VARIATIONS IN THE ATMOSPHERIC DRAG COEFFICIENT
DUE TO CHANGES IN SEA STATE

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Abstract

Measurements made on both sides of an atmospheric front of the neutral atmospheric drag coefficient, \( C_{dn} \), and ocean surface wave spectra showed a large variation in \( C_{dn} \) at constant wind speed that was unexplained by conventional formulations. In our investigation, which for the first time made wave spectral measurements together with correlation flux measurements in the marine surface layer, coincident wave spectral measurements showed an unexpected increase in energy in the 6-sec wave band (.15-.17 Hz), which was well correlated with the variation of \( C_{dn} \) and the surface roughness, \( z_0 \). We derived a relation of \( z_0 \) to the wave spectra which reduces to Charnock's relation in situations of dynamic equilibrium between the wind field and the surface waves. In situations of disequilibrium, the relation shows a strong dependence of \( z_0 \) on the 6-sec wave field. This relationship of \( z_0 \) to the surface wave spectra helps explain the apparently high values of \( C_{dn} \) behind moving fronts reported in the literature.

1. Introduction

Recent work suggests that the long-standing disagreement between micrometeorologists and other users of drag coefficient on the nature of the neutral drag coefficient, \( C_{dn} \), may be a result of the preference of micrometeorologists for taking measurements during conditions of steady wind.

Micrometeorologists, in general, have concluded, through wind profile observations and direct measurements of Reynolds stresses, that the drag coefficient is a slowly increasing function of wind speed and that it is a factor of two or three smaller than the values of \( C_{dn} \) used to bring the computed results of numerical models into close agreement with observations. Most determinations of \( C_{dn} \) made by micrometeorologists are done under conditions of horizontal homogeneity and steady state and, therefore, their over-the-oceans measurements appear biased towards a fully developed wave field, where, it has been shown, the drag coefficient approaches a pronounced minimum (Kitaigorodskii, 1970; Krauss, 1967). Models, on the other hand, are usually verified using measurements made during "significant events" that are usually associated with rapid changes in the wind and, correspondingly, with a young wave field, for example, the passage of a storm over the Great Lakes.

This investigation uses measurements of the drag coefficient made in regions where there were time and spatial variations of both the surface wind and sea state. Our goal was to provide a more realistic representation of the drag coefficient in the turning-wind condition, which would allow greater accuracy in calculation of the momentum exchange and heat and water vapor fluxes during a storm passage over the ocean.

The effort involved measuring, by the correlation method, the Reynolds stress in the marine boundary layer close to a moving atmospheric front. The Reynolds stress measurement was correlated with the bulk surface-layer meteorological measurements together with thermal stability in order to characterize the change in the neutral drag coefficient as the surface wind changes direction abruptly over an established wave field.

Parametric Description of \( C_{dn} \)

In our investigation, we can represent the momentum flux as:

\[
\tau = -pu'w' = pu_x^2
\]

where \( u' \) and \( w' \) are the longitudinal and vertical components of wind velocity fluctuations, \( p \) is the air density, and \( u_x \) is the dynamic velocity. \( \delta U \) characterizes the total flux of momentum intersecting the underlying surface in almost the entire thickness of the boundary layer.

The most frequently sought formulation of the small-scale interaction between the atmospheric surface layer and the underlying surface reduces to a determination of these fluxes by means of an external, readily measured variable, such as:

\[
\delta U = U_a - U_w \approx U_a
\]

where subscript \( a \) represents standard height.
measurement in the atmosphere—for example, 10 meters, and \( v \) correspond to the surface measurement, i.e., the speed of the water.

From this quantity, together with density, \( \rho \), we can form a combination that has the dimension of momentum flux \( \rho (\bar{u})^2 \). The ratio of the true turbulent flux \( \bar{v} \) to the easily measurable combination is the drag coefficient, \( C_d \)

\[
C_d = \frac{\bar{v}}{\rho \bar{u}^2}
\]

For any selection of measurement height in the layer \( z \), the largest contribution to \( \bar{v} \) is made by the lowest part of the layer. \( C_d \) will depend on the characteristics of the turbulent regime in the immediate vicinity of the underlying surface. Therefore, in order to discuss the variability of \( C_d \), we need to know the laws governing vertical turbulent exchange in the main thickness of the surface layer, \( z = z_s \), and the hydrodynamic properties of the underlying surface, the ocean surface.

The complete group of dimensionless parameters governing \( C_d \) can be expressed as:

\[
C_d = C_d \left( \frac{z}{h_s}, \frac{L}{h_s}, \frac{\delta_v}{\delta_n} \right)
\]

where \( L \), is the Monin-Obukhov length scale, \( h_s \) is the scale of surface roughness, \( \delta_v \) is the thickness of the viscous sublayer, and \( z \) is the measurement altitude.

The ratio \( z \) is related to the Richardson number

\[
R_i = \frac{g \delta_v}{\theta_v (du/dz)^2}
\]

where \( \theta_v \) is the virtual potential temperature, which can be used to characterize the conditions of stability in the temperature-stratified surface layer.

We can then write

\[
C_d = C_d \left( z, R_i, Re_s \right)
\]

where \( Re_s \) is the roughness Reynolds number.

The roughness regime breaks down into three areas. The first, aerodynamically smooth flow is scaled by \( \delta_v \), which is determined by parameters \( v \) and \( u_* \) such that

\[
\delta_v = v/u_*
\]

The expression for \( Re_s \) then is

\[
Re_s = \frac{h_s u_*}{v}
\]

For aerodynamically rough flow when the underlying surface is homogeneous and reasonably steep but has vertical dimensions still smaller than the depth of the sublayer, i.e., \( h_s \ll L \), then for large values of \( Re_s \) the main resistance to the flow will come from the normal pressure on the roughness elements. In this case, the drag coefficient has to be independent of the air viscosity, \( v \), and therefore

\[
f(Re_s) \sim \text{constant}
\]

For an incompletely rough surface, there must be a transition zone between the two cases. Numerous studies have been made in an effort to determine the form of the function \( f(Re_s) \) for a range of \( Re_s \). Nikuradse' classic studies (1933) are often quoted as a basis for development of the limiting-drag regimes for immobile boundaries. An adequate representation of aerodynamic roughness has been sought extensively in the oceanographic literature (Kraus, 1966; Kondo, 1975; Sethuramen, 1978).

Most studies of smooth flow or aerodynamically incomplete roughness use wave tanks or short-fetch limited regimes. Our investigation will deal only with field data and completely rough regimes.

Therefore, we can treat \( Re_s \) as a constant and

\[
C_d = C_d \left( z, R_i, Re_s \right)
\]

We use the work of Businger et al. (1971) as a basis for calculating the effect of stability on the drag coefficient and, by compensation for stability, effectively calculate the drag coefficient for a neutral atmosphere, \( C_{dn} \). Then

\[
C_{dn} = C_d \left[ 1 + \left( \frac{k}{Re_s} \right)^{1/2} \right]
\]

where \( k \) is the von karmen constant, \( \psi_n \) is the diabatic correction as shown in Businger et al. (1971) and evaluated at \( z = 10 \) meters. Now,

\[
C_{dn} = C_d \left( z \right)
\]

Roughness length and relation to \( C_{dn} \)

Utilizing the formulas of logarithmic boundary layer theory for the neutral case, we have

\[
du(z) = u_* \frac{dz}{k} \]
where \( z \) is the integration constant. Remembering the equation
\[
C_{dn} = \frac{1}{k} \ln \left( \frac{z}{z_0} \right)
\]
we have
\[
C_{dn} = \frac{1}{k} \frac{1}{z_0} (\ln z)^2
\]

\( C_{dn} \) will depend only weakly on \( z \). The only unknown, therefore, is \( z_0 \). However, \( z_0 \) is difficult to measure on the ocean surface, and usually is inferred from atmospheric wind profile measurements when they are available, assumed a constant, or derived from a model. As a result, the underlying sea state has been neglected in previous work on determining a momentum drag coefficient necessary for relating bulk measurements to momentum flux.

Garrett (1977) reviewed all available data for calculating drag coefficients, terrestrial and marine, and concluded that \( C_d \) is a slightly increasing function of wind speed. Smith and Banke (1975), using additional data ranging from a gentle breeze (7 m/sec) to strong-gale force winds (21 m/sec), found a significant increase in the drag coefficient with wind speed which was well described by the equation of Charnock (1955). Smith and Banke also reviewed some earlier results and concluded that

\[
10^3 C_d = 0.63 + 0.066 U_{10} \pm 0.23
\]

with \( U_{10} \) in m-sec\(^{-1} \) which describes all of the results well. They noted that

\[
C_{dn} = k^2 \left[ \ln \frac{z}{z_0} \right]^2
\]

fits the data as well if

\[
z_0 = \frac{a u_w^2}{g}
\]

with \( k = 0.4 \) and \( a = 1.44 \times 10^{-2} \), which is close to Charnock's original suggestion of \( a = 1.23 \times 10^{-2} \).

Several investigators have computed the drag coefficient from profile measurements and found substantial variations in \( C_d \) which were virtually uncorrelated with the wind speed. Ruggles (1970) used 299 mean wind profiles to find the drag coefficient as a function of wind speed. He termed the region of large variabilities in \( C_d \) discontinuities, but noted that failure to consider the momentum partition of energy going into wave development left unanswered the question whether the values of \( C_d \) are characteristic of a rising, falling, or fully arisen sea. He speculated that the nodes or discontinuities may be a result of differences in physical states of the ocean surface, such as wave-generating conditions as contrasted with non-generating conditions.

Kitaigorodskii (1970) found that \( C_d \) was generally a constant function of wind speed \( U_{10} \) to approximately 7 m/sec, where it then showed a tendency to increase. He noted, however, that the concept of critical wind speed for mean values of \( C_d \) is rather arbitrary, because the differences in mean values of \( C_d \), even for large changes in wind speed can be smaller than the differences in values of \( C_d \) at different stages of wave development, but at the same wind speed. Since he had shown that there is an absence of a single-valued relationship between \( z \), \( u_w \), and \( C_d \), which may be due to the variability of the wave characteristics that affect the drag, Kitaigorodskii suggested grouping \( C_d \) measurements according to values of \( c/u_w \) where \( c \) is the phase velocity of the wave with the frequency of the peak of the wave spectrum. He showed that the drag coefficient as a function of \( c/u_w \) tends to reach a minimum as the wave field ages. His data were limited to wind speed between 4 and 15 m/sec.

Kitaigorodskii has also shown the absence of a single-valued relationship between \( z \) and \( u_w \), which may be due to the variability of the wave characteristics that affect the drag. These characteristics in turn are not in single-valued relationship with the local wind, but depend rather on the history of the wind speed and direction and the fetch available for development of the wave field.

The hypothesis of this study is that sea state has an effect on roughness length and \( C_d \), that cannot be accounted for as merely a function of wind speed. We include more comprehensively than previously the surface wave characteristics to properly account for the scatter and variation of \( C_d \). We also also represent the response of the wave field to the atmospheric input by analyzing the complete wave spectrum rather than only \( H_{1/3} \) and the peak frequency.

\( C_d \) can be expressed as a function of only \( z_0 \). Therefore, it would suffice to know a single parameter \( z_0 \). The basic parameter \( z_0 \) would assume an elementary form that returns the
basic parameterization derived earlier. We have then

\[ z_o \propto h_s \]

where \( h_s \) is a characteristic scale of the surface roughness elements.

Previous studies of marine drag coefficient have either neglected the wave field completely (Large and Pond, 1981), or used measurements of only the height of the dominant waves (Hsu, 1979), which severely limits the capability to model \( z_o \) because the effective density of physical elements of height scale \( h_s \) cannot be represented. Similarly, using the height of only the significant wave to calculate marine boundary layer values of \( z_o \) neglects the additional possibility that waves which are shorter than those of the peak frequency could influence the effective scale of \( z_o \).

Examining the relationship of the complete spectral energy distribution of the wave field to the calculation of \( z_o \) enabled us to account for the contribution of waves that are not at the spectral peak to the surface roughness felt by the marine boundary layer.

2. Experiment Description

The Storm Transfer Response Experiment (STREX) (Fleagle et al., 1982) provided an opportunity to calculate the drag coefficient from aircraft gust-probe measurements coincidentally with wave spectral measurements and measurements of atmospheric stability. These measurements allowed the calculation of the magnitude of the drag coefficient and a parameterization of the effect of sea state on the drag coefficient.

Wind speed analysis was derived from aircraft level measurements and vertical camera photography. An assessment of the coverage of surface foam and white caps provided an estimate of surface wind speeds. Figure 1 shows the wind speed estimate from the two methods plotted as a function of increasing distance from the front. It is noteworthy that the foam coverage surface winds drop below the phase speed of 6-sec waves at the distance where both the drag coefficients and the 6-sec wave energy drop.

Figure 2 is a plot of the drag coefficient \( C_d \) and the stress, \( \tau \), as a function of distance from the front. The surprising aspect of this figure is the dip in stress and \( C_d \) immediately to the rear of the front before the stress and \( C_d \) increased as the distance behind the front increased. When the frontal distance had exceeded 200 km, \( C_d \) and the stress began to return to lower values.

The unique data set for the 15 November frontal passage allows us to use the measured eddy correlation momentum flux and the bulk wind speed \( U_{10} \) to recover a \( C_d \) that is independent of modeling errors.

3. Wave Measurements

Comparing two of the wave spectra before we proceed, we find that the surface wave spectra measured ahead of the front in the warm sector of the storm, as shown in Figure 3 has the typical shape as described in the JONSWAP report (Hasselmann et al., 1973) a very steep front face and rapid drop behind the spectral peak which occurs at .10 Hz. Note especially that the values for Phillips' parameter reach

Figure 1. Surface wind speed vs frontal distance. Upper panel wind speeds were derived from aircraft level winds and reduced to \( U_{10} \) using stability correction. Lower panel wind speeds were derived from surface foam coverage.

Figure 2. The drag coefficient \( C_d \) shown by the open circles and the stress \( \tau \) shown by for the upwind frontal penetration into the cold sector on 15 November 1980.
equilibrium at the peak frequency, which implies that this spectrum behaves as Phillips' model predicts; that is,

$$S(\omega) = \beta g^2 \omega^{-5}$$

where $S(\omega)$ is the spectral density, $\beta$ is Phillips' constant, and $g$ is the local gravity. These spectra are very likely in dynamic equilibrium with the warm sector wind field. One of the spectra from the cold side of the front is shown in Figure 4 and exhibits quite a different spectral shape. The spectral peak here is less sharp and it drops less rapidly on the rear face. Moreover, Phillips' parameter does not come to equilibrium until well beyond the region of maximum spectral energy which implies that this wave field does not fit Phillips' model.

Figure 5 shows the values of $\beta$ and energy density in the 6-sec wave field, $E_6$, for the wave spectra measured on the westward passage through the front. It can be seen that the energy in the saturated part of the spectrum, characterized by the value of $\beta$, at first increases rapidly immediately behind the front, and then decreases towards an equilibrium value as the measurements proceed farther into the cold sector. The energy density in the 6-second waves is rather low immediately behind the front but increases systematically to a maximum approximately 200 km behind the front. It then decreases at larger distances behind the front.

The region of the wave spectrum that is of most interest is the region where the phase speed of the waves is less than the surface wind speed. We can see from Figures 2 and 5 that the energy in the 6-second wave field, which are the largest waves moving slower than the wind, correlates well with the rapid increase in $C_d$. Closer to the front, before the 6-second wave field has a chance to develop, the drag is correlated to the even shorter, therefore, slower waves in the saturation region of the spectrum which are characterized by Phillips' parameter $\beta$.

4. Discussion

A large body of data points to a representation of $z_0$ as occurring according to Charnock's original equation, i.e.,

$$z_0 \propto \frac{a v_n^2}{g}$$

and, therefore, statistically $C_d$ is a slowly increasing function of wind speed.

It appears that there is a dynamic equilibrium situation existing between waves and the overlying wind field. This dynamic equilibrium is characterized by a wave spectral peak frequency whose phase speed is less than or very close to that of $U_{10}$. The JONSWAP spectra and fetch-limited spectra, in general, seem to

![Figure 3. Surface wave spectra from the warm side of the front.](image)

![Figure 4. Surface wave spectra on the cold side of the front.](image)

![Figure 5. Energy density in the 6-second wave field.](image)
follow this equilibrium. It is with this type of
dynamic equilibrium that a representation of \( z \)
in terms of the sea state should converge to
Charnock's dimensional formulation, and we need
a representation of \( z \) that reflects this.

Kurtzback (1961) performed a pilot experi-
ment on wind profile modification on the ice of
Lake Mendota, Wisconsin. He varied \( z \) from
10\(^{-4}\) m to 10\(^{-1}\) m. In a simplification of
Kurtzback's results Lettau (1969) proposed the
following representation for \( z \):

\[
\begin{aligned}
z_0 &= 0.5 \, h^* \, s/S \\
\end{aligned}
\]

where

\[
\begin{aligned}
h^* &\text{ is the height of the object meters} \\
s &\text{ is its silhouette area meters}\^2 \\
S &\text{ is the specific area meters}\^2 \\
\end{aligned}
\]

This type of representation allows for an
additional boundary condition equivalent to
packing density. In a stationary terrestrial
situation, \( z \) can be computed straightforwardly
for any given situation.

It is somewhat different, however, with a
mobile underlying surface. We use Lettau's idea
as a starting point to develop a representation
of \( z \) that responds to the underlying wave
field. 0

Suppose we write

\[
\begin{aligned}
z_0 &= f_w \, h^* \, s/S \\
\end{aligned}
\]

where \( f_w \) is some "average" drag coefficient and
\( h^* \) is the wave height.

For a two-dimensional wave field moving in
the direction of the prevailing wind, we can show
that

\[
\begin{aligned}
s &= h^* \times \text{width} \\
S &= \lambda \times \text{width} \\
\end{aligned}
\]

where \( \lambda \) is the wavelength of the waves.

Therefore, the \( s/S \) ratio reduces to the
slope of the wave, \( h^*/\lambda \), for each individual
roughness element as it is seen by the wind. Now

\[
\begin{aligned}
z_0 &= f_w \, h^* \frac{h^*}{\lambda} \\
\end{aligned}
\]

We know from deep-water wave theory that

\[
\begin{aligned}
c_p &= \frac{\sqrt{gk}}{2\pi} \\
\end{aligned}
\]

where \( c_p \) is the phase velocity of the wave with
wavelength \( \lambda \). But

\[
\begin{aligned}
c_p^2 &= \frac{g^2}{w^*} \\
\end{aligned}
\]

or substituting

\[
\begin{aligned}
\lambda &= \frac{2\pi c_p^2}{g} = \frac{2\pi}{w^*} \\
\end{aligned}
\]

therefore

\[
\begin{aligned}
z_0 &= f_w \, h^* \frac{h^*}{\lambda} \frac{w^2}{2\pi g} \\
\end{aligned}
\]

\( z_0 \) is now expressed as a function of \( h^* \)
and \( w^* \) and an undefined drag coefficient. But
\( h^* = h^*(w) \). We know that:

\[
\begin{aligned}
h^* &= a^2 \frac{w^2}{2\pi g} \\
\end{aligned}
\]

where \( a^2 \) is the amplitude squared of the wave.
We also know that the rms amplitude squared, \( a^2 \),
can be written as,

\[
\begin{aligned}
a^2 &= \int S(w) \, dw \\
\end{aligned}
\]

where \( S(w) \) is the spectral energy density of the
surface waves.

Now

\[
\begin{aligned}
z_0 &= f_w \, a^2 \frac{\int S(w)a^2 \, dw}{2\pi g} \\
\end{aligned}
\]

In the situation of dynamic equilibrium, we can
expect the wave spectrum to behave according to
Phillips' model

\[
\begin{aligned}
S(w) &= \beta \, g^2 \, w^{-5} \quad w > w_{\text{max}} \\
S(w) &= 0 \quad w < w_{\text{max}} \\
\end{aligned}
\]

and \( w_{\text{max}} \) is the peak of the wave spectrum. We
also expect the peak frequency of the spectrum
to be very close to \( \omega = g/U_{10} \). We found in the
warm sector on 15 November that this was the
case. In this situation, the entire energy of
the spectrum is characterized by \( \beta \) and the peak
frequency \( \omega \), and we can write

\[
\begin{aligned}
z_0 &= f_w \, \beta \frac{g^2}{2\pi g} \int w^{-3} \, dw \\
&= f_w \, \frac{g}{g/U_{10}} \\
\end{aligned}
\]

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We want to use only the part of the spectrum that moves slower than the wind and let \( \omega \) be the phase velocity of the surface waves moving slower than the wind. Since \( \frac{\omega}{u} \leq 20 \) for actively developing waves (Kitaigorodskii, 1973), we could also use \( \omega = \frac{g}{c_0} \) as the limit in the integral. Evaluating the integral we have

\[
\frac{z_0}{a} = \frac{f_\omega b_\omega u_0^2}{4ng}
\]

If we had taken the limit to be \( \frac{g}{20u} \), the functional form of \( z_0 \) would be

\[
\frac{z_0}{a} = \frac{200 f_\omega b_\omega u_0^2}{2ng} = \frac{a' u_0^2}{g}
\]

which is the original relation derived by Charnock on dimensional grounds.

There is always some point in the spectrum at high enough frequency that the remaining spectrum could be modeled by Phillips' relation \( S(\omega) = \beta g^2 \omega^{-5} \). Usually when \( \omega \geq 1.88 \).

If, however, the wave field is not in dynamic equilibrium and for some reason cannot be modeled by Phillips' spectra, then the evaluation of \( z_0 \) would have to proceed differently. For these cases where the spectrum cannot be modeled analytically, the representation for \( z_0 \) could be written as,

\[
\frac{z_0}{a} = \frac{f_\omega}{2ng} \left[ \frac{\omega = 1.88 \sum S(\omega) \omega^2 d\omega + \beta g^2}{\int \omega^{-3} d\omega} \right] \omega = 1.88
\]

where \( \omega \) is the frequency of the lowest frequency wave that has a phase velocity less than \( U_{10} \). We can see that we will still have a contribution that will scale as

\[
\frac{\beta u_0^2}{2ng} \text{ or } \frac{\beta u \omega^2}{2ng}
\]

from the second term in the brackets. But the additional summation term will weight the roughness elements by the inverse square of the wave speed

\[
\omega^2 \propto \frac{1}{c^2}
\]

As \( \omega \) increases from the \( \omega_1 \) toward \( \omega = 1.0 \), the relative weight of \( S(\omega) \) in the non-equilibrium case is not decreasing as fast as \( \omega^{-5} \); therefore, the contribution from the first term in the square bracket will always be larger than the equivalent integral calculation, using Phillips' representation. The result is that the representation for \( z_0 \) will always be larger than if the wave field had the dynamic equilibrium distribution of roughness elements. It is interesting to note that the area where the \( \omega^2 \) weighting begins to exceed 1 is the beginning of the six-second wave band \( \sim 0.159 \) Hz. Therefore, the increase in energy in this band should correlate well with an increase in \( z_0 \) and \( C_{dn} \). As we have seen, the energy in the 6-second wave band does correlate well with the variation in \( C_{dn} \) when the 6-second energy is above a certain threshold.

5. Summary and Conclusions

The experimental data showed that \( C_{dn} \) varies over a large range and that variations exceeded any that could be accounted for by changes in surface wind speed using relationships reported previously in the literature. The wind speed for most of the measurements made in the unstable sector was nearly constant (to within \( \pm 1 \) m/sec). Previous work, which attempted to relate drag coefficients to surface wind speed, paid slight attention to the coincident surface wave field. As a result, no explicit dependence of \( C_{dn} \) on the wave field has ever been reported. Our investigation, which measured the complete one-dimensional surface wave spectrum, constitutes the first time wave spectral measurements have been made together with eddy correlation flux measurements in the marine surface layer.

The data sets used during the study are unique in that it measures momentum flux, bulk wind speed and temperature, and surface wave spectra. This measurement group includes all of the variables needed to check the relations of \( C_{dn} \) to \( z_0 \).

The wave spectral energy distribution in the duration-limited sea region behind the 15 November front was found to differ markedly from the energy distribution of a fetch-limited case, while the short gravity waves continued to increase in energy with increasing distance behind the front.

Variations in the drag coefficient, \( C_{dn} \), measured in the cold sector appeared to be well correlated with the increase in energy in these short gravity waves.

Using only the wave spectral information, we derived a representation of \( z_0 \), which reduced to Charnock's relation in situations of dynamic equilibrium between the wind field and the surface waves. In situations of disequilibrium between the wave field and the surface wind, the derived relation shows a strong dependence of \( z_0 \) on the 6-second wave field. We found that the \( C_{dn} \) calculated from measurements...
made during the 15 November 1980 frontal penetration followed the derived z_o relationship closely. This relationship of z_o to the surface wave spectra helps explain the apparently high values of C_{dm} behind moving fronts that have been reported in the literature.

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