PART TWO

8 The Influence of Swell on the Drag

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8.1 Introduction

Swell is formally defined as old wind sea that has been generated elsewhere. The term "old" is meant to signify that at some past time the swell energy, propagating through a given defined point, had been directly forced by the wind elsewhere. In view of the rather specific notion of "wave age" it might be better to think of swell as "escaped" wind sea. Having come from elsewhere, bearing the imprint of a different storm, swell may propagate at any speed relative to the wind or at any angle to the wind. Indeed, the vector difference in speed of the swell and peak wind sea may provide the only unambiguous criterion for identifying and separating swell from actively growing wind sea. Frequency dispersion separates the components of swell as they propagate away from the source area, and so swell tends to have a narrower spectrum than wind sea; but this provides only a qualitative selection criterion since the bandwidth of wind sea and swell may have considerable variation. For clarity we consider only two clearly defined cases of swell: (1) a distinct peak in the spectrum having peak phase speed greater than the wind component in the direction of propagation of the peak; (2) a distinct peak in the spectrum having peak phase velocity at an angle greater than 90 degrees to the wind.

The addition of a swell component to an existing wind sea may affect the drag in two ways: (a) the direct interaction of the wind and swell could enhance the drag when the swell propagates counter to the wind (case 2 above), or could reduce the drag when the swell runs ahead of the wind (case 1 above); (b) the effect of swell on altering the wind sea spectrum will also change the aerodynamic roughness of the surface. In the following sections these two effects are discussed with samples from published work and new data.

8.2 Wind–Swell Interaction

The idea that swell running ahead of the wind can return momentum to the atmosphere and produce a wave-driven wind appears to have first been promulgated by Harris (1966). Subsequent field observations by Davidson and Frank (1973) and Donelan (reported in Holland 1981 and reproduced in Fig. 8.1) provide clear evidence for an upward transfer of momentum, and Dobson (1971) reported direct measurements of the momentum transfer from a swell group over-running the wind. It remains to parametrize this in terms of a roughness length, z_0 or drag coefficient, C_D but it is likely that the effects of over-developed swell on the drag will only be significant in very light winds when the difference in phase speed and wind speed is large and the wind sea's contribution to roughness will be small.

The case of swell running against the wind has not yet been well documented, at least in connection with the modification to the wind stress. Clearly, the largest effect would be expected when wind direction and swell direction are directly opposite. On the open ocean this circumstance is unusual and, to our knowledge,

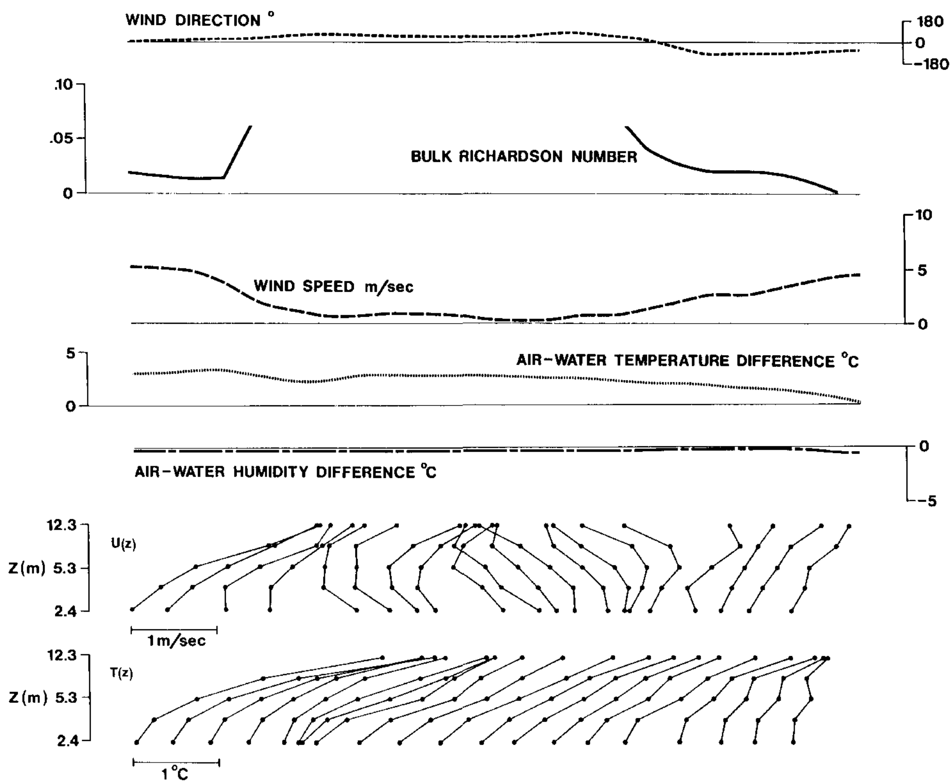

Figure 8.1. Wind and temperature profiles measured over Lake Ontario showing formation and decay of a wave-driven wind. The profiles are running averages over 30 min plotted at 10 min intervals. The humidity difference is expressed in buoyancy equivalent degrees Celsius. The time series of wind speed and air–water differences are obtained from measurements at the top level and the surface water temperature. (From Holland 1981.)

has only recently been observed with concurrent stress and wave directional measurement (Dobson et al. 1994; Donelan et al. 1997).

Lake Ontario's long axis is aligned (with the prevailing wind direction) WSW to ENE and a few times a year a low pressure centre crosses the lake from south to north bringing first east winds and then, soon after, west winds. The research platform of the Canadian National Water Research Institute (Donelan et al. 1985) is located 1.1 km from the western shore. So the east winds, blowing over 300 km of fetch, produce large waves that persist for many hours after the wind has abated or turned to the west. Under these latter conditions the east swell faces an adverse wind and produces an enhanced wind drag. Figure 8.2 shows a record of wind and wave properties under these circumstances. It is seen that the drag coefficient

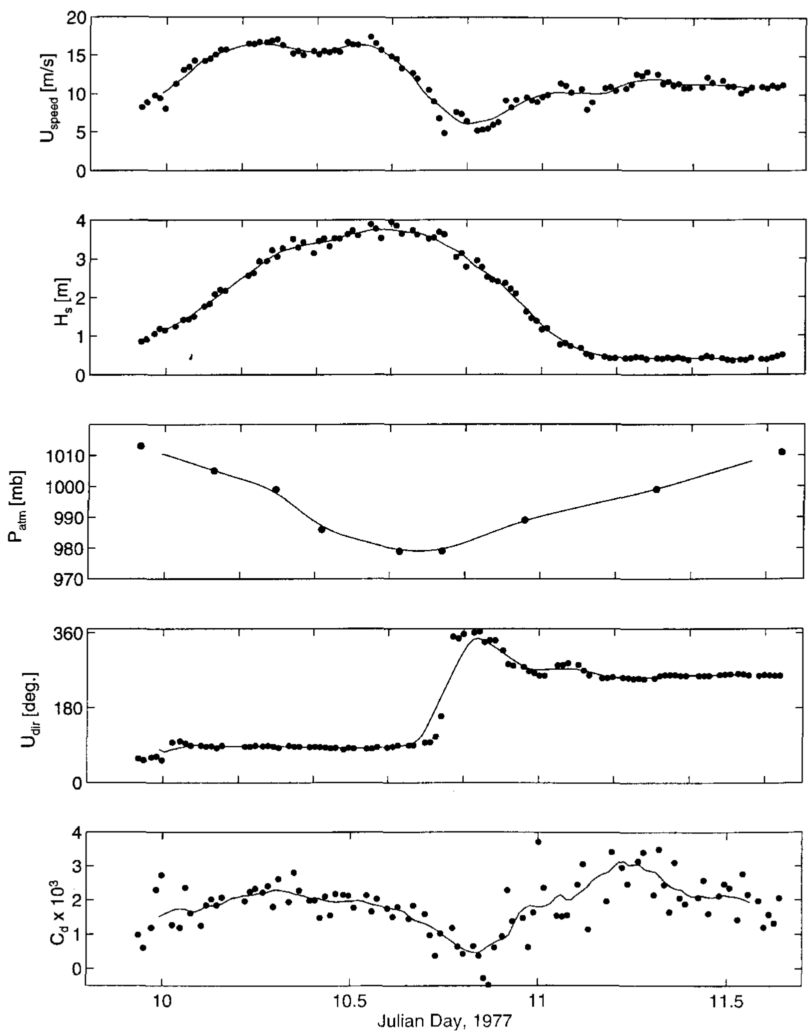

Figure 8.2. Observations of wave height, wind and the drag coefficient 1.1 km from the western shore of Lake Ontario during the passage of a low pressure centre, causing the wind to reverse direction in a few hours.

decreases as the wind speed decreases when the wind is from the east and the waves are mature, having propagated over the entire 300 km fetch of the lake. The wind abruptly turns to the north and, in blowing across the swell propagation direction, i.e. along the crests, sees very little drag. Finally, the wind turns to the west and intensifies; the rate of decay of the waves increases and the drag coefficient rises up to and above its value when the wind was from the east and 50% stronger and the waves were more than six times larger.

8.3 Swell–Wind Sea Interaction

It has often been noticed that the addition of paddle-generated swell to wind waves in a tank produces a pronounced reduction in the energy of the wind sea (Mitsuyasu 1966; Phillips and Banner 1974; Hatori et al. 1981; Bliven et al. 1986; Kusaba and Mitsuyasu 1986; Donelan 1987). Figure 8.3, reproduced from Donelan (1987), illustrates the dramatic effect on the wind sea caused by quite gentle monochromatic swell waves. The swell is seen to grow rapidly in response to the wind because these paddle waves are travelling slowly (2.2 m/s) relative to the wind (11 m/s at 26 cm height). The growth of the swell corresponds to an additional transfer or stress. However, the increased stress from direct wind–swell interaction is more than offset by the substantial reduction in the wind sea and its attendant roughness. The variance of the wind sea has been reduced by a factor of four, while the variance of the swell has been approximately doubled. In fact, the total variance of surface elevation when wind and paddle are operated together is only 67% of the sum of the two acting separately. In the open ocean, swell propagation speeds may be much closer to the wind speed and so the direct contribution to the surface stress from the swell will be small when the swell propagates in, or close to, the wind direction. The largest effect will be brought about by the attenuation of the wind sea in the presence of swell. In Fig. 8.2 the swell was propagating directly against the wind. In that case the contribution to the stress from the counter-swell would be expected to be very large and is seen to more than compensate for the reduction in the portion of the stress supported by the wind sea.

8.4 Swell Attenuation in Adverse Winds

The direct attenuation of swell in an opposing wind has never been observed in the field, and laboratory tests have had conflicting results. Experiments performed by Young and Sobey (1985) yield insignificant attenuation rates of swell as estimated from pressure-slope measurements, while those of Mizuno (1976), though somewhat scattered, show significant attenuation of counter-swell. In a series of experiments in a large (100 m) flume, in which both favourable and adverse winds were applied to swell, Donelan (1999) found that attenuation rates in adverse winds were significant although smaller than growth rates in favourable winds (Fig. 8.4). The large increase in momentum transfer described above in an opposing swell (Fig. 8.2) appears to come from drag on the swell itself since the wind sea

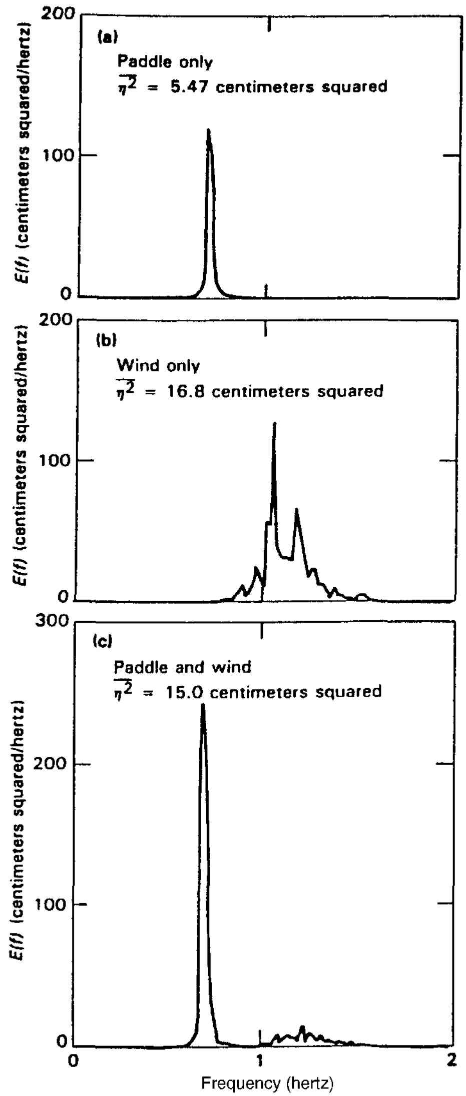

Figure 8.3. Wave spectra at 50 m fetch in a laboratory wind-wave tank. (a) The spectrum of a continuous train of 0.707 Hz paddle-generated waves of steepness $ak - 0.067$; (b) The spectrum of a pure wind sea with measured wind of 11 m/s at 26 cm height. (c) The spectrum of waves with wind and paddle excited together as in (a) and (b).

is flattened by the swell. Such a momentum transfer from the swell corresponds to its rapid attenuation.

On the other hand, observations of swell propagation, across the widest oceans and through various meteorological conditions, Snodgrass et al. (1966), seem to suggest that when the steepness of swell is reduced enough it no longer interacts directly with the wind. These very gentle long swell components may carry significant energy and on approaching the coast the energy is concentrated near the surface.

8.5 Change in the Drag due to Swell

The addition of swell to a locally wind-generated sea alters the roughness of the surface in two distinctly different ways: (1) the swell contributes directly to the

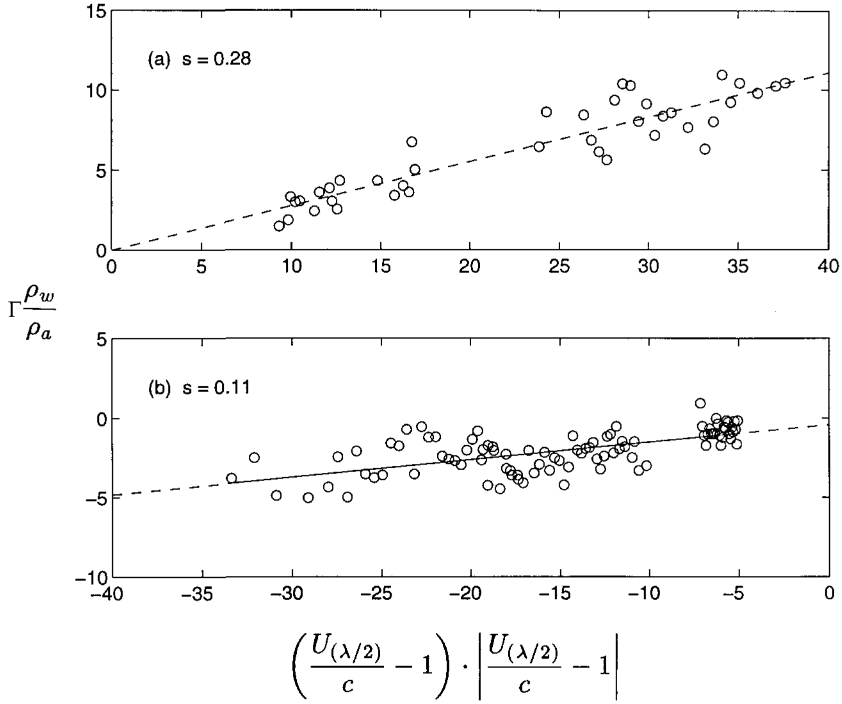

Figure 8.4. The magnitude of the fractional energy change per radian x the density ratio: (a) growth rates for the wind sea; (b) attenuation rates for the paddle-generated waves travelling against the wind. The regression lines to the data are shown and the corresponding sheltering coefficients (s = line slopes) are indicated on the figure.

surface roughness and the importance of this contribution depends sensitively on the direction of propagation of the swell relative to the local wind; (2) the swell attenuates the wind sea and, although the mechanism of attenuation is poorly understood, it may be expected to depend on the steepness of the swell and its propagation direction relative to that of the wind sea components.

Any attempt to predict the effect of swell on the drag will require detailed information on the directional properties of both wind sea and swell. Consequently, measurements at sea of the wind stress without concomitant information on the wave directional properties will exhibit considerable noise, much of which may be caused by swell. In a recent paper Yelland and Taylor (1996) have obtained a wealth of data on drag coefficients inferred from the high frequency spectrum of horizontal air velocity fluctuations via the inertial dissipation method. Their estimates of the drag coefficient, reproduced here as Fig. 8.5, show substantial scatter about some average value that is taken to depend on stability and wind speed only.

Dobson et al. (1994) made simultaneous measurements of wind stress with the inertial dissipation technique (Anderson 1993) and directional wave spectra with a pitch-roll buoy in the open ocean during the passage of several winter weather systems. Their intent was to determine a relationship between wind stress and sea state in the open sea; to do this it was necessary to partition their wave spectra into

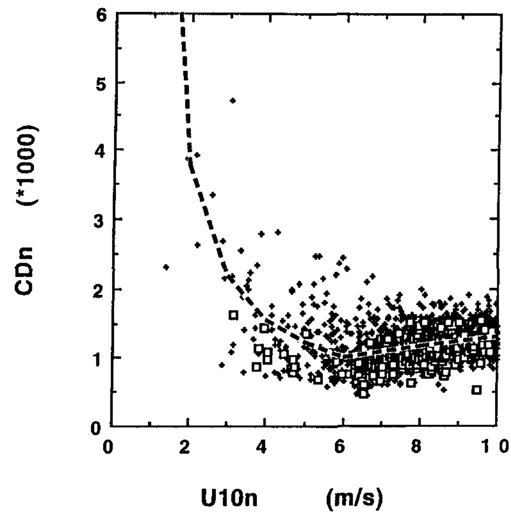

Figure 8.5. Neutral drag coefficients from two ships using the inertial dissipation method. (From Yelland and Taylor 1996.)

sea and swell, which they did using the energy and mean direction at each frequency of their buoy spectra. Wavelength and direction from the buoy spectra compared favourably with image spectra taken by ship-, air-, and space-borne radar systems.

Their findings can best be described as scattered. First, no clear wind stress versus sea state relation emerged from the wind sea parts of their wave spectra – merely a general confirmation, over a severely limited range of ages (all near $c_p/U_c = 1$, where c_p is the wave phase speed at the sea peak and U_c is the component of U_{10} in the wave direction), of the HEXOS result (Smith et al. 1992). Second, their wind stress measurements, although giving drag coefficients (Fig. 8.6a) consistent with the open-sea results of Smith (1980, 1988) and exhibiting the typical large scatter, did not stratify significantly with either the swell amplitude relative to that of the sea or with the swell direction relative to the wind (Fig. 8.6b).

They concluded that better definitions were needed of "sea" and "swell" and "wave age" in the presence of propagating and developing weather systems containing fronts. For such systems the lack of a clear understanding of the mix of physical mechanisms by which energy and momentum were transferred from the wind and from the swell into the sea severely hampered their ability to extract information from a well-calibrated, carefully made set of simultaneous wind and wave measurements.

Donelan et al. (1997) made direct observations of Reynolds stresses and wave directional properties from the SWATH ship *Frederick G. Creed* during the Surface Wave Dynamics Experiment (SWADE). Their estimates of the neutral drag coefficient are tagged with the general swell condition in Fig. 8.7. It is apparent that the presence of swell greatly increases the variability of the drag coefficient over that which obtains in a pure wind sea. Generally, when the swell is counter to the wind the drag is higher, and cross and following swell tend to produce lower drag. In Fig. 8.7 the very high drag coefficients near wind speeds of 5 m/s occurred when there was a large swell running directly against the wind. These results are in

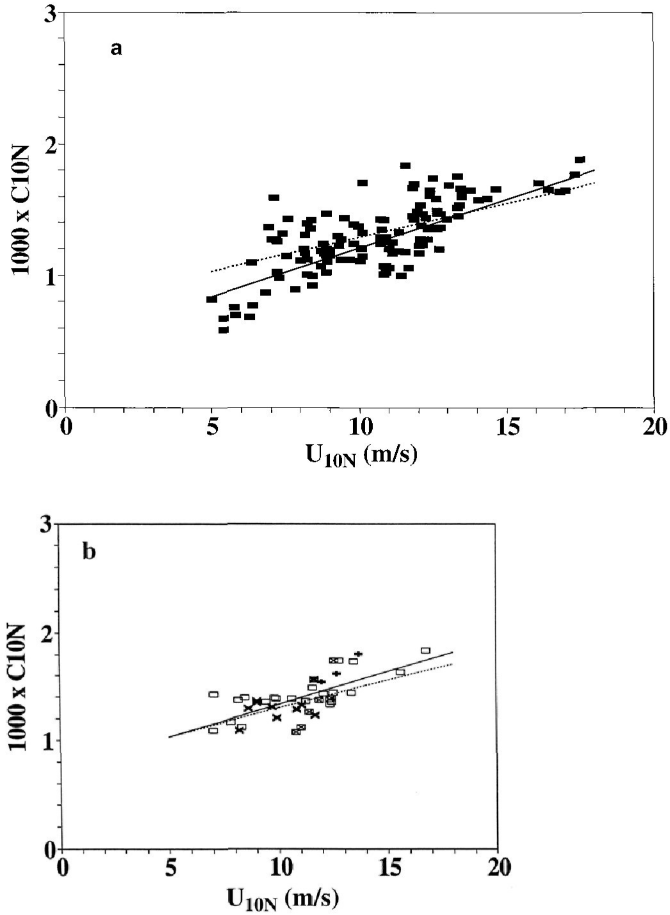

Figure 8.6. (a) Neutral 10 m drag coefficients from ship's bow anemometer runs (corrected for ship-induced distortions of the mean flow and the turbulence at the anemometer): lines are: (solid) regression to these data and (dotted) Smith (1988); (b) Neutral 10 m drag coefficients interpolated to times of wave runs: (×) mature waves ($U_c/c_p < 1.1$), (open square) middle-aged waves, (+) younger waves ($U_c/c_p > 1.7$), (box with ×) wind sea energy > swell energy. Data points deleted for cases when there was no well-defined wind sea peak or when the wind sea direction was more than 30 degrees off the wind direction. Lines are: (solid) regression on these data and (dashed) Smith (1988).

general accord with the effects of swell on the drag discussed above. As pointed out by Donelan et al. (1997), the inertial dissipation method, which was used by Yelland and Taylor (1996) and Dobson et al. (1994), responds only to the turbulent Reynolds stress through its interaction with the wind profile. In the presence of long waves (swell) some fraction of the total stress is carried by wave-coherent (non-turbulent) motions. These tend to cause a reduction on the slope of the wind profile that is more pronounced near the surface but may reach up to heights of the order of the swell wavelength. The inertial dissipation estimates, which depend on the production of kinetic energy through the interaction of the turbulent stress with the wind profile, tend to underestimate the total stress in these conditions. Some

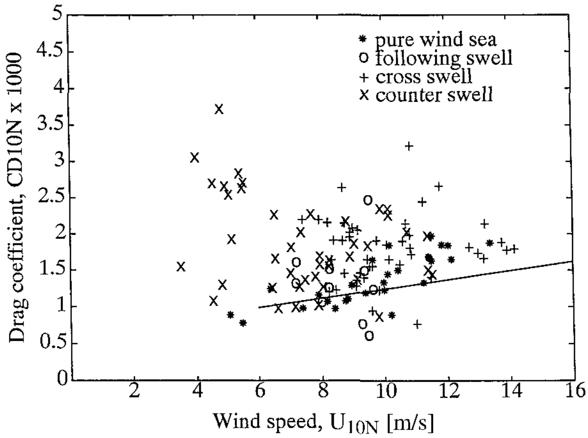

Figure 8.7. Neutral drag coefficients from the SWATH ship in SWADE from direct measurements of the Reynolds stress during various conditions of wind sea and swell. The pure wind sea relation of Smith (1980) is indicated with a solid line.

comparisons of the inertial dissipation and direct eddy correlation methods of estimating the wind stress are given in Donelan et al. (1997), but a full resolution of the matter will be realized only by the concurrent estimates of all the terms in the kinetic energy budget that are required in a rigorous determination of the stress via a balance of production with the dissipation and local divergence of kinetic energy.

8.6 Summary

When swell is relatively steep and travels with the wind, the wind waves are suppressed and the drag is lower than in swell-free cases. When swell is relatively steep and travels against the wind, the wind waves are again suppressed, but now the momentum transfer (the drag) to the swell is large enough to enhance the drag coefficient. For cross-wind swells and low-slope swells the effect on the drag coefficient appears to be small.