a significant role in the observed pressure-coupled response pictured in Fig. 11.
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