Open Ocean Momentum Flux Measurements in Moderate to Strong Winds

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ABSTRACT

Measurements of the momentum flux were made by the Reynolds flux and dissipation methods on a deep water stable tower operated by the Bedford Institute of Oceanography. A modified Gill propeller-vane anemometer was used to measure the velocity. Drag coefficients from 196 Reynolds flux measurements agree well with those reported in Smith (1980) based on independent observations at the same site. Based on 192 runs, a comparison of the dissipation and Reynolds flux results shows excellent agreement on average, for wind speeds from 4 to 20 m s^{-1}. The much more extensive dissipation data set (1086 h from the tower and 505 h from the weathership PAPA, CCGS *Quadra*) was used to investigate the dependence of the drag coefficient on wind speed, fetch and stability. The drag coefficient reduced to 10 m height and neutral conditions (CDN), is independent of stability and fetch (for fetch/height $\geq$ 800) but increases with wind speed above 10 m s^{-1}. Some time series of the momentum flux and drag coefficient are shown to demonstrate additional sources of variation in the drag coefficient. CDN is observed to be smaller, on average, during rising winds than during falling winds or after a change in wind direction. Based on our results and many deep water results of others, we obtain

$$10^3\,\mathrm{CDN} = \begin{cases} 1.2, & 4 \leq U10 < 11\ \mathrm{m\,s^{-1}} \\ 0.49 + 0.065U10, & 11 \leq U10 \leq 25\ \mathrm{m\,s^{-1}}, \end{cases}$$

where $U10$ is the wind speed at a height of 10 m. A method for calculating the stress from this CDN and observations of wind speed and surface minus air temperature at heights other than 10 m is also given.

1. Introduction

The interaction of the atmosphere and ocean is important to a number of processes, including the large scale circulations of the atmosphere and ocean and, at smaller scales, mixed layer development and wave generation. Modeling and predicting large scale features require the surface fluxes, which are too difficult and costly to measure directly on this scale, to be estimated from more easily measured quantities through parameterizations. For this purpose, extension of drag coefficient measurements to 20 m s^{-1} ought to suffice, since higher winds contribute little to fluxes averaged over a month or more (Fissel *et al.*, 1977). A further extension to 25 m s^{-1} should clearly reveal any wind speed dependency. The capability of obtaining flux measurements in winds up to 40 m s^{-1} would allow the entire time histories of large storms to be followed, which is desirable for the study of synoptic and mesoscale processes. In the case of hurricanes, a specialized investigation is probably required, because of their very high winds and unique pattern of occurrence. A great deal of scatter is typical of turbulence measurements, suggesting that a large amount of data is required before a representative picture of the open sea can be obtained.

There are several obstacles that make direct measurements in high winds difficult and that have restricted the majority to low and moderate winds and to near or onshore platforms. The most common methods, the Reynolds flux (or eddy correlation) and profile work best on stable platforms with minimal flow distortions, but these conditions are not easily satisfied during storms at sea. For open sea work a ship is a convenient platform and the dissipation method of measuring fluxes should be possible in spite of the motion and flow distortion in high winds. An experimental program was designed, therefore, first to establish the validity of the dissipation method by comparison with simultaneous Reynolds flux measurements from a stable deep water tower, then to obtain a large quantity of high wind speed, shipboard, dissipation measurements.

This paper is based on the momentum flux portion of a Ph.D. thesis (Large, 1979) which also presents concurrent sensible heat flux measurements and describes the instruments, experiments and data analysis in greater detail. These sensible heat flux

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data and measurements taken during the JASIN, 1978 experiment, will be the subject of a future publication.

2. Measurement techniques

Exchanges between the atmosphere and ocean are most easily measured in the atmospheric surface layer where the fluctuating vertical velocity bodily transports fluid properties up and down. The resulting Reynolds fluxes are defined by

$$\left.\begin{array}{lr} \text{Momentum flux} & \\ \quad\text{(surface stress)} & \tau = -\rho\langle uw\rangle \\ \text{Sensible heat flux} & H_s = \rho C_p\langle wt\rangle \\ \text{Latent heat flux} & H_L = L_E\langle wq\rangle \end{array}\right\}, \quad (1)$$

where ρ is the mean air density, C_p the specific heat at constant pressure and L_E the latent heat of evaporation. Following Reynolds' convention, the downstream, cross-stream and vertical velocities, the air temperature and the absolute humidity are partitioned into a mean, $\langle\ \rangle$, (time average) and fluctuation and are given by $\langle U\rangle + u$, $\langle V\rangle + v$, $\langle W\rangle + w$, $\langle T\rangle + t$ and $\langle Q\rangle + q$, respectively. The behavior of this turbulent layer is well described by Monin-Obukhov similarity theory. This theory introduces the velocity scale.

$$u^* = (\tau/\rho)^{1/2} = |\langle uw\rangle|^{1/2}, \quad (2)$$

known as the friction velocity. Another important scale is the Monin-Obukhov length L whose magnitude gives the height at which the magnitude of the buoyant turbulent kinetic energy B produced in a non-neutral air column, is equal to the purely mechanical production in the equivalent neutral case, P_0. It can be shown (Lumley and Panofsky, 1964) that

$$B = g\langle wT_v\rangle/T_0 \quad \text{and} \quad P_0 = u^{*3}/\kappa Z, \quad (3)$$

where Z is the height above the sea, κ von Kármán's constant, g the acceleration due to gravity and T_v is the virtual temperature (K) with T_0 its local average. An appropriate stability parameter, Z/L, is

$$\frac{Z}{L} = \frac{-B}{P_0} \quad \text{with} \quad L = \frac{-u^{*3}T_0}{\kappa g\langle wT_v\rangle}, \quad (4)$$

where the minus sign makes $Z/L > 0$ in stable stratification. An important consequence of similarity theory is the profile of the mean wind described by

$$\frac{\kappa Z}{u^*}\frac{\partial\langle U\rangle}{\partial z} = \phi_m(Z/L), \quad (5)$$

where $\phi_m(0) = 1$, making the neutral profile logarithmic. Measured values of κ range from 0.35 to 0.42 (Busch, 1977), so a 5% uncertainty is present in our assumed value of 0.40. Dyer (1974) suggests

$$\phi_m(Z/L) = \begin{cases} 1 + 5Z/L & 0 < Z/L < 0.2 \\ (1 - 16Z/L)^{-1/4} & -1.0 < Z/L < 0. \end{cases} \quad (6)$$

At a height Z, the mean wind, UZ, is found by integrating (5) (Paulson, 1970):

$$UZ = [u^*/\kappa][\ln(Z/Z_0) - \psi_m(Z/L)],$$

where

$$\psi_m(Z/L) = \begin{cases} -5Z/L, \quad \text{stable } (Z/L > 0) \\ 2\ln[(1 + X)/2] + \ln[(1 + X^2)/2] \\ -2\tan^{-1}X + \pi/2, \quad \text{unstable } (Z/L < 0) \end{cases}$$

with

$$X = (1 - 16Z/L)^{1/4}. \quad (7)$$

The roughness length Z_0 describes the surface as "seen" by the turbulence and it may be a complicated function of sea surface parameters (Burling and Stewart, 1967).

a. *The Reynolds flux method*

The most direct flux measurement is the Reynolds flux or eddy correlation method. Integration of the u, w cospectrum, $\Phi uw(f)$, over all contributing frequencies f gives the covariance and hence momentum flux. Unfortunately, measurements cannot be extended to more than about one hour before stationarity begins to be lost, hence the low-frequency flux contributions cannot be well established. Other disadvantages of this method are the large amount of data required and the sensitivity of the method to instrument orientation, which is a great handicap on ships and buoys. Although the Reynolds flux method is not easily applicable to remote open sea operation, it has become the standard to which the other methods are compared.

A Reynolds flux system was designed to record automatically the necessary data for Reynolds flux estimates from 4 (and sometimes 3) sequential 13.5 min groups. Each group consisted of 256 values of the three velocity components sampled simultaneously at the slow rate of ⅓ Hz (12.8 min) and 128 values sampled at the faster rate of 3 Hz (0.7 min). Before being sampled at the slow rate the velocity signals were low-pass filtered. These filters were 3dB down at about the Nyquist frequency (⅙ Hz), so that the filter loss and aliasing would nearly cancel one another in the low frequency cospectral values (0.00065–0.167 Hz) calculated from the slow samples. Before being sampled at the faster rate the velocity signals were prewhitened. These circuits behaved as time differentiators at low frequencies in order to increase signal levels and eliminate spectral distortion from nonsampled frequencies. At higher frequencies they rolled off as R-C lowpass filters. By design their response was maximum and nearly flat at the Nyquist frequency (1.5 Hz), so that aliasing would occur without much loss of contributions below a few Hz. The loss of a small amount of covariance was expected as a consequence of the aliased frequencies being undercorrected for sensor response. The poorly determined lower frequency cospectral values from the fast samples were not required, since the covariance of the lower 6 (and part of the seventh) frequency bands (0.023–0.167 Hz) was available from the slow samples. Although the u, w cospectra, formed by averaging $\Phi uw(f)$ from each group, were distorted at frequencies above about 1 Hz by aliasing, they should have contained almost all the covariance from 0.00065 Hz to about $fn \simeq 4$ Hz.

In an attempt to minimize the variability caused by the uncertain low frequency contributions to the momentum flux, the integration of the u, w cospectra began at different frequencies, f_1, which always corresponded to the same natural frequency, $n_1 = f_1 Z/\langle U \rangle = 0.004$. This integral was then multiplied by a factor E to compensate for the excluded low frequencies. In order to determine E, the quantity

$$\frac{f\Phi uw(n)}{\int_{f_1}^{f_n} \Phi uw(f)df},$$

where the denominator is a first approximation to $\langle uw \rangle$ and the multiplication by f preserves the covariance of a plot against $\log n$, was first calculated at all the discrete natural frequencies from all the available runs, then averaged over bands of constant $\log n$. These band averages were taken to be normalized cospectral values, $N\Phi uw(n)$, with n the midpoint of the logarithmic band, as plotted in Fig. 2. This Reynolds flux method is expressed by

$$\langle uw \rangle \text{ FLUX} = E \int_{f_1}^{f_n} \Phi uw(f)df,$$

where

$$E = \frac{\int_{n=0}^{n2} N\Phi uw(n)d(\log n)}{\int_{n=0.004}^{n2} N\Phi uw(n)d(\log n)}. \quad (8)$$

With aliasing, the upper limit of integration, $n2$, was effectively about 10.

b. The dissipation method

The dissipation method is very attractive because it does not involve an explicit measurement of the vertical velocity, allowing moving platforms to be used and reducing measurement errors. Instead, the major sources of error arise from the uncertainty of various constants and from the necessary assumptions. The method is based on the consideration of the balance of turbulent kinetic energy per unit mass, $e = (u^2 + v^2 + w^2)/2$, which in steady, horizontally homogenous flow (Busch, 1977) becomes

$$\begin{aligned} 0 &= u_*^2 \frac{\partial\langle U \rangle}{\partial z} + g\frac{\langle wT_v \rangle}{T_0} - \epsilon - \frac{\partial}{\partial z}[\langle we \rangle + \langle wp \rangle/\rho] \\ &= P + B - \epsilon - D, \end{aligned} \quad (9)$$

where p is the fluctuating pressure. Turbulent fluctuations are chiefly produced by mechanical interactions P and lost to molecular dissipation ϵ. The complete divergence term D has been investigated by McBean and Elliott (1975), whose measurements show that $[\langle we \rangle + \langle wp \rangle/\rho]$ is nearly independent of Z/L, making $D \ll P$ a reasonable approximation on average. Substituting (3), (4) and (5) into the remaining terms of (9) $(P + B = \epsilon)$, allows u^* to be expressed as a function of ϵ and Z/L:

$$u^{*3} = \kappa Z\epsilon[\phi_m(Z/L) - Z/L]^{-1} \quad (10)$$

Three other methods have been tried: Denman and Miyake (1973) assumed the neutral case $P_0 = \epsilon$, Pond *et al.* (1971) used $P_0 + B = \epsilon$, while Smith and Banke (1975) employed $P = \epsilon$.

A direct measurement of ϵ is difficult, but it can be inferred from the spectrum of the downstream velocity, $\Phi u(f)$, at frequencies f in the $-5/3$ region where the Kolmogoroff hypothesis predicts

$$\Phi u(f) = K'\epsilon^{2/3}(2\pi f/\langle U \rangle)^{-5/3}, \quad (11)$$

where Taylor's hypothesis has replaced the downstream radian wavenumber with $(2\pi f/\langle U \rangle)$. A reasonable value of the one-dimensional Kolmogoroff constant K' is 0.55 ± 10% (Paquin and Pond, 1971 and Busch, 1977). In our dissipation system the power averaged over 4 to 5 min from three band-pass filters, centered nominally at $fc = 0.4$, 0.8 and 1.6 Hz, was used to find the discrete spectral values, $\Phi u(fc)$, at each fc. Dissipation estimates of the momentum flux were then calculated from (10), as

$$\langle uw \rangle \text{ DISS} = (\kappa Z\epsilon)^{2/3}[\phi_m(Z/L) - Z/L]^{-2/3}, \quad (12)$$

with ϵ the average of the three estimates found from the $\Phi u(fc)$'s.

In all four methods $\langle uw \rangle$ DISS is proportional to

$$[\kappa Z]^{2/3}(K')^{-1}\Phi u(fc),$$

hence the uncertainties in κ, in the measurement height (perhaps 0.5 m in 10 m) and in K' could combine to produce a 15% error in $\langle uw \rangle$ DISS. These errors and the assumption errors were likely somewhat systematic, but fortunately they were not the same in the Reynolds flux method. Intercomparisons were essential in order to establish (10) as the best dissipation method and to ensure that there were no major systematic errors (Large, 1979).

c. *The bulk aerodynamic method*

The bulk aerodynamic formulas parameterize the air-sea fluxes in terms of the mean wind speed UZ, potential temperature θZ, and absolute humidity QZ, at the height Z, the surface humidity Q_s, and sea temperature T_s:

$$\left.\begin{aligned} -\langle uw \rangle \text{ BULK} &= \text{CD}UZ^2 \\ \langle wt \rangle \text{ BULK} &= \text{CT}UZ\Delta\theta \\ \langle wq \rangle \text{ BULK} &= \text{CE}UZ\Delta Q \end{aligned}\right\} , \quad (13)$$

where $\Delta\theta = T_s - \theta Z$ and $\Delta Q = Q_s - QZ$. The air temperature TZ and sea minus air temperature difference ΔT are given by

$$TZ = \theta Z - \gamma Z$$

$$\Delta T = \Delta\theta + \gamma Z,$$

where γ is the adiabatic lapse rate, 0.01 K m^{-1}. The nondimensional transfer coefficients CD, CT and CE are the drag coefficient, Stanton number and Dalton number, respectively. In practice they must be determined from direct flux measurements and the mean or bulk quantities. To eliminate its variation with height, the drag coefficient is commonly evaluated at 10 m as $C10$, using the wind speed at 10 m, $U10$, obtained from (7). The stability dependence is also removed by calculating the coefficient in the equivalent neutral case at 10 m (CDN) using Z_0 from (7):

$$\left.\begin{aligned} U10 &= UZ - [u^*/\kappa][\ln(Z/10) - \psi_m(Z/L) \\ &\qquad + \psi_m(10/L)] \\ C10 &= -\langle uw \rangle / U10^2 \\ \text{CDN} &= \kappa^2[\ln(10/Z_0)]^{-2} \\ &= C10[1 + (C10)^{1/2}\kappa^{-1}\psi_m(10/L)]^{-2} \end{aligned}\right\} , \quad (14)$$

with Z, Z_0 and L in meters.

An experimental formulation of CDN allows the momentum flux to be estimated from UZ, through (13), with Z/L, if known, providing a stability correction. The drag coefficient CD at the measurement height Z (m) is given by

$$\text{CD} = \text{CDN}\{1 + (\text{CDN})^{1/2}\kappa^{-1} \times [\ln(Z/10) - \psi_m(Z/L)]\}^{-2}. \quad (15)$$

If CDN is given as a function of $U10$, UZ can be shifted to 10 m using

$$UZ/U10 = 1 + (\text{CDN})^{1/2}\kappa^{-1} \times [ln(Z/10) - \psi_m(Z/L) + \psi_m(10/L)] \quad (16)$$

(assuming $C10 \approx$ CDN, which introduces an error of only 1% into $U10$ for $-1 < 10/L < 0.2$) and (16) can be quickly solved iteratively. If the stability is in the range $-1.0 < Z/L < 0.2$, but is unknown and assumed to be neutral, errors arise from an inaccurate $U10$ used to find CDN and because CD is not shifted to the proper stability. For $Z > 5$ m the total error in $U10$'s found from (16) will be less than 10%, which at 20 m s^{-1} reduces to 7% in CDN and the momentum flux. With $\psi_m(Z/L)$ ranging from 1.0 to -1.0, taking it to be zero could result in an additional 20% error in CD and in $\langle uw \rangle$ BULK. Only in the range $-0.25 < Z/L < 0.1$ does the total error in assuming neutral stability remain less than 10%. If, however, ΔT is also available, Z/L can be estimated from UZ and ΔT (Section 3c), reducing the associated error in CD to ~5%, throughout $-1.0 < Z/L < 0.2$.

3. The experiments

a. *Instrumentation*

A complete description of both the Reynolds flux and dissipation systems, including error analysis, design criteria and sensor requirements can be found in Pond and Large (1978) and Large (1979). The essential considerations for their remote operation were low power consumption, a large recording capability and robust sensors. The detailed analysis of the velocity measurements sytem (based on a Gill propeller-vane anemometer) is also in Pond *et al.* (1979). A horizontal propeller, Gill-u, sensed the downstream velocity component, while the vertical velocity was derived from a propeller, Gill-w, whose axis was tilted down at an angle $\alpha \approx 60°$ to the axis of Gill-u. In winds over 4 m s^{-1}, the axial component is sufficient to keep Gill-w in its linear operating regime, but corrections for its non-cosine behavior are necessary.

For Reynolds flux calculations the average vertical velocity was made exactly zero by assuming that Gill-u was tilted from the horizontal by an appropriate small angle δ. Removal of a portion of Gill-u from Gill-w left the instantaneous vertical velocity W, which was then removed from the Gill-u signal to give the horizontal velocity. The vane's signal was used to resolve the cross-stream V and downstream U components, such that $\langle V \rangle = 0$. The important errors associated with this procedure come from determining the offset and correcting for the non-cosine behavior (which both produce apparent tilts), calibrating the propellers and measuring α. With some errors tending to cancel, $\sim \pm 10\%$ instrument error could be left in the *average* drag coefficient.

Only the Gill-u signal was used for dissipation calculations, where the above errors are not important. More serious errors arise from the assumptions and constants used, from the estimate of Z/L and in extracting a ship's velocity. Again the error in the average CD is expected to be $\sim \pm 10\%$.

b. The experimental sites

The Reynolds flux and dissipation methods were compared from measurements on the Bedford tower (44°29′44″N, 63°23′30″W) off the harbor of Halifax, Nova Scotia, between September 1976 and April 1977. A description of the tower and the results from a Thrust anemometer system operated by the Bedford Institute of Oceanography (BIO) can be found in Smith (1980). The tower was a floating spar buoy moored in 59 m of water making the site essentially a deep water wave regime (Smith, 1980). The shortest fetch was about 10 km, while open fetch conditions extended over a 170° range. The anemometer was located ~12.5 m above mean sea level and the tidal range was rarely >2 m. An air temperature was taken near the Gill and a sea temperature sensor was tied to the tower ~ 10 m below mean sea level. Routine observations, including the atmospheric pressure PA in kPa, for density calculations ($\rho = 1.29[273/\langle T\rangle][\mathrm{PA}/101]$ in kg m^{-3}) were available from the Shearwater "A" meteorological land station, about 15 km north of the tower.

The dissipation measurements were extended to higher wind speeds and more open sea conditions in a second experiment conducted from the CCGS *Quadra* during its four patrols at ocean weather station PAPA (50°N, 145°W) from July 1977 to April 1978. The anemometer and air temperature sensor were installed on the ship's foremast about 22 m above its waterline. A sea surface "bucket" temperature and the air pressure were recorded as part of the ship's three-hourly meteorological observations. Some useful data were obtained as the ship steamed at 7–12 kt while en route to PAPA, however, the majority were collected on station as the ship steamed into the wind at <4 kt. With winds over the bow, measured tilt angles δ were typically only 7°, indicating that the ship did not seriously distort the mean flow.

c. The bulk stability parameter, Z/L

Since a measure of the virtual temperature was never available, a bulk estimate of the stability parameter was found from the mean wind speed and air temperature at the measurement height, UZ and TZ, and the sea surface temperature T_s. The surface humidity Q_s, was found from T_s (assuming 98% saturation over salt water) and a relative humidity of 75% (the expected median value) was assumed at Z in order to obtain an absolute humidity, QZ ($\Delta Q = Q_s - QZ$). The saturation humidity over pure water as a function of temperature (K) was approximated by

$$QSAT(T) = 6.4038 \times 10^8 \exp(-5107.4/T),$$

so $Q_s = 0.98QSAT(T_s)$ and $QZ = 0.75QSAT(TZ)$ were calculated. Over temperate seas a good measure of T_0 is

$$T_0 \approx TZ + TZ^2 QZ(1.72 \times 10^{-6})$$

and expanding $\langle wT_v\rangle$ (Large, 1979) in (4) gives

$$Z/L \approx \frac{-\kappa Z g\langle wt\rangle}{u^{*3}T_0}\left(1 + T_0^2 \frac{\Delta Q}{\Delta\theta} 1.72 \times 10^{-6}\right), \quad (17)$$

where the bulk formulas (13), with CT ≈ CE (Large and Pond, 1980), makes $\langle wq\rangle/\langle wt\rangle \approx \Delta Q/\Delta\theta$. Following Deardorff (1968), the bulk formulas can be substituted for u^{*3} and $\langle wt\rangle$ in (17) and a portion of the Z/L expression identified as a bulk Richardson number

$$\mathrm{Ri}(\Delta\theta) = \frac{-gZ\Delta\theta}{(UZ)^2 T_0}\left(1 + T_0^2 \frac{\Delta Q}{\Delta\theta} 1.72 \times 10^{-6}\right).$$

We obtained adequate estimates of Z/L using ΔT in place of $\Delta\theta$. During the experiments $\Delta\theta$ and ΔT differed by ~0.2 K on the *Quadra* and 0.1 K at the tower. These differences are not significant at large values of $|Z/L|$ (due to large $|\Delta T|$), when the stability correction is important in finding CD, (15), for bulk estimates, (13), and in calculating $U10$ and CDN, (14). The use of ΔT in place of $\Delta\theta$ results in Z/L being reduced by about 0.02 (5 m s^{-1} winds at the tower or 8 m s^{-1} winds on *Quadra*). Close to neutral stability, where the flux estimation, (12), is most sensitive to this error in Z/L, this difference would make the momentum flux larger by ~4% if $Z/L > 0$ and 3% for $Z/L < 0$. The difference in Z/L and the errors in the flux quickly decrease as $(UZ)^{-2}$. Hence, bulk estimates of Z/L were calculated from

$$Z/L(\Delta T) = \kappa\ \mathrm{CT}\ \mathrm{CD}^{-3/2}\ \mathrm{Ri}(\Delta T). \quad (18)$$

At the lower wind speeds where stability has its largest range, a reasonable average CD is 1.25×10^{-3} (Garratt, 1977), while Friehe and Schmitt (1976) show CT to be $\sim 1.0 \times 10^{-3}$ in unstable stratification and $\sim 0.86 \times 10^{-3}$ in stable. The bulk estimate (18) is very practical and it is plotted against the more exact expression (from 17)

$$Z/L(u^*, \langle wt\rangle) = \frac{-\kappa Z g\langle wt\rangle}{u^{*3}T_0}\left(1 + T_0^2 \frac{\Delta Q}{\Delta T} 1.72 \times 10^{-6}\right)$$

in Fig. 1. These are the 27 unstable and 33 stable Bedford tower runs for which $\langle wt \rangle$ FLUX is available from calculations analogous to $\langle uw \rangle$ FLUX (8). There is overall good agreement, with a typical deviation of only ± 0.05. However, the two calculations occasionally differ by about ± 0.2, implying that a bulk estimate of Z/L [from (18)], from an individual run, may be considerably in error.

The results of this paper and of our sensible heat flux measurements will allow the values of CD and CT to be updated for future bulk Z/L estimates. Our latent heat flux measurements from JASIN will allow these estimates to be compared to calculations of Z/L using (4). Better agreement may be found by using $\mathrm{Ri}(\Delta\theta)$ in (18).

4. The Reynolds flux results and dissipation intercomparison

We have analyzed 196 Reynolds flux runs from the Bedford tower. An investigation of the velocity spectra and turbulence statistics has shown that the Gill anemometer performed as expected (Large, 1979). Turbulence levels of the horizontal components are in good agreement with Smith and Banke (1975) and Smith (1980). The overall averages (± 1 standard deviation) of $\sigma u/\langle U \rangle$ and $\sigma v/\langle U \rangle$ are 0.092 ± 0.018 and 0.080 ± 0.031, respectively (where the σ's are the standard deviations of the velocity fluctuations). Much of the scatter in $\sigma v/\langle U \rangle$ is a consequence of its stability dependence. Both these averages ought to be increased (possibly by 10% at $\langle U \rangle = 5$ m s^{-1} and 2% at 20 m s^{-1}) in order to account for contributions to the variances from frequencies below 0.00065 Hz (the lowest included in the integration of spectra). After applying these corrections to the linear regressions, we obtain $\sigma u/\langle U \rangle = 0.70 + 0.0023\langle U \rangle$ and $\sigma v/\langle U \rangle = 0.043 + 0.0033\langle U \rangle$. For the vertical component *only*, not fully correcting for the response limits of the propellers above the fast sampling Nyquist frequency is serious. Therefore, our averages, $\sigma w/\langle U \rangle = 0.042 \pm 0.006$ and $\sigma w/u^* = 1.24 \pm 0.01$, are too small.

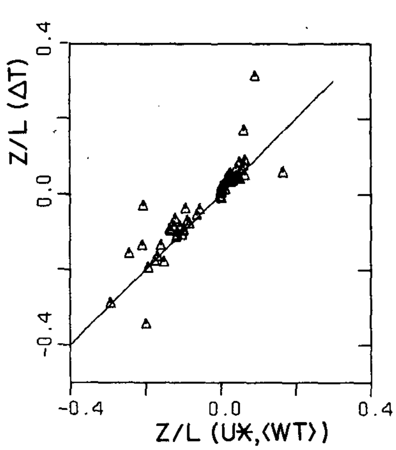

FIG. 1. Comparison of the bulk estimate of the stability parameter $Z/L(\Delta T)$ to the more exact expression $Z/L(u^*, \langle wt \rangle)$ from 33 stable and 27 unstable runs.

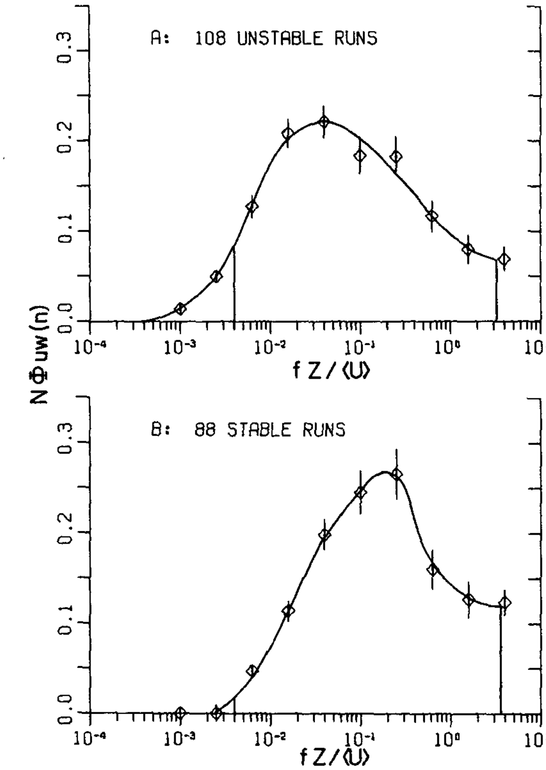

FIG. 2. Normalized unstable (A) and stable (B) u, w cospectra, $N\Phi uw(n)$, which are average of $f\Phi uw(n)/\langle uw \rangle$. Vertical bars represent an estimation of the standard deviation of the mean.

Fig. 2 shows the averaged normalized u, w cospectra; the unstable runs contain more low-frequency covariance than do the stable ones. At frequencies not distorted by aliasing, these cospectra over the sea are very similar to the over land results of McBean and Miyake (1972). The areas under the curves were found to be 1.06 and 1.005 times the area from $n = 0.004$ in the unstable and stable cases, respectively. Thus, E of (8) was made a function of stability: 1.06 for $Z/L < 0$ and 1.005 for $Z/L \geqslant 0$. The uncertainty in Z/L and its small range makes further stability adjustments to E impractical. A comparison of $\langle uw \rangle$ FLUX of (8) with integrals of $\Phi uw(f)df$ over all frequencies sampled, showed that the former included the average contributions from frequencies below $n = 0.004$, while avoiding some large random contributions from these uncertain low frequencies. Thus this method of integration should reduce the scatter in the calculated drag coefficients. In one particularly bad example at 6 m s^{-1}, the contribution to $\langle uw \rangle$ from $f = 0.00065$ Hz to

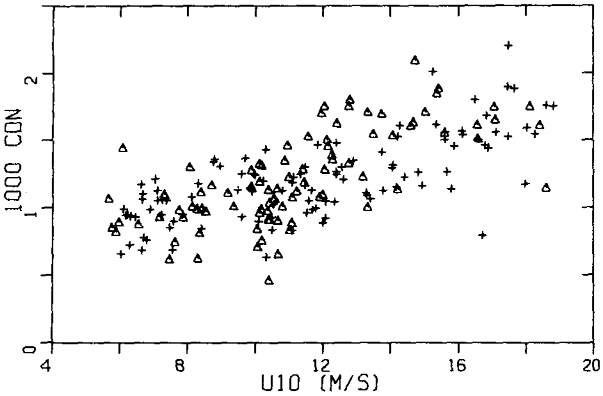

FIG. 3. The neutral 10 m drag coefficient versus wind speed from the 108 unstable (pulses) and the 88 stable (triangles) Reynolds flux runs.

$f \approx 0.002$ Hz ($n = 0.004$) was positive and of sufficient magnitude to make the integral from $f = 0.00065$ Hz to fn less than half the magnitude of $\langle uw \rangle$ FLUX.

The neutral drag coefficient (CDN) from all 196 Reynolds flux runs is plotted against $U10$ in Fig. 3. This plot is nearly identical to all the 1976–78 BIO tower results over a similar range of wind speeds (Smith, 1980), where a regression of 120 $C10$ values on $U10$ gives $0.44 + 0.063U10 = 10^3C10$ as compared to $0.46 + 0.069U10 = 10^3$ CDN (or $0.43 + 0.069U10 = 10^3C10$) from the data in Fig. 3. The higher coefficients at the higher speeds are commonly observed, but at all wind speeds the CDN's tend to be distinctly smaller than those at similar wind speeds compiled by Garratt (1977). In Fig. 3, the stable (triangles) and unstable (pulses) data are not distinguishable, supporting the theoretical prediction that CDN is independent of Z/L. A more complete view of the behavior of CDN is offered in Section 5, based on the more extensive dissipation data set.

Nearly simultaneous dissipation runs were recorded during 192 of the Reynolds flux runs. They began less than 4 (and usually less than 2) min before the start of the Reynolds flux runs and lasted for either 56 or 44 min in the case of 54 or 40.5 min (4 or 3 group) Reynolds flux runs, respectively. The dissipation ϵ was taken as the average of the individual values obtained from the power from three Gill-u bandpass filters, averaged over the run. The individual values of dissipation never differed from their average (ϵ) by more than 15% (in only 10 runs was this difference more than 10%). Much of the deviation was probably due to the spectrum's fluctuations about a $-5/3$ line. The u^* calculation, u^* DISS, is compared to the simultaneous u^* FLUX in Fig. 4. The excellent fit of the solid 1:1 line was achieved *without adjusting* the constants $\kappa = 0.40$ and $K' = 0.55$. In the region $0.15 < u^*$ FLUX < 0.4 m s^{-1}, u^* DISS from the six most stable runs ($Z/L > 0.1$) tends to be greater than u^* FLUX by more than 15%, on average. As a consequence, a least squares fit to Fig. 4, u^* DISS $= 0.96u^*$ FLUX $+ 0.025$ m s^{-1}, has a positive offset and the average u^* DISS:u^* FLUX ratio is 1.03 (standard deviation 0.10). The two techniques differ by at most 28%, and usually by less than 20%, in u^*. This amount of scatter is expected because the uncertainty in an individual calculation from *either* method is 20% in $\langle uw \rangle$ (10% in u^*).

The overall agreement between Fig. 3 and the BIO tower results (suggesting that u^* FLUX is not seriously in error) and the good agreement in Fig. 4 indicate that the proper propeller response was used, since this correction is important to u^* DISS only. Conversely, the non-cosine behavior of Gill-w is a potential source of substantial error in u^* FLUX only, and the good agreements imply that it, too, was treated satisfactorily. These conclusions are also supported by a favorable comparison between u^* from the 20 Thrust 6.4 runs (October–December 1976) in Smith (1980) and simultaneous values of u^* DISS (Large, 1979).

The simpler neutral dissipation calculation, u^* DISS1 $= (\kappa Z\epsilon)^{2/3}$, is often a good approximation and it has been plotted against u^* FLUX for all 192 simultaneous runs in Pond *et al.* (1979). The use of u^* DISS1 should be valid over a range of near neutral stabilities where the B and D terms of (9) and the effects of $\phi_m(Z/L)$ on the velocity profile either are small or tend to cancel. Such a regime was observed to extend throughout the range $-0.1 < Z/L < 0.05$. For $Z/L > 0.10$, every run gave u^* DISS1 greater, often by more than 20%, than u^* FLUX.

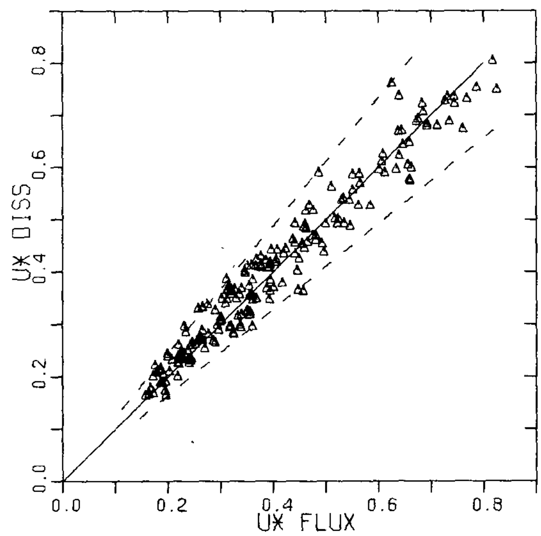

FIG. 4. Intercomparison of the dissipation and Reynolds flux methods using u^* (m s^{-1}) from all 192 simultaneous runs. Dashed lines indicate ±20% deviations from the solid 1:1 line.

Runs in the range $-0.3 < Z/L < -0.1$ showed u^* DISS1 to be smaller than u^* FLUX by an average of only 10%. In more unstable stratification the two calculations again tended to be equal, so that in the near neutral regime $-0.45 < Z/L < 0.05$, a very good estimate of the momentum flux was obtained, without an explicit knowledge of the stability, from u^* DISS1. This is a useful result because open ocean conditions are often within this range and because an accurate Z/L is difficult to obtain.

The Reynolds flux and dissipation methods usually give nearly the same momentum flux. With the possible exception of the most stable runs ($Z/L > 0.10$), the estimates of $\langle uw \rangle$ from both techniques appear to be reliable and free of major systematic errors. The ratio $\langle uw \rangle$ DISS:$\langle uw \rangle$ FLUX averages 1.05 and exhibits neither a trend with wind speed nor a stability dependency, for $Z/L < 0.05$. The use of $\phi_m(Z/L) = (1 + 7Z/L)$, for $Z/L > 0$, would make this ratio less than 1.05 for each of the six most stable runs. This is an acceptable formulation of the flux-profile relationship (Yaglom, 1977) in view of the revaluation of the 1968 Kansas atmospheric surface layer experiment (Wieringa, 1980). We have used the Dyer (1974) formulation given in (6), but the overall results are much the same, since $Z/L > 0.05$ is not common. These results establish that the dissipation method is a viable means of measuring the momentum flux, at least up to the 20 m s^{-1} winds of the intercomparison. It is now possible to have confidence in dissipation estimates from ships, where the Reynolds flux method is not practicable.

5. The drag coefficient

The dissipation system has provided a total of 1086 h of momentum flux data from the Bedford tower and 505 h from the CCGS *Quadra*, which have satisfied a variety of criteria for data reliability. The tower analysis was limited to wind speeds above 4.0 m s^{-1} because in lower winds the propeller response corrections become very large and there may be a considerable loss of flux between the surface and the measurement height. On CCGS *Quadra* the anemometer was nearly twice as high so the limit was set at 8.0 m s^{-1}. A further restriction required the stability to be in the range $-0.6 < Z/L < 0.15$, where there is experimental evidence to support the dissipation method. In processing a dissipation run, ϵ was found from 20 min averages of the $\Phi u(fc)$'s; however, if any individual bandpass value differed from the average ϵ by more than 15%, the run was rejected. This criterion ensured that all spectral values fell within 10% of the average $-5/3$ slope. An average over 20 min or less may be sufficient for the dissipation method, but this hypothesis cannot be tested be-

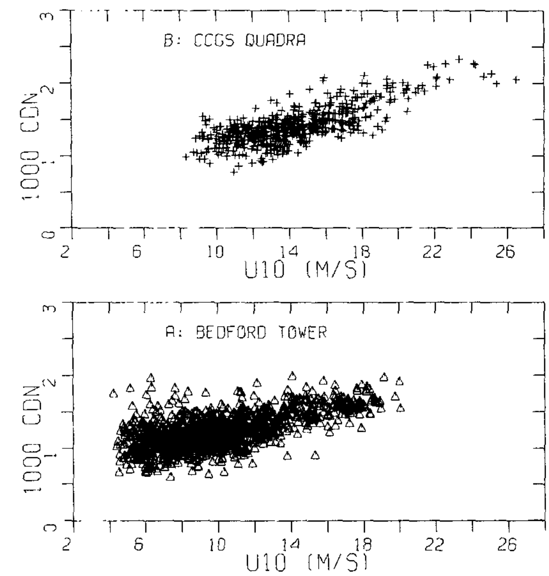

FIG. 5. The neutral 10 m drag coefficient versus wind speed, using hourly averages from dissipation measurements on (A) the Bedford tower, and (B) the CCGS *Quadra*.

cause the low-frequency contributions in the Reynolds flux method become more uncertain as the time interval decreases. Therefore, three sequential $\langle uw \rangle$ DISS estimates were averaged to obtain hourly averaged values of $\langle uw \rangle$ and hence of u^*, CD and CDN.

Hourly CDN values are plotted against $U10$ separately for the tower and ship data in Fig. 5. The observed scatter is typical of drag coefficient measurements and is uniform between $U10 = 4$ to 11 m s^{-1}. At higher speeds a tendency for higher values develops and there appears to be less scatter. Above $U10 = 20$ m s^{-1} there are few data, so average conditions may not be represented. The abundance of data at the lower winds should give meaningful averages and allow sources of the variability to be investigated. Fortunately, in the region $8 < U10 < 18$ m s^{-1}, there is a very good overlap of tower and ship results and band averaged CDN's from one platform differ by less than one standard deviation (usually by less than $\frac{1}{2}\sigma$) from the other platform's mean. This agreement establishes that the tower site was representative of open sea conditions provided that no serious error was introduced by moving to the ship. Large (1979) shows that the velocity fluctuations involved in calculating $\Phi u(f)$, and hence $\langle uw \rangle$ DISS, are at distinctly higher frequencies than those contaminated by ship motion. Agreement with the tower results has also been observed in JASIN measurements from F.S. *Meteor* (Large and Pond, 1980) at: 23 m on a mast, in 8 to 15 m s^{-1} winds and 9 m from a boom extending 10 m out from the bow, in 4 to 12 m s^{-1} winds. It seems

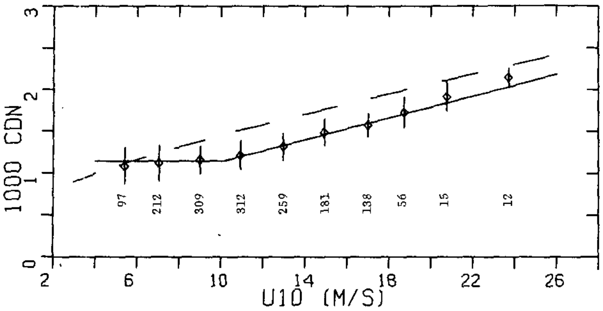

FIG. 6. CDN averaged over wind speed bands. Vertical bars extend $\pm 1\sigma$ and the number of points in a band is shown below each average. Lines show the Charnock representation of Garratt (1977) (dashed) and Eq. (19) (solid).

unlikely that this agreement would persist at these very different locations and over these large wind speed ranges, if ship influences were causing serious errors. Mast measurements were restricted to head-on winds, because from other directions superstructure interference was indicated by lack of a $-5/3$ spectrum from the three bandpass filter outputs. In contrast boom measurements were restricted to winds from either port or starboard, since the mean speeds of winds from other directions were much lower than those recorded at nearby buoys. Thus, the *Quadra* data appear to be reliable and, therefore, both data sets of Fig. 5 have been combined in the following investigations into the behavior of CDN.

Below $U10 = 10$ m s^{-1}, CDN does not vary appreciably with wind speed, so these data were used to investigate the influences of fetch and stability. They include 291 hours from unlimited fetch and 200 hours when the fetch was between 10 and 20 km and in both cases the mean 10^3 CDN is 1.14 (standard deviation 0.2). Thus, there is an indirect implication that the surface roughness Z_0 does not depend strongly on surface wave parameters that are not fully developed after 10 km during 10 m s^{-1} winds. In all subsequent analyses, data from all fetches will be considered because the better statistics resulting from the additional data should more than compensate for any unresolved fetch effects. Similarly, data from all stabilities between $Z/L = -0.60$ and 0.15 are also included, because no systematic stability dependency was observed. The average stable CDN's are larger than the unstable CDN's and they seem to increase with Z/L, but the large drag coefficients responsible for these features are associated with changes in wind direction which happen to coincide with stable stratification, as will be shown later.

The 1591 neutral coefficients from all the data are band averaged and plotted at the $U10$ mean (vertical bars show $\pm 1\sigma$) in Fig. 6. Although a smooth curve could fit all the points, the slight increase of the average CDN with wind speeds below 10 m s^{-1} should be adequately described by a constant and the more rapid rise to 26 m s^{-1} appears to be approximately linear. In order to quantify this behavior, the 618 hourly CDN values with $U10 \leq 10$ m s^{-1} were averaged and all 973 with $U10 > 10$ m s^{-1} were regressed against $U10$ (correlation coefficient 0.74) to give the solid line of Fig. 6:

$$10^3\,\mathrm{CDN} = \begin{cases} 1.14, & 4 < U10 \leq 10\ \mathrm{m\ s^{-1}} \\ 0.49 + 0.065\,U10, & 10 < U10 < 26\ \mathrm{m\ s^{-1}}. \end{cases} \tag{19}$$

The behavior of CDN described by (19) is distinctly different from the dashed curve of Fig. 6, which is the Charnock formulation suggested in Garratt's (1977) review. Eq. 19 does fit Smith's (1980) eddy correlation data from the Bedford tower site extremely well. For winds below 10 m s^{-1}, his average 10^3 CDN is 1.11 from 14 runs (1976–78) and 10^3C10 averages about 1.18 from 14 near neutral runs (1968–69). Again a regression of all 120 of his runs (1976–78) between 6 and 22 m s^{-1} yields $10^3C10 = 0.44 + 0.063\,U10$. However, the Sable Island Thrust anemometer results from the same system (Smith and Banke, 1975) are significantly higher ($10^3C10 = 0.61 + 0.075\,U10$), suggesting that onshore and perhaps shallow water locations, which are included in Garratt's review, may not be representative of the open ocean situation. Within measurement error, the deep water (>50 m) data in Garratt (1977) are generally consistent with (19). The overall average of the 69, near neutral, Reynolds flux CD's at 7.5 m from DeLeonibus (1971) is 1.24×10^{-3} (corresponding to $C10 \approx 1.17 \times 10^{-3}$). Brocks and Krügermeyer (1970) present $C10$'s from their 152, near neutral, 15 min profiles and from the 787, near neutral, 10 min profiles of Hoeber (1969). The averages of 10^3C10 are 1.30 ± 0.18 (4 to 12 m s^{-1}) and 1.23 ± 0.25 (5 to 11 m s^{-1}), respectively. The Bass Strait eddy correlation drag coefficients of Hicks (1972) were corrected for surface drift and if they are reduced by ~7%, they conform to the CDN's of Fig. 6 and the 30 values between 3 and 8 m s^{-1} average 1.13×10^{-3}; however, six runs between 8 and 10 m s^{-1} average $\sim 1.4 \times 10^{-3}$. Garratt also uses the eighteen 10^3C10 eddy correlation values (3 to 11 m s^{-1}) from Hasse (1970), quoting an average of $1.21 \pm 20\%$. Because of flow distortion due to R/V *Flip*, Paulson *et al.* (1972) used an empirical correction factor, chosen to make their average $\langle uw \rangle$ equal to the average of the corresponding Reynolds flux measurements from Pond *et al.* (1971). The latter data (20 runs between 4 and 8 m s^{-1}) give an average CD at ~8 m of 1.52×10^{-3} in slightly unstable conditions. The equivalent average CDN is expected to be about 1.3×10^{-3} and accordingly, Paulson *et al.* find 10^3 CDN equal to 1.32 from 19 profiles.

Recent measurements (10 hours) by Antonia *et al.* (1978) in winds between 5 and 10 m s^{-1} and $-0.1 < Z/L < 0$ give an average 10^3 CD at 5 m of 1.05 and 1.25 (~0.9 and 1.1 at 10 m) from the Reynolds flux and dissipation methods, respectively.

Since a constant CDN is a good fit to most individual data sets over deep water, it is likely to remain a good description of the overall open ocean situation below about 10 m s^{-1}. The "best" constant should be in the range 1.1×10^{-3} to 1.3×10^{-3}. The average 10^3C10's (neutral or near neutral) from *all* the discussed open ocean, eddy correlation, dissipation and profile measurements in winds < 10–12 m s^{-1} are 1.17 (181 runs, which do *not* include our Reynolds flux values which are not independent of our dissipation estimates), 1.14 (628 h) and 1.24 (958 profiles, 188 h), respectively. The overall average is 1.20, but only 1.18 if the results from Hoeber and from Brocks and Krügermeyer are weighted by 1/6 and 1/4, respectively, to make them comparable to hourly values. At higher wind speeds, the vast majority of deep water drag coefficients come from Smith (1980) and Fig. 6 which are both well described by (19). It is an interesting feature of (19), that an extrapolation to 50 m s^{-1} fits the hurricane and wind flume data compiled by Garratt (1977) as well as does a continuation of the Charnock line of Fig. 6.

In the next section, the bulk aerodynamic method [using CDN from (19)] is to be comparted to direct estimates of the momentum flux that include data with $-1.0 < Z/L < 0.3$. The error in $Z/L(\Delta T)$ introduces about a 5% error in CD and the possible 2% error in $\langle U \rangle$ becomes $\pm 4\%$ in $\langle uw \rangle$. The standard deviation in Fig. 6 and the range of other measurements suggest at least a 10% uncertainty in any CDN formulation. The error in individual bulk estimates (13), is, therefore, comparable to the error in other methods (~ ±20%).

6. Time variability of the momentum flux

At any given wind speed the measured momentum flux is highly variable (hence, the scatter in measured CD's) and time histories of the winds and fluxes can be used to explore the sources of this variability. In Figs. 7–9, the momentum flux or Reynolds stress (in N m^{-2}) is calculated from the dissipation (pluses), the Reynolds flux (squares) and the bulk aerodynamic (solid lines) methods. Fig. 7 shows the results when a front was passing the tower. Despite its large range (0.8 to 2.0×10^{-3}), CDN only displays ~ a 10% random variability about a running average over a few hours. As expected, CDN follows the rising wind from hour 8 to 21, but it does not follow the falling wind after

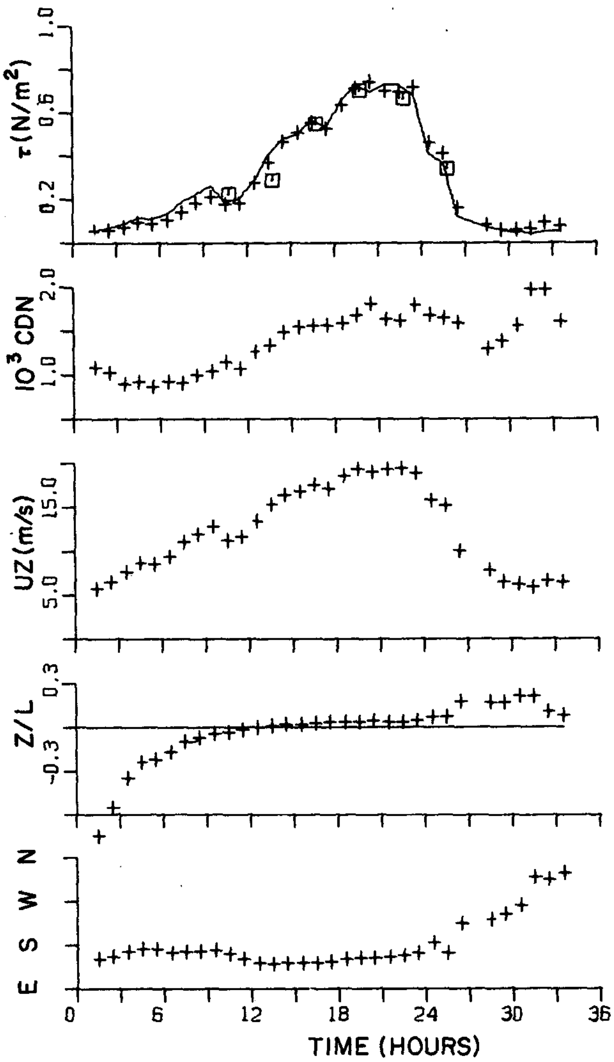

FIG. 7. Time series from 0440 GMT 7 December 1976. Momentum fluxes are from the dissipation (pluses), Reynolds flux (squares) and bulk (solid line) methods.

hour 24. There is almost a complete reversal in wind direction between hours 26 and 35, when the wind speed becomes nearly constant at 6.0 m s^{-1} and this is the situation, previously mentioned, that gives rise to drag coefficients that are much larger than the average for these wind speeds. The time series of Z/L shows the coincident stable stratification, but the high coefficients do appear to be more correlated with the change in wind direction and perhaps to the preceding rapid fall in wind speed. The time series of the stress τ suggests that on the rising wind below 10–12 m s^{-1} the dissipation calculation is systematically less (smaller calculated CDN's) than the bulk calculations, which use an average CDN. An investigation has shown (Large, 1979), that there is an effect, on the average, for winds below ~10–12 m s^{-1}, with CDN from

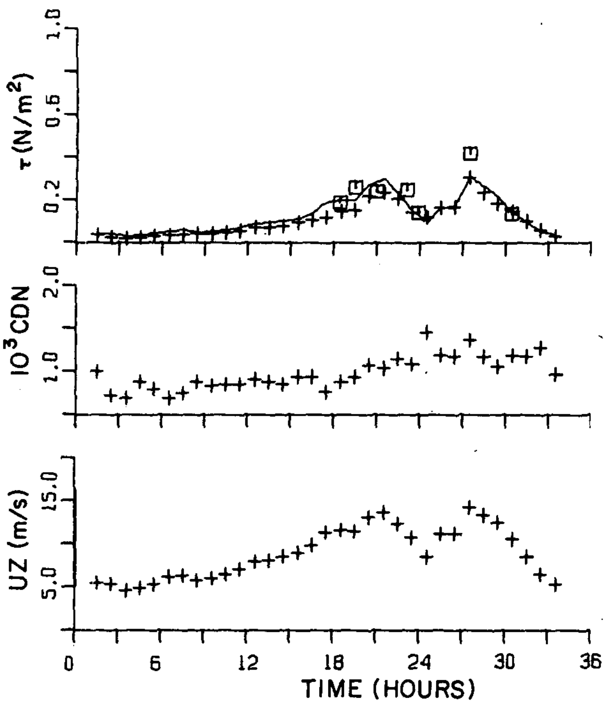

FIG. 8. Time series from 1700 GMT 26 September 1976. Momentum fluxes are from the dissipation (pluses), Reynolds flux (squares) and bulk (solid line) methods.

periods (at the Bedford tower) of falling winds or following a direction change, averaging 20–40% higher than it does when the wind is rising. However, these effects are not important to the stress averaged over the entire 34 hours of Fig. 7, as the cumulative bulk and dissipation stresses differ by less than 1%. A typical September situation is shown in Fig. 8. During the rising wind (to hour 22) the bulk method gives a total momentum input of 2.23 N·h m^{-2}, which is 34% higher than found from the dissipation calculations. Upon reaching the peak winds, the two methods are in closer agreement and the deviation over the whole time series is reduced to 18%. Rapid variations in the mean wind vector are often associated with the passage of low-pressure systems and it was observed (Large, 1979), that following such an event at the Bedford tower, the dissipation measurements over an 18 hour period gave an overall momentum input 12% higher than the bulk calculation. The behavior of the drag coefficient during this time and throughout Figs. 7 and 8 is well described by the observation of Denman and Miyake (1973), that drag coefficients tend to increase on the leading side of a storm, then either remain constant or decrease slightly. With access to wave spectra, they concluded that the drag coefficient was dependent on the nature of the wave field to the order of 20%. It appears likely, therefore, that some of the observed scatter is due to the past history of the wind as "remembered" by the sea surface. When the winds are reasonbly steady, as they were throughout the two day period shown in Fig. 9, hourly averages are long enough for all three methods to be in excellent agreement. During this time, the winds remained westerly (offshore) and they rose from 5 to 12 m s^{-1} by hour 15, then stayed constant until falling to 8 m s^{-1} after hour 43. The range of Z/L was between -0.1 and 0.1. Over this entire period the cumulative momentum input from the dissipation calculations is <3% higher than that found from the bulk formula.

In general the Reynolds flux method (squares) verifies the dissipation calculations, but an exception is found at hour 28 of Fig. 8, when there may have been a feature unresolved by the hourly averaging. The Reynolds flux run is a forty minute average starting fifteen minutes after the hour 28 dissipation average, but in this interval there was sudden 4 m s^{-1} increase in wind speed. The simultaneous dissipation run (used in Fig. 4) is in much better agreement, emphasizing that nearly identical time intervals are required for intercomparisons. A similar situation is found at hour 14 of Fig. 7. Figs. 7 and 8, together, suggest that the Reynolds flux data set (Fig. 3) probably has a bias to small drag coefficients at the lower wind speeds, because most of these runs were collected in September when steadier wind conditions tended to produce small drag coefficients. Later, when the more variable winds produce the high coefficients at wind speeds less than 10 m s^{-1}, the Reynolds flux system was set to record only the higher wind speeds, which is the reason why Reynolds flux calculations are available only between hours 10 and 26 in Fig. 7. As a further consequence, Fig. 3 exhibits a trend with wind speed over the entire range of measurements.

The time series suggest that in steady winds the hourly bulk estimates are good measures of the momentum flux, but with variable winds they may consistently differ over periods up to a day from the dissipation method. It is not obvious which method is the more accurate. Varying winds could

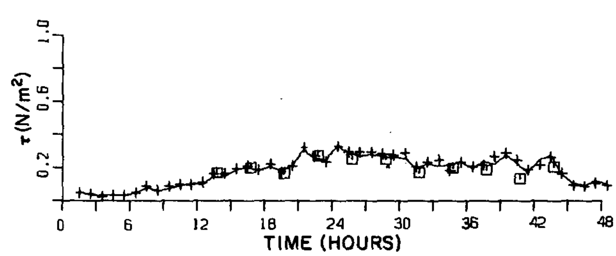

FIG. 9. Momentum flux time series from the dissipation (pluses), Reynolds flux (squares) and bulk (solid line) methods, starting at 1600 GMT 12 March 1977.

produce systematic errors in the dissipation method; however, such behavior would likely result in the scatter of Fig. 5 being greater than the scatter in drag coefficient calculated from other methods (Fig. 3), which is not observed. Consider the two measurements, hour 2 and hour 32 of Fig. 7. Although the wind speeds are about the same (7 m s^{-1}), the dissipation stress estimates differ by a factor of 2. It is felt that errors in the dissipation method are not entirely responsible for this difference and that much of it is real. It is likely that, compared to hour 2, a greater surface roughness, Z_0, at hour 32 (due to varying winds) produced a larger stress and shear (5), but the same $U10$. The resulting difference in the mechanical production, and hence in ϵ, of (9), allowed the dissipation method to distinguish between the two situations, whereas the bulk method (which responds to changes in wind speed only) could not. Since dissipation estimates are based on relatively high frequency spectral estimates, (11) and (9), they ought to be able to follow the hourly variability of the real stress, that is caused by changes in both Z_0 and wind speed. Therefore, over periods such as hours 2 to 22 (Fig. 8) the dissipation estimates (pulses) should be more representative of the behavior of the real stress than the bulk estimates (solid line). A bulk formula based on long-term averages may not, therefore, be the best approach to small-scale or short-term phenomena, such as wave development during a rising wind, but when averaged over a day or two should give good results even if the winds are variable.

7. Conclusion

The momentum flux was measured from the Bedford tower, in winds up to 20 m s^{-1}, by both the Reynolds flux and dissipation methods. The eddy correlation drag coefficients were comparable to those of Smith (1980); however, there was probably a sampling bias to small values at low wind speeds. Universal shapes for the u, w cospectra in the stable and unstable cases were found from averages over the 88 stable and 108 unstable Reynolds flux runs. These curves indicated that the integral from $n = 0.004$ of an unstable and a stable $\Phi uw(f)$ should be multiplied by 1.06 and 1.005, respectively, in order to obtain the momentum flux. This method of integration preserved covariance, on average, and reduced the scatter caused by the uncertain low frequency contributions to the momentum flux. In the near neutral stability regime ($-0.45 < Z/L < 0.05$) the neutral dissipation method could have been used to give good estimates of the momentum flux, without an explicit knowledge of the stability. The corresponding values of u^* were generally within $\pm 20\%$ of simultaneous Reynolds flux measurements. The agreement between the two techniques improved (especially in the more stable cases) when the stability modification to the mean wind profile and the buoyant production of turbulent kinetic energy were incorporated into the dissipation method. These corrections involved the stability parameter, for which a bulk estimate $Z/L(\Delta T)$ was a reasonable approximation, on average. In all but the most stable stratification ($Z/L > 0.05$) the Reynolds flux and the best dissipation estimates of the momentum flux were found to be reliable and in good agreement (4% on average) at all wind speeds. Similar good agreement could be extended through the range $0.05 < Z/L < 0.20$, by using an updated formulation of $\phi_m(Z/L)$. The dissipation method was established as a viable means of measuring the momentum flux in high winds over the open sea.

A favorable comparison of dissipation drag coefficients from the CCGS *Quadra* and the Bedford tower showed the latter to be essentially an open ocean site, allowing a single open ocean data set of 1591 hours to be analyzed. The dependence of the neutral drag coefficient at 10 m on wind speed was approximated by (19). Using the formulation, the bulk aerodynamic method gave good measures of the momentum flux averaged over a day or two and good hourly averages when the wind was steady. Over periods nearing a day, the bulk and dissipation estimates could consistently differ by as much as 30%. Discrepancies were associated with varying winds, with the dissipation estimates being smaller on the rising wind and larger on the falling wind or after a change in wind direction. The surface roughness and hence drag coefficient may depend on sea surface parameters that are a product of both past and present winds.

Over the deep ocean a constant neutral 10 m drag coefficient is an adequate description of most results throughout the wind speed range 4–12 m s^{-1}. A reasonable compromise would be to use 10^3 CDN = 1.2 in the bulk formula for winds below about 11 m s^{-1}. At $U10 = 11$ m s, the high wind speed regression of (19), 10^3 CDN = $0.49 + 0.065U10$, is about 1.2 and since this regression fits the only other large open sea high wind speed data set (Smith, 1980), it should give satisfactory CDN's at wind speeds up to at least 25 m s^{-1}.

If one is interested in periods shorter than a day or so, some allowance could be made for the past history of the wind by using 10^3 CDN = 1.3 for $U10 < 12.5$ m s^{-1} when the wind is decreasing (hourly average speed decreasing continuously and 4 m s^{-1} or more lower than 6 h earlier) or changing direction (60° or more within 2 h sometime in the past 6 h). A value of 10^3 CDN = 1.1 could also be used for $U10 < 9.5$ m s^{-1} for rising winds (hourly average speed increasing continuously and at least 4 m s^{-1} larger than the speed 6 h earlier) and the formulation of the previous paragraph for all other cases.

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