1.5 The Evidence on Momentum Transfer

Meteorologists have expended a great deal of effort on the empirical determination of a momentum transfer law for the air-sea interface. Hundreds of contributions to the subject have appeared in the literature in the past few decades. Some excellent review articles (Donelan, 1990; Garratt, 1977; Smith, 1988) summarize the evidence, and some recent papers describing major observational projects paint a vivid picture of the difficult problems of observation at sea (DeCosmo et al., 1996; Smith et al., 1992; Yelland and Taylor, 1996). Here we can only give a glimpse of this vast effort, prior to examining the evidence on the momentum transfer law.

Dimensional arguments above revealed the potential complexity of such a law, while Charnock's formula for it, distilled from observation, held out the hope of great simplification. We ask now: Does the much larger body of evidence available today, within the accuracy of the observations, contradict Charnock's law? If that law holds only within a range of parameter space, what is that range? What other law should replace it outside that range?

1.5.1 Methods and Problems of Observation

Progress toward a consensus on the momentum transfer law was slow, because observation at sea of any variable is difficult. The earliest estimates of momentum flux over the sea came from the "profile method." This exploited the logarithmic law, by observing wind speed at several levels, and if a substantial logarithmic range could be identified, inferred not only u^*, but also the displacement of the logarithmic line from where it would have been over a smooth solid surface. The displacement defined a roughness length z_0 or alternatively a drag coefficient C_D. The process of fitting a straight line in a semi-log plot to a sparse set of points, provided by observation at a few levels on a spar-buoy, tower, or a ship-mast, is not a particularly accurate process, however. Errors were fairly high, both in the friction velocity and in the roughness length, the latter an exponential function of the observed log-line displacement.

"Direct" observation of the Reynolds flux $\overline{u'w'}$, meaning via two separate instruments at a single level, recording time series of u' and w', runs into other difficulties. It is necessary to average the velocity product over a fairly long time, one hour is common, to make sure that all of the surface layer eddies have been fairly sampled. However, the wind speed itself changes in such a long period: Choice of an averaging period in effect defines what changes are deemed fluctuations, and what slower ones contribute to a "trend" in the average wind speed and Reynolds flux. This problem is common to the determination of any "mean" quantity in turbulent flow, which means extracting it from a long but finite record. Additional problems are the alignment of instruments with the true vertical on a fixed tower, or allowing for ship-mast motions on board ship (in order to ensure that the vertical velocity fluctuations are not contaminated by the horizontal wind). Tower or mast structures may also interfere with the air flow. Upto 20% errors in individual observations of Reynolds fluxes have been common. Yelland et al. (1998) carried out a comprehensive investigation of flow distortion effects on a ship, reporting: “Originally, the four anemometers [on the foremast of the research vessel Charles Darwin] gave drag coefficient values that differed by up to 20% from one to another,” and were up to 60% too high. They also found flow distortion to depend on wind direction relative to the ship.

The “dissipation method” of determining wind stress is perhaps most convenient: It consists of extracting from a time series a velocity spectrum, and determining energy dissipation from the observed spectral amplitudes in the inertial subrange of wavenumbers, with the aid of Equation 1.13. With ε known at some height h above Mean Sea Level, and its distribution supposed governed by the surface layer law, $\varepsilon = u^{*3}/\kappa h$, u^* follows. Because only relatively high frequencies need to be observed, the dissipation method requires shorter averaging periods than the direct observation of $\overline{u'w'}$, therefore allowing more frequent observations on a cruise. Its weakness is, however, that it relies on allowances for stability effects on the energy balance of turbulence, and on the values of some empirical constants in the underlying theory. As Yelland and Taylor’s (1996) careful discussion points out, the method “works” nevertheless, yielding estimates of the Reynolds flux in good agreement with direct correlation measurements, or more accurately, with similar scatter and a not too different mean value.

1.5.2 The Verdict of the Evidence

In light of the difficulties of observation, what is surprising is that a reasonably well-defined momentum transfer law emerges, not that the scatter of individual observations is large. As an example of what a well-conducted major experiment yields, consider the results reported by Yelland and Taylor (1996), and discussed further by Yelland et al. (1998). The observations come from large open stretches of the Southern Ocean at fairly high latitudes, as well as from the subtropical Atlantic near the Azores. The authors employed the dissipation method, and presented drag coefficient and friction velocity against ten-meter wind speed, corrected for buoyancy effects to “neutral” values, observed on thousands of occasions, apparently the largest single data set available on momentum transfer today. Figures 1.6 to 1.9 portray their results, over the moderate to high wind speed range encountered, and separately with higher resolution in low winds. As pointed out before, the friction velocity and the wind speed may be taken to be proxies for the nondimensional parameters $u^*/\sqrt{gh}$ and $U/\sqrt{gh}$.

The friction velocity-wind speed relationship is remarkably tight, as such observations go, (Figure 1.6) defining a momentum transfer law between wind speeds of about 3 and 25 m s^{-1}. At 10 m s^{-1} wind speed, the range of the scatter expressed as U/u^* is about 8, or ± 4 (Figure 1.7). Figure 1.8 shows the averaged drag coefficient-wind speed relationship, the focus of many prior summaries of observations, with Charnock’s law as dotted lines, the lower one with the constant C equal to 11.3, the upper one with 10.2. Data above 6 m s^{-1} fit with few exceptions between these two lines, error bars considered.

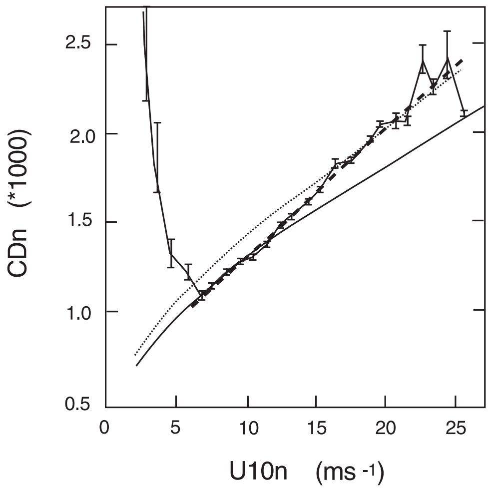

Figure 1.6 Drag coefficient averaged over many observations in the open ocean, at wind speeds above 6 m s^{-1}, with error bars indicating standard deviations. Charnock's law is shown by dotted lines, with constants $C = 11.3$ (lower line) and $C = 10.2$ (upper line). From Yelland and Taylor (1996).

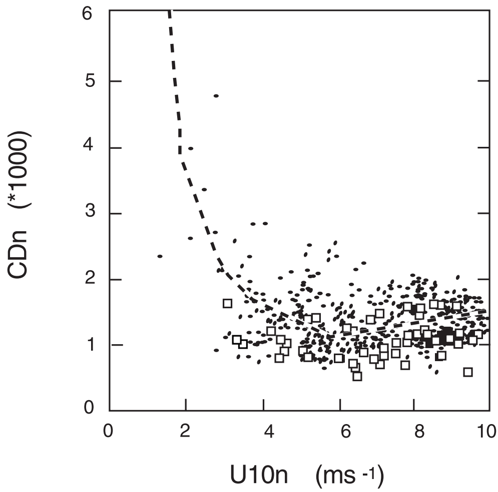

Figure 1.7 Drag coefficient at low wind speeds, individual observations. From Yelland and Taylor (1996).

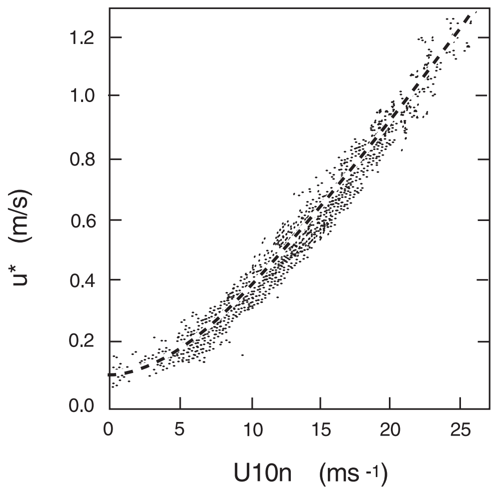

Figure 1.8 Individual observations of friction velocity at moderate winds and higher. From Yelland and Taylor (1996).

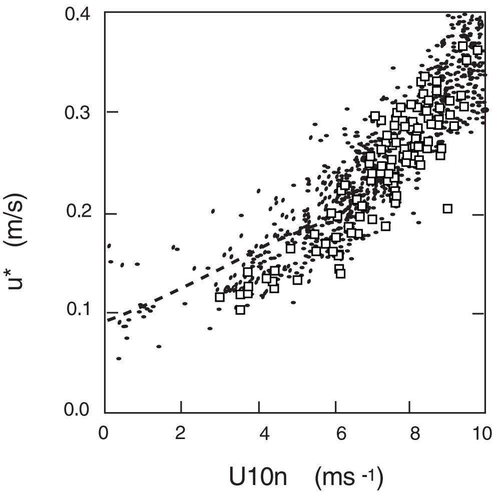

Figure 1.9 Friction velocities at low wind speeds. From Yelland and Taylor (1996).

In a remarkably rigorous further analysis of these results, Yelland et al. (1998) concluded that the drag coefficients reported earlier were on the average 6% too high, owing to airflow distortion by the research vessel. Correcting the data brings them very close to what a prior authoritative summary of momentum flux data found: Smith (1988) confirmed Charnock's law, and yielded the best estimate for C of 11.3. Thus, Yelland and Taylor's averaged data agree with the prior estimate within an error in the U/u^* ratio of less then 1.0, or less than 3%. The averaged data also firmly fix the constant in Charnock's law at $C = 11.3$. Yelland et al.'s (1998) analysis furthermore ties the much larger scatter of individual observations to airflow distortions varying with wind direction. All in all, Smith's (1988) synthesis of earlier information and Yelland et al.'s report on the sources of errors ten years later constitute as good support for Charnock's law as one could possibly get.

Another way to express the verdict of the evidence is that Charnock's law yields the ten-meter wind speed $U(10)$, as a multiple of the friction velocity, with an error of less than one u^*, which is the order of magnitude of the neglected common surface velocity of air and water. The conclusion holds for winds between 6 and 25 m/s, in the open ocean, with corrections applied for buoyancy flux to reduce observed U to the equivalent wind speed under neutral conditions. The corrections for buoyancy flux could not be directly verified, but a fairly wide range of z/L having been covered, the relatively low scatter of the observed U/u^* ratio argues that the corrections "worked."

Charnock's law with buoyancy flux corrections applied thus falls into the category of many other laws of physics valid under certain idealized conditions (steady wind, no swell, etc.), similarly to Hooke's law or the "perfect" gas law, and within a limited range of the Force, the wind speed. Within those limitations, it is accurate to a few percent, in the U/u^* ratio.

The situation today may be contrasted with the confusing state of the subject just a decade ago. A number of different formulae claimed to represent the observations, as individual investigators packaged their conclusions into one coded form or another. Most common were formulae for the drag coefficient $C_D = u^{*2}/U^2$, although many investigators offered recipes for the "roughness" of the sea surface as expressed by z_0. Much of the difficulty in arriving at a consensus was indeed due to this unfortunate focus on z_0, which amplifies errors owing to the exponential dependence of z_0 on the constant in the logarithmic law. According to a study of Blanc (1985), C_D according to ten different formulae then in use varied within wide limits, by about $\pm 20\%$.

1.5.3 Other Influences

At least some of the scatter in observed $U(h)/u^*$ values is bound to be due to genuine influences of other external variables, quantified by the nondimensional parameters in Equation 1.19, but not appearing in Charnock's law. One independent nondimensional variable of potential importance is the Keulegan number, $Ke = u^{*3}/g\nu_a$. This may be regarded as the ratio of the waveheight scale u^{*2}/g that appears in Charnock's law, and the viscous length scale, ν_a/u^*. Keulegan (1949) showed that above a "critical" value of this number instability waves appear on a two-fluid interface. On a wind blown water surface, as several authors reported, short, sharp-crested waves become visible above a wind speed of about 6 m s^{-1}, when $Ke = 100$, or so. This is at the lower limit of validity of Charnock's law, suggesting that in weaker winds the sea surface may behave similarly to a smooth solid wall, above which a viscous sublayer separates the interface from the wall layer.

Outside the viscous sublayer, in the wall layer above a smooth surface, the velocity distribution is:

$$\frac{U(z)}{u^*} = \kappa^{-1} \ln\left(\frac{u^* z}{\nu_a}\right) + 5.7. \tag{1.36}$$

The independent variable here is z divided by the viscous length scale, yielding a Reynolds number. At suitably low wind speeds, when waveheights are small, the smooth law may well apply over the sea and provide a transfer law with $z = h$. This is far from certain, however, horizontal motions are unhindered in a fluid surface, unlike in a solid wall, and eddies in contact with the free surface may behave differently, (e.g., surface tension variations may affect them) modifying the mechanism of viscous momentum transfer.

Experimental evidence on momentum transfer in light winds has been sparse and conflicting until recently. Smith (1988) gave the smooth law as the best representation of the limited data then available in the low wind speed range. Later studies of Geernaert et al. (1987), Bradley et al. (1991), and Yelland and Taylor (1996) have confirmed the validity of the smooth law around a wind speed of 4 m s^{-1}, but showed a much more definite increase with reducing U, see Figure 1.7 above and Figure 1.10 here, from Bradley et al. (1991). In a 1 m s^{-1} mean wind, Bradley et al.'s data cluster around $C_D = 2.5 \times 10^{-3}$, or u^* of 0.05 m s^{-1}, instead of $C_D = 1.56 \times 10^{-3}$, according to the smooth law. Yelland and Taylor's results depart from the smooth law already

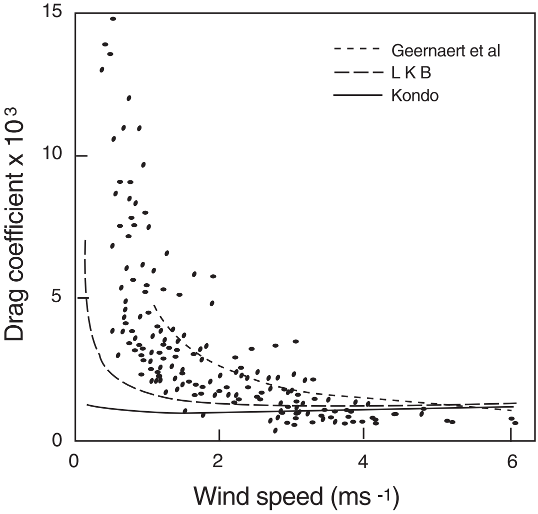

Figure 1.10 Drag coefficients at low wind speeds, from Bradley et al. (1991). The lines show formulae recommended by different authors.

at 3 m s^{-1}, although the observations taken in the subtropical Atlantic remain closer to it than the Southern Ocean ones. The very high drag coefficients at very low wind speeds, reported in all three recent studies, were obtained by the dissipation method, the validity of which under such challenging conditions remains to be confirmed. Another point is that, as the language of meteorological forecasts teaches us, light winds are also variable, so that $U + u'$ may even vary between positive and negative values. The mean wind stress then falls between extremes of the product $\rho|U + u'|(U + u')C_D$, at a higher value than $\rho U^2 C_D$. For $U \ll \sqrt{u'^2}$ it could be much higher.

Even more puzzling than the high drag coefficients at low wind speeds, are some very low ones, found by Sheppard et al. (1972) and Portman (1960), over inland lakes in light winds. The velocities they observed at each level were higher than the smooth law predicts (Figure 1.11), showing the interface to be "supersmooth," at friction velocities less than about 0.18 m s^{-1}, the drag coefficients very low. Similarly, low drag coefficients were reported by Barger et al. (1970) over artificially produced sea slicks in Buzzard's Bay in Massachusetts, that depressed the surface tension of the water surface. The "supersmooth" character of the water surface in these observations may have been the result of energy drain from air-side turbulence to the free surface, similar to the effect of drag-reducing chemicals on boundary layers over solid surfaces. Exactly how, and under precisely what conditions, such drag reduction over water surfaces occurs, is not clear. Much thus remains to be learned about momentum transfer in very light winds.

One other factor that may cause significant departures from Charnock's law is "wave age," quantified by C_p/u^*. Field studies of wave age effects are of recent date, however, and the results have remained controversial for some time. A major field study off the Dutch coast code-named HEXMAX (the main experiment of the Humidity Exchange

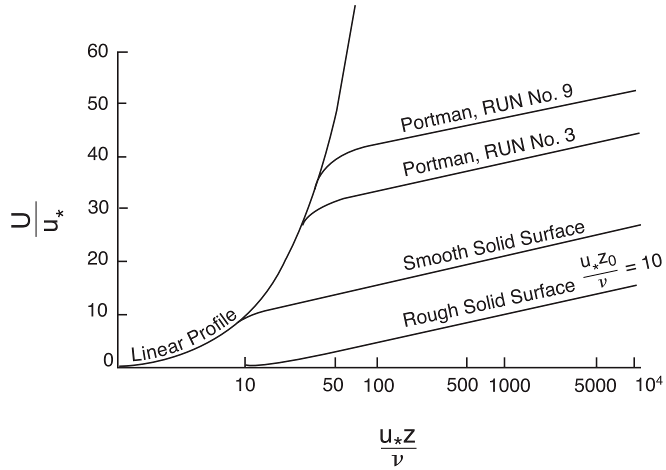

Figure 1.11 Logarithmic velocity distributions in light winds over "supersmooth" surfaces in Lake Michigan, observed by Portman (1960), compared with typical distributions over a smooth and rough solid surface. From Csanady (1974).

over the Sea, HEXOS, program, see Maat et al., 1991 and Smith et al., 1992) led to the conclusion that the constant C in Charnock's law is constant only at long fetch (i.e., C_p/u^* around 30). At younger age it varies according to the HEXOS results as:

$$C = 1.83 + \kappa^{-1} \ln(C_p/u^*) \tag{1.37}$$

supposedly valid within the range of C_p/u^* from 4 to 14 or so. Smith et al. state this result in terms of the nondimensional roughness length as: $gz_0/u^{*2} = 0.48/(C_p/u^*)$.

Figure 1.12 from Donelan et al. (1993), shows the data underlying this law, in a log-log plot. The ordinate is proportional to $-C$, if the logarithmic scale is replaced by a linear one ($C = 11.5$ at $gz_0/u^{*2} = 10^{-2}$, decreasing by 5.76 with each decade, to a minimum of about $C = 5$ at the highest observed points). On the abscissa, the logarithmic scale of C_p/u^* unduly compresses the explored range of this variable. At the high C_p/u^* end at the left, the large cluster of badly scattered C-values comes from mature waves, apparently subject to many unexplained influences. At rather lower wave age, the HEXOS law, Equation 1.37, supposedly applies as far as the field data reach, to about $C_p/u^* = 4$. The scatter is considerable, however. Only laboratory data are available at still lower C_p/u^*, seen on the right of Figure 1.12: they seem to represent a different population of randomly scattering values. This is how Donelan et al. (1993) interpret the data, contending that laboratory waves differ qualitatively from field waves. Toba et al. (1990) argue, on the other hand, that laboratory waves correctly portray the roughness of sea surface at very low C_p/u^*. In that case, according to Figure 1.12, the "constant" C in Charnock's law first increases then decreases with increasing C_p/u^*.

The solution of this conundrum seems to be that the wave age dependence of wind stress found in the HEXOS project was in reality an effect of long waves shoaling and

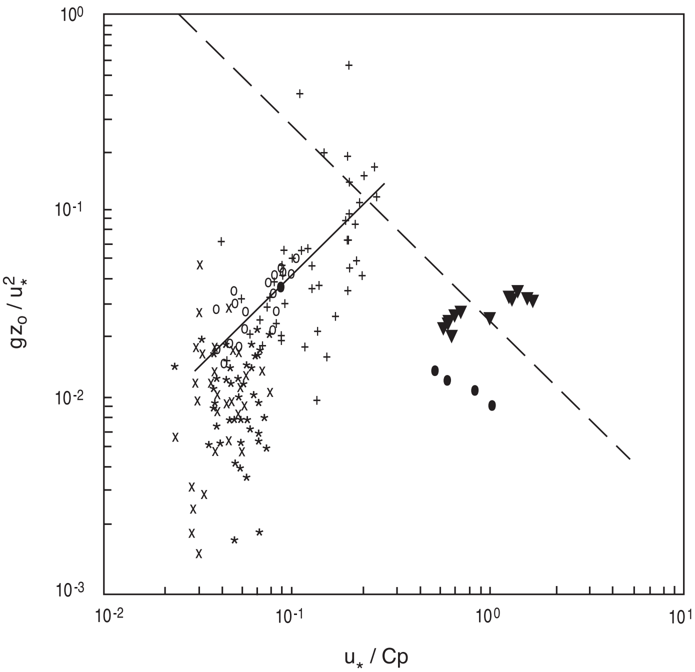

Figure 1.12 Nondimensional roughness gz_0/u^{*2} versus inverse wave age u^*/C_p, representing velocity distributions observed over the windsea. The lines show supposed functional relationships deduced from these data by two different groups of investigators. From Donelan et al. (1993).

steepening in shallow water (Oost 1997), and that it does not apply in the open ocean. As Yelland et al. (1998) stated in their report on flow distortion effects on Yelland and Taylor's (1996) data: "This study examined open ocean drag coefficient measurements for evidence of significant anomalies that can be related to the sea state or wave age. None were found." Dobson et al. (1994) reached a similar conclusion. Yelland and Taylor (1996) also point out that the large scatter of data on wind stress is likely due to flow distortion varying with wind direction relative to the ship or other platform of observation.

This completes our discussion of the momentum transfer law of the air-sea interface. Charnock's law corrected for buoyancy flux emerges as an empirical formula valid within close error bars in moderate to fairly strong winds. We turn next to transfer laws for scalar properties.

1.6 Sensible and Latent Heat Transfer

The air-sea interface is the seat of a whole array of complex thermodynamic interactions: it receives energetic short-wave solar radiation but reflects a large fraction of