Part Four

Ocean and Atmosphere

17 Air–Sea Interaction

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17.1 Introduction

In preparing a chapter for a book such as this some nostalgia is excusable, even inevitable. For those who came to the subject, or came back to the subject, after the war there was a bewildering range of work going on. Barber and Ursell were publishing their results on the long-distance propagation of ocean swell, having used an analogue device to estimate the wave spectrum. Sverdrup and Munk had done wartime work on waves too, but by 1947 Munk was writing on a possible critical velocity for air–sea transfer processes and Sverdrup was working up his classical paper on currents driven by the curl of the wind stress. Jacobs was continuing his long-term study of the climatology of energy exchange between sea and air: Budyko was just starting his. Sheppard was publishing his direct determination of the shearing stress by use of a drag plate and Roll was making new wind-profile measurements over the Wattenmeer. Obukhov had already developed the dimensional arguments leading to the Monin–Obukhov length and had contributed to the Kolmogoroff small-scale similarity hypothesis with Onsager and Weizsäcker: in Cambridge, Batchelor was exploring its consequences. Priestley was off to Australia to set up a powerful group on near-surface turbulence, and was making pioneering calculations of the poleward heat and momentum transfer by covariance of wind and temperature fluctuations. Eady was in London working up his idea about baroclinic instability, Charney his at Princeton. Henry Stommel, relatively recently at Woods Hole, was interested in convection in the atmosphere and ocean (it was the time of the Woodcock-Wyman expedition) and had discovered the phenomenon of entrainment into cumulus clouds.

The importance of air–sea interaction to the larger-scale flows of the atmosphere and ocean was in no doubt, though it was a somewhat minority interest. Most of the work at that time concerned the estimation of the surface fluxes of heat, water vapor, and momentum from the only data base then foreseen, namely, the routine observations of temperature (dry bulb, wet bulb, sea) and wind made from the merchant vessels that reported to national meteorological agencies. Given suitable formulas it was thought that one could perhaps calculate the poleward heat transfer by the ocean and make some progress on relating winds to near-surface currents.

There were obvious difficulties of observation over the sea rather than the land but these were compensated for by the importance of the results and by the relative uniformity of the surface, both in space and, due to the high thermal capacity of the ocean, in time. Also the problem was close enough to a laboratory shear flow to allow comparisons with flow in pipes and channels. Much of the early work therefore concerned itself with the fluid mechanics of the air flow over the sea to a height of, say, 50 m. Since then the concept of air–sea interaction has been much broadened to include consideration of phenomena on larger space and time scales. General problems such as the teleconnections between sea surface temperature anomalies and subsequent weather patterns and specific aspects such as El Niño have been included. One thinks of climate as a complex of interactions between the air, the sea, and the surface of the earth, and in this sense air–sea interaction can be argued to include much of the physics and dynamics of the atmosphere and the ocean. But this review will consider only the small-scale processes by which heat, water, and momentum are transferred near the sea surface. In a fundamental sense the air and the sea interact only in a thin interfacial layer, but it is convenient to consider processes confined to the coupled boundary layers of the atmosphere and the ocean, which, as will be seen, extend typically to a height of 1000 m and a depth of 30 m from the sea surface.

The first section deals with the surface layer of the atmosphere, which constitutes about the lowest tenth of the whole atmospheric boundary layer. This is the only region for which a satisfactory (though empirical) treatment is available in the form of "similarity theory" that relates small-scale properties of the airflow (gradients, turbulence spectra) to the vertical fluxes of momentum, heat, and water vapor.

To get the mean profiles, or the exchange coefficients (which are important in practice), requires boundary conditions within the interfacial layer. These are not at all well understood—observational results are briefly summarized in section 17.3.

Many of the difficulties associated with the interfacial layer are due to the complications introduced by surface waves. The relation between the wind stress (or the aerodynamic roughness) and the surface wave field (or the geometrical roughness) has proved an intransigent problem. Recent advances in our knowledge of the wave spectrum, and of the pressure distribution at the moving sea surface, are indicated in section 17.4.

The development of computer models of the atmosphere, and increasingly of the atmosphere and ocean combined, have much reduced the emphasis on the near-surface meteorological variables. The surface fluxes are no longer related to ships' observations so much as to winds, temperatures, and humidities in the atmosphere and ocean at levels where the flow can be taken to be frictionless and adiabatic. This requires increased understanding of the structure of the boundary layer as a whole. Section 17.5 describes our regrettably limited knowledge of the climatology of the atmospheric boundary layer and of the complicated processes that affect the distribution of density and wind within it. Some of the processes are similar to those that determine the structure of the oceanic boundary layer: for others, such as clouds, there is no obvious analogy.

17.2 The Surface Layer

The lowest 50 m of the boundary layer of the atmosphere has a special importance and simplicity that together with its accessibility have attracted intensive study. The importance of the surface layer comes from the fact that although its depth is only a small fraction of the whole boundary layer, it is within it that most of the change of wind speed, temperature, and humidity between the free atmosphere and the surface takes place. Its simplicity comes about because the fluxes of momentum, heat, and water vapor undergo only small fractional changes within the surface layer, so they may commonly be regarded as independent of height. For this reason it is convenient to take the fluxes of momentum and potential density as the basic independent variables governing the motion, and to consider the mean gradients and all the properties of the turbulence as being determined by them.

17.2.1 Near-Surface Profiles in Neutral Conditions

Many measurements of the vertical profile of velocity have been made over sites uniform for an upwind distance great compared to the height of observation z in conditions steady for times greater than z/u_*, where u_* is the friction velocity defined by $\tau_0 \equiv \rho u_*^2$, τ_0 being the surface shearing stress and ρ the air density.

If the potential density is independent of height (neutral hydrostatic stability) the velocity gradient is found to vary quite accurately as the inverse of the height measured from a reference plane near the top of the roughness elements and

$$\frac{dU}{dz} = \frac{u_*}{\kappa z}, \qquad (17.1)$$

where κ is constant.

It is easy to see that (17.1) is a reasonable relation, though the "proofs" of it to be found in the literature are to be treated with caution. If, away from the surface, the turbulent motion is not affected by viscosity or other processes by which the stress is communicated to the surface, nor by the fact that the boundary layer is of finite thickness, but has its intensity and scale determined by the Reynolds stress and the height, then (17.1) follows on dimensional grounds. It is written in terms of dU/dz rather than U because a uniform translation can have no effect on the internal dynamics of the flow.

The profile of a transferable scalar such as potential temperature, specific humidity, or the concentration of gases such as carbon dioxide can be treated similarly. Limiting the discussion again to neutral hydrostatic stability means that any variation in temperature or humidity must be small or combined in such a way as to maintain the potential density independent of height. As for momentum the vertical transfer by the turbulence will be governed by u_* and z: the potential temperature profile will be given by

$$\frac{d\Theta}{dz} = \frac{\theta_*}{\alpha_0 \kappa z}, \tag{17.2}$$

where θ_* is a scale temperature defined by $u_*\theta_* \equiv \langle w\theta \rangle$, w being vertical velocity, and the constant α_0 is introduced to allow for the possibility that the transfer of a scalar quantity may differ from that of momentum.

In a similar way the humidity profile is given by

$$\frac{dQ}{dz} = \frac{q_*}{\alpha_0 \kappa z} \tag{17.3}$$

with $u_*q_* \equiv \langle wq \rangle$. The same constant α_0 is used because it seems unlikely that different scalars will have different transfer properties in fully turbulent flow.

In (17.2) and (17.3), Θ is the mean temperature and Q the mean humidity, θ and q being the respective fluctuating quantities. The covariances $\langle w\theta \rangle$ and $\langle wq \rangle$ measure the heat flux H and the water vapor flux E as

$$H = c_p \rho \langle w\theta \rangle \qquad \text{and} \quad E = \rho \langle wq \rangle$$

by analogy with the Reynolds stress $\tau = -\rho\langle wu \rangle$ (c_p is the specific heat at constant pressure, and u is the horizontal component of velocity).

It is perhaps surprising that the mean gradients are unaffected by the characteristics of the surface—one might expect the expressions for them to be valid only at heights large compared with some height typical of the surface geometry. But by choosing the zero plane suitably, often just below the tops of the surface-roughness elements, the formulas fit quite well down to heights only just above them.

17.2.2 Near-Surface Profiles in Nonneutral Conditions

It has been known for a long time that a vertical gradient of potential density can have a profound effect on turbulence (Richardson, 1920). When the density increases upward, so that the mean situation would be statically unstable, the mixing action of the turbulence produces a downward density flux, and buoyancy forces feed energy into the turbulence so as to augment the action of the windstress. So in unstable conditions, for a given value of the shear stress, the turbulence will be more vigorous and its ability to transfer heat and momentum greater than in neutral conditions. In stable conditions the converse is true.

How the buoyancy forces operate is only partly understood, although some theories based on the insertion of simple physical approximations into the Friedman–Keller equations for the variances and covariances of the velocity components and the density have had considerable success. It may be noted that since the work done by the buoyancy forces involves a product of their magnitude with the distance over which they operate, their effect is most pronounced on the large scales of motion. Hence one expects large eddies to be preferentially destroyed in stable conditions and preferentially sustained in unstable conditions: the scale of the most active part of the turbulence will be smaller in stable conditions than in unstable. Also, since the scale of the motion decreases as the surface is approached, so also does the effect of the buoyancy. It follows that sufficiently near the surface the active part of the motion is governed by the laws appropriate to neutral conditions.

In the surface layer great simplification has been achieved by the use of dimensional arguments to develop what is called "the similarity theory of the surface layer." It applies to the components of the motion that have scales smaller than the depth of the surface layer and so are generated and controlled within it. Recognizing that the fluxes of momentum and potential density are nearly independent of height in the surface layer, and that the mean gradients are unaffected by the detailed transfer processes at the boundary, Russian workers (Obukhov, 1946; Monin and Obukhov, 1954) were led to use the fluxes as key quantities in the surface layer. This was an imaginative development, at a time when fluxes were much harder to measure than mean gradients: it has provided a very useful means of systematizing many varied observations.

The assumption is made that turbulent quantities in the surface layer are unaffected by all quantities external to it, such as the total thickness of the boundary layer and detailed transfer processes at the surface. The basis of the theory is to use as the independent variables z, u_*, and δ_* (defined by $\delta_* u_* \equiv \langle \delta w \rangle$, where δ is the buoyancy fluctuation). All the properties of the turbulence are expressed in terms of them.

From these variables only one dimensionless group can be found. It is

$$\zeta = z/L, \qquad \text{where} \quad L = u_*^2/\kappa\delta_* . \tag{17.4}$$

L is called the Monin–Obukhov length after the originators of the theory. κ has been introduced because they included it in their initial definition.

It follows that all dimensionless properties of the turbulence must depend solely on ζ. In particular the dimensionless velocity profile will be a function of ζ alone,

$$\frac{\kappa z}{u_*}\frac{dU}{dz} = \phi_M(\zeta), \tag{17.5}$$

as will be the profile of a transferable scalar

$$\frac{\kappa z}{\theta_*}\frac{d\theta}{dz} = \phi_H(\zeta). \tag{17.6}$$

17.2.3 Alternative Stability Parameters

The use of the Monin-Obukhov length has the disadvantage that it requires knowledge of the fluxes, which is not always available. It is sometimes more convenient to work with the gradient form of the Richardson number *Ri*, which is defined by

$$Ri = \frac{g}{\rho}\frac{d\rho}{dz}\bigg/\left(\frac{dU}{dz}\right)^2, \tag{17.7}$$

g being the acceleration due to gravity. According to the similarity theory *Ri* should be a universal function of ζ in the surface layer.

In his original paper Richardson (1920) showed that the rate per unit mass at which work had to be done by the turbulence against buoyancy forces was $\delta_* u_*$. He pointed out that the ratio of this to the rate at which the shear stress produced turbulent energy, $u_*^2(dU/dz)$, could not exceed unity unless energy was being brought into the region from outside. This ratio is now called the flux Richardson number, *Rf*, and is related to *Ri*, to ϕ_M, and to ζ by

$$Rf = \delta_*\bigg/\left(u_*\frac{dU}{dz}\right) = \alpha Ri = \zeta/\phi_M, \tag{17.8}$$

where $\alpha = \phi_M/\phi_H$ is itself a function of *Ri* or ζ.

It may be noted that only two empirical functions are needed to describe mean profiles and that the relations between the variables enable all the functions required in connection with mean profiles to be derived from whichever two functions can be conveniently measured.

17.2.4 Flux-Gradient Observations

To verify (17.1) for the mean-velocity profile in neutral conditions and to determine κ, it is necessary to measure the surface stress. From laboratory pipe measurements it was known that $\kappa \simeq 0.4$, and the measurements of Sheppard (1947), who simulated an area of ground surface and measured the forces on it with a spring balance, confirmed that the same value applied to the lower atmosphere. Later measurements using the drag-plate technique have given excellent results in suitable conditions.

A second method is to measure fluctuations of the horizontal u and vertical w components of the air flow. The turbulent stress is (nearly) $-\rho\langle uw\rangle$. It is necessary to use a fast-response instrument that responds to the whole range of frequencies contributing to the stress, and to have computer processing for the spectra, covariance, etc., but several workers have succeeded in producing consistent results. Such techniques can also be used, given measurement of fluctuating temperature and humidity, to estimate heat and water-vapor fluxes from $\langle w\theta\rangle$ and $\langle wq\rangle$.

Such rapid-response devices and analysis facilities can also be used to estimate the dissipation rates for turbulent energy, temperature variance, and humidity variance. In suitable conditions these can be related to the respective fluxes.

In spite of a good deal of work the value of κ is not universally agreed upon: this is partly due to the difficulty of allowing for fluctuations in the surface stress. When this is taken into account u_* in (17.1) should be replaced by its mean value $\langle(\tau/\rho)^{1/2}\rangle$, which is less than $\langle\tau/\rho\rangle^{1/2}$.

Pruitt, Morgan, and Laurence (1973) made careful measurements using a large drag plate to determine that $\kappa = 0.42$; allowing for a slight overestimate due to fluctuating stress it seems that the generally accepted value, $\kappa = 0.40$, is not far wrong.

There is not space here to deal adequately with the many observations that have been made, particularly over land, which have established the forms of ϕ_M and ϕ_H. They have been reviewed by Plate (1971), Monin and Yaglom (1965, 1967), and Högström (1974). According to Busch (1977), most atmospheric data are well represented by

$$\phi_M = \begin{cases} 1 + 5\zeta & (\zeta \ge 0) \\ (1 - 15\zeta)^{-1/4} & (\zeta \le 0), \end{cases}$$

$$\frac{\phi_H}{\phi_H^{(0)}} = \frac{\phi_{\mathrm{OTS}}}{\phi_0} = \begin{cases} 1 + 6\zeta & (\zeta \ge 0) \\ (1 - 9\zeta)^{-1/2} & (\zeta \le 0), \end{cases} \tag{17.9}$$

where OTS means other transferable scalar. The value of $\phi_H^{(0)}$ $(= \alpha_0^{-1})$ values are scattered, but a reasonable value is 0.8.

There are some plausible arguments to support these forms but no satisfactory theory. Nevertheless it seems clear that the similarity theory of the surface layer provides an excellent framework in which to systematize observational studies of mean gradients and of turbulent fluctuating quantities in the surface layer. The basis of the theory is that such quantities are unaffected by the characteristics of the underlying surface, so the results so far given apply over both land and sea: the only requirement (by no means easy to satisfy) is that of uniformity in space and steadiness in time.

17.3 The Lower Boundary

Previous expressions for profiles have been written in terms of gradients, like dU/dz, because by the hypotheses used a uniform translation can have no bearing on the internal structure of the flow. To integrate, so as to get an expression for, say, $U(z)$, requires a boundary condition within what can be called the *interfacial layer,* which includes the surface itself and the air above up to a height comparable to that of the elements that make up the surface. Processes in this layer are complicated and not well understood.

On integration of (17.1) we have for the wind profile in neutral conditions

$$\kappa \frac{U}{u_*} = \ln\left(\frac{z}{z_s}\right) \tag{17.10}$$

z_s is merely a constant of integration, whose value is determined by the surface geometry and surface processes. It has no influence on the internal dynamics of the flow.

Similarly, integration of (17.2) and (17.3) for the temperature and humidity profiles in neutral conditions gives

$$\alpha_0\kappa(\Theta - \Theta_0) = \theta_* \ln(z/z_\theta), \tag{17.11}$$

$$\alpha_0\kappa(Q_0 - Q) = q_* \ln(z/z_q) \tag{17.12}$$

where Θ_0, Q_0 are the potential temperature and the humdidity at the surface and z_θ, z_q are constants of integration analogous to z_s. Like z_s they have no influence on the internal structure of the flow: changing them has the effect of adding a constant amount to the temperature and the humidity.

Turning to the more complicated expressions, (17.5) and (17.6), and integrating to get the profiles of velocity and potential temperature in nonneutral conditions we have

$$\kappa U/u_* = \int_{z_0}^{z} (\phi_M/z')\,dz'$$

$$= f_M(\zeta) + \phi_M(0)\ln(z/z_0)$$

where

$$f_M(\zeta) = \int_0^{\zeta} [(\phi_M - \phi_0)/\zeta']\,d\zeta'. \tag{17.13}$$

The lower limit has been taken as zero instead of z_0/L since $z_0 \ll L$ and ϕ_M is assumed continuous at the origin.

Thus the departure from the neutral logarithmic form is given by f_M with ζ positive for stable, and negative for unstable conditions. All the stable profiles are similar to each other, as are all the unstable ones, the neutral profile being a limiting case.

Analogously,

$$\kappa(\Theta - \Theta_0)/\theta_* = f_H(\zeta) + \phi_H(0)\ln(z/z_\theta) \tag{17.14}$$

where

$$f_H(\zeta) = \int_0^{\zeta} [(\phi_H - \phi_H(0))/\zeta']\,d\zeta'.$$

These rather formal results can be summarized by remarking that profiles like U/u_* are functions of ζ $(= z/L)$ and of z/z_0. The basic requirement of the similarity theory of the surface layer is that the internal dynamics of the flow is unaffected by the boundary processes so that

$$\kappa U/u_* = f(\zeta, z/z_0) = f_1(\zeta) + f_2(z/z_0)$$

$$= f_M(\zeta) + \ln(z/z_0).$$

The basic unknowns in the problem are those in the interfacial layer, represented by z_0, z_θ, z_q.

17.3.1 Transfer Coefficients over the Sea

So far as the relation between stress and velocity gradient is concerned, (17.1) indicates that the turbulence acts as an effective (eddy) viscosity of magnitude

$$K_M = \kappa u_* z.$$

This is usually much greater than the molecular viscosity ν, but below a height $\nu/\kappa u_*$ it is smaller, and molecular transfer will dominate the motion. If the surface is fairly smooth, so that the typical height of the roughness elements h_r is smaller than this, they will be submerged in the viscous layer and play little part in communicating stress to the surface. The flow is then said to be aerodynamically smooth (though in fact it is fully turbulent), and, since h_r is irrelevant, dimensional reasoning gives

$u_* z_s/\nu = \text{constant} = 0.11$ by observation.

If, on the other hand, $h_r \gg \nu/\kappa u_*$, the stress is communicated to the surface by the form drag of the roughness elements. Then the molecular viscosity is irrelevant and

$$z_s = z_0,$$

where the so-called roughness length z_0 depends in a complicated way on the size, shape, and spacing of the roughness elements. There is no good theory for relating z_0 and h_r: typically for close-packed granular roughness elements $z_0 = h_r/30$.

In the intermediate case $h_r \simeq \nu/\kappa u_*$, z_s/h_r is a function of $u_* h_r/\nu$ that is known from laboratory observation.

A complication is that the wind can modify the geometry of the surface over which it blows. Long grass is flattened by the wind, and E. L. Deacon (1953) observed that z_0 for grass 700 mm long falls from 90 mm in light winds to less than 40 mm in strong winds. On the other hand, when particles from the surface are carried into the air, as in blowing sand or snow, the value of z_0 is much larger than for the undisturbed surface.

The most important surface whose geometry is affected by the wind is the ocean. Its aerodynamic roughness has been the subject of much research over the last decades.

At very low wind speeds, before waves or ripples have been generated, the sea would be expected to behave as an aerodynamically smooth surface, and this is generally observed. When waves appear, the wind profile in neutral conditions remains closely logarithmic down to levels close to the surface, but the effective roughness length increases from the aerodynamically smooth value. From a series of careful wind-profile measurements over an artificial lake, and taking into account earlier observations, Charnock (1955) suggested that the aerodynamic roughness length was determined by the shearing stress, and used the simplest nondimensional relation

$$z_0 = \alpha_1 u_*^2/g. \qquad (17.15)$$

The same expression but with a different value for α_1 had been found by Ellison (1956) using observations reported by Hay (1955). The implication of such a formula is that, while z_0 depends in a complicated way on the waves generated, the wave structure in turn is determined by the stress on the surface. The coefficient α_1 is at most a weak function of the faster, and so of the larger, waves, with the possible implication that the stress is transmitted locally, and to the short waves and ripples. This raises the question why g is used in (17.15) rather than other properties of the fluid such as its viscosity or surface tension. Lengths can be formed using u_* and ν (as in aerodynamically smooth flow), and using the surface tension S and u_*: in both cases the lengths decrease as u_* increases, so it is not likely that z_0 depends on ν or S in any simple way. But the fluid mechanics of the wavy surface is complicated and no adequate theory exists.

The usefulness of (17.15) is tested by observation, and here there has been considerable difference of opinion. Observations of the surface stress over the sea have been made by numerous workers, using various methods. The most common methods have been the use of the wind profile and eddy correlation. Most workers have preferred to express their results in terms of a drag coefficient C_D given by

$$\tau_0 = C_D(10)\rho U_{10}^2$$

the 10 being inserted as a reminder that the value of C_D depends on the height at which U is measured: 10 m is commonly used. U_{10}, and so $C_D(10)$, is also affected by the static stability, but this can be allowed for using similarity theory:

$$C_D = \frac{C_{DN}}{[1 + \kappa^{-1}C_{DN}^{1/2}f_M(\zeta)]^2},$$

where f_M is given by (17.13) and

$$C_{DN} = \frac{\kappa^2}{(\ln z/z_0)^2}$$

is the neutral drag coefficient.

Garratt (1977) has recently made a thorough review of previously reported values of C_{DN} in relation to U_{10}. Until about 1970 the values were scattered (though less so than they were 20 years before—see Charnock, 1951). They are shown in figure 17.1 and table 17.1. Since 1970 many more observations have been reported, using better methods, and Garratt has estimated C_{DN} from the 17 publications listed in table 17.2, excluding some, for reasons detailed in his paper. The resulting values are plotted in figure 17.2, in which reasonable agreement with (17.15) is shown, though the considerable scatter increases at wind speed greater than 15 m s^{-1}.

Some authors have estimated the surface stress in hurricanes by integrating the ageostrophic wind component. These are given in table 17.3 and figure 17.3 (again due to Garratt, 1977): there is some indication that (17.15) is satisfied in winds up to 50 m s^{-1}. Garratt gives $\alpha_1 = 0.0144$ as an acceptable value.

It seems from Garratt's review that (17.15) is sufficient for many purposes. But its physical basis is still very unsatisfactory: the implied roughness lengths are small ($\sim 10^{-1}$ mm), and we have no clear idea as to how they are determined. That the high-wavenumber range of the wave spectrum is involved seems probable, and is supported by experiments using surface films and detergents that eliminate short waves and much reduce the drag for a given wind.

Our knowledge of z_θ, z_q, and the physical properties on which they depend is even less satisfactory. Owen and Thompson (1963) have put forward a theoretical framework that allows comparisons between measurements of heat and of vapor transports from fixed rough surfaces. They give a formula that, assuming α_0 [equation (17.3)] to be 1.3, becomes

$$\ln(z_0/z_\theta) = 2.0Pr^{0.75}(u_* z_0/\nu)^{0.33}, \qquad (17.16)$$

though the numbers are tentative. $Pr = \nu/\nu_T$ where ν_T is the kinematic molecular diffusivity for the property being transferred. Fortunately $u_* z_0/\nu$ is small over the ocean so $z_0 \simeq z_\theta$ is a reasonable approximation. But if a formula like (17.16) does apply, and if z_0 is given by (17.15), then z_θ will gradually become less than z_0 as u_* increases. Kitaigorodskii (1970) gives a critical review of existing observations, as do Friehe and Schmitt (1976) and Busch (1977), but the experimental scatter makes it difficult to generalize.

Table 17.1 Main Reviews of the Neutral Drag Coefficient over the Sea[a]

Source	Wind speed range (m s^{-1})	$C_{DN}(10)$ ($\times 10^3$)	Variability (%)	Number of references
A. Priestley (1951)	2.5–12 strong	1.25[b] 2.6[c]	?	Not stated
B. Wilson (1960)	~1–5 9–20	1.42 2.37	±50 ±25	47
C. Deacon and Webb (1962)	2.5–13	$1 + 0.07\,V$	±25–50	9
D. Robinson (1966)	3–8.5 2.5–14	1.8[d] 1.48[e]	±30 ±15	14
E. Wu (1969b)	3–15 15–21	$0.5\,V^{0.5}$[f] 2.5	±30 ±10	30
F. Hidy (1972)	2–10	1.5	±30	8

a. Showing wind speed range, best estimate of $C_{DN}(10)$ (either as a constant or a function of wind speed), and typical data variability as a percentage of $C_{DN}(10)$ value over the wind speed range considered (see figure 17.1). [After Garratt (1977), who summarized the reviews.]
b. Actually based on Deacon (1950): *Nature* 165, p. 173.
c. Quotes Sverdrup et al. (1942) and Munk (1947).
d. Micrometeorological data.
e. Geostrophic departure.
f. Overall variation close to Charnock relation with $\alpha = 0.016$.

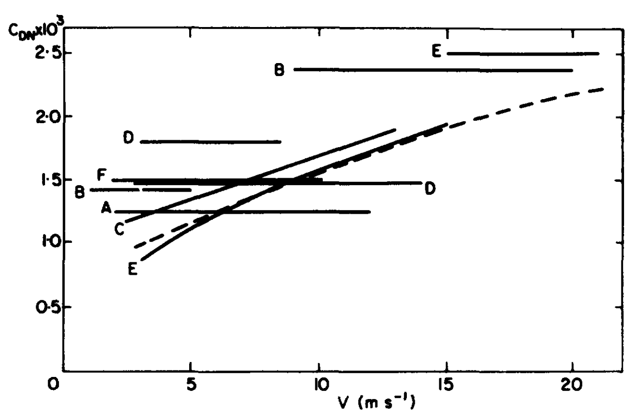

Figure 17.1 Mean curves of $C_{DN}(10)$ plotted against V (10 m) for review sources shown in table 17.1. Dashed curve is based on $z_0 = \alpha u_*^2/g$ with $\alpha = 0.016$ and $\kappa = 0.41$.(Garratt, 1977.)

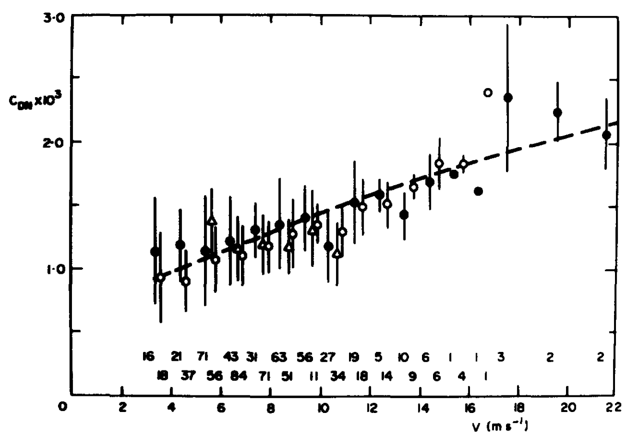

Figure 17.2 Neutral drag coefficient values as a function of wind speed at 10-m height, based on individual data taken from the recent literature (see table 17.2 and Garratt, 1977). Mean values are shown for 1-m-s^{-1} intervals based on the eddy correlation method (●) and wind profile method (○); Hoeber's wind profile data are also shown (△). Vertical bars refer to the standard deviation of individual data for each mean, with the number of data used in each 1-m-s^{-1} interval shown above the abscissa axis: top line refers to (●), bottom line to (○). The dashed curve represents the variation of $C_{DN}(10)$ with V based on $z_0 = \alpha u_*^2/g$ with $\alpha = 0.0144$. (Garratt, 1977.)

Table 17.2 Neutral Drag Coefficient Values over the Ocean[a]

Source	Wind speed range ($m\,s^{-1}$)	$C_{DN}(10)$ ($\times 10^3$)	Variability σ (%)	Number of data (n)	Method	Platform	Comments
1. Smith and Banke (1975)	2.5–21	$0.63+0.066V$	30	111	ec	Mast	Also utilizes data of Smith (1973) using thrust and sonic anemometers
2. Kondo (1975)	3–16	$1.2+0.025V$	15	—	waves	Tower	Utilizes data on wave amplitudes from Kondo et al. (1973)
3. Davidson (1974)	6–11.5	1.44	?	114	ec	Large buoy	Does not correct for stability effects
4. Wieringa (1974)	4.5–15	$0.62V^{0.37}$ or $0.86+0.058V$	20	126	ec	Tower	Surface tilt and wp estimates are excluded
5. Kitaigorodskii et al. (1973)	3–11	0.9 (at 3 $m\,s^{-1}$) to 1.6 (at 11 $m\,s^{-1}$)	?	29	ec	Tower	Plots C_{DN} as a function of u_*z_0/ν
6. Hicks (1972)	4–10	$0.5V^{0.5}$	25	74	ec	Tower	Accepts C_{DN} relation as same as Wu (1969b)
7. Paulson et al. (1972)	2–8	1.32	25	19	wp	Large buoy	Uses $\kappa = 0.40$
8. Sheppard et al. (1972)	2.5–16	$0.36+0.1V$	20	233	wp	Tower	Uses $\kappa = 0.40$
9. De Leonibus (1971)	4.5–14	1.14	30	78	ec	Tower	
10. Pond et al. (1971)	4–8	1.52	20	20	ec	Large buoy	
11. Brocks and Krügermeyer (1972)	3–13	$1.18+0.016V$	15	152	wp	Buoy	Data from North Sea and Baltic Sea—uses $\kappa = 0.40$
12. Hasse (1970)	3–11	1.21	20	18	ec	Buoy	See text on data interpretation
13. Miyake et al. (1970)	a. 4–9 b. 4–9	1.09 1.13	20 20	8 8	ec wp	Mast Mast	See text on data interpretation—uses $\kappa = 0.40$
14. Ruggles (1970)	2.5–10	1.6	50	276	wp	Mast	C_D anomalies found at a number of wind speeds—uses $\kappa = 0.42$
15. Hoeber (1969)	3.5–12	1.23	20	787	wp	Buoy	Data from equatorial Atlantic—uses $\kappa =$ 0.40
16. Weiler and Burling (1967)	a. 2–10.5 b. 2.5–4.5	1.31 0.90	30 75	10 6	ec wp	Mast Mast	 Uses $\kappa = 0.40$
17. Zubkovskii and Kravchenko (1967)	3–9	$0.72+0.12V$	15	43	ec	Buoy	wp estimates of u_* show low correlation with ec; possible effect of buoy motion

a. Taken from the recent literature for a reference height of 10 m: ec = eddy correlation method; wp = wind profile method. σ is the standard deviation of n data points about the mean value. [After Garratt (1977), who compiled and evaluated the source material.]

Table 17.3 Neutral Drag Coefficients over the Ocean[a]

Source	Wind speed range ($m\,s^{-1}$)	$C_{DN}(10)$ range ($\times 10^3$)	Comments
A. Miller (1964)	17-52	1.0-4.0 (linear)	Hurricanes Donna and Helene—ageostrophic
B. Hawkins and Rubsam (1968)	23-41	1.2-3.6 (discontinuous)	Hurricane Hilda—ageostrophic
C. Riehl and Malkus (1961)	15-34	2.5	Held constant to achieve angular momentum balance
D. Palmén and Riehl (1957)	5.5-26	1.1-2.1 (linear)	Composite Hurricane data—ageostrophic
E. Kunishi and Imasoto (see Kondo, 1975)	14-47.5	1.5-3.5	Wind flume experiment
F. Ching (1975)	7.5-9.5	1.5	Vorticity and mass budget at BOMEX

a. Taken from the literature, for hurricane and vorticity-mass-budget data analyses. Also included are wind flume data of Kunishi and Imasoto (see Kondo, 1975). [After Garratt (1977), who compiled and evaluated the source material.]

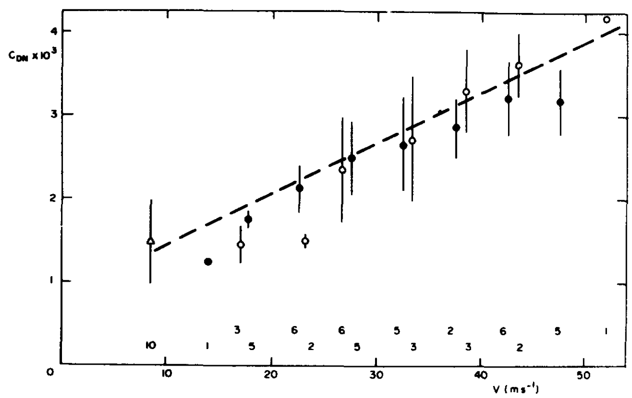

Figure 17.3 Mean values of the neutral drag coefficient as a function of wind speed at 10-m height for 5-m-s^{-1} intervals, based on individual data from hurricane studies (O), wind flume experiment (●), and vorticity mass budget analysis (△)—see table 17.3. Vertical bars as in figure 17.2. The number of data contained in each mean is shown below each mean value, and immediately above the abscissa scale. The dashed curve represents the variation of $C_{DN}(10)$ with V based on $z_0 = \alpha u_*^2/g$ with $\alpha = 0.0144$. (Garratt, 1977.)

Although our knowledge of the complicated processes in the interfacial layer is very unsatisfactory, we can, by using similarity theory and empirical knowledge of z_0, z_θ, etc., derive formulas from which the surface fluxes can be estimated from ships' observations in the near-surface layer of, say, temperature, humidity, and wind speed at a known height, together with sea-surface temperature. The errors in such estimates will be considerable, but they are more likely to be due to the errors in the ships' observations than to deficiencies in the formulas.

Calculations of the fluxes from climatological data [Jacobs (1951), Privett (1960), Budyko (1956), and more recent work by Bunker (1976) and Saunders (1977)] are of great value even though their accuracy is limited by the low precision of the ships' observations and by lack of uniformity of their cover of the ocean. They are thought unlikely to provide estimates from which the poleward heat transport by the ocean can be deduced, but will be useful in attempts to interpret the work of Oort and Vonder Haar (1976).

17.4 Waves

The most obvious effect of the wind on the sea is the generation of waves. They have been much studied, for there is no doubt of their economic importance: the design of ships, of harbors, and of sea defenses all need estimates of the waves to be encountered, to say nothing of the questions raised by the reflection of sound and light at the sea surface.

What is less obvious is how they fit into the coupled mechanics of the ocean and the atmosphere—how the winds and currents would differ if by some magic device the surface waves were eliminated. The drag coef-