The turbulence regime of a very stable marine airflow with quasi-frictional decoupling

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Abstract. Data from an air/sea interaction experiment over the Baltic Sea during conditions with very large land - sea surface temperature contrast have been analyzed. It is found that the measured shearing stress and sensible heat flux were strongly suppressed compared to what would be expected from current theory. Analysis of spectral behavior shows that a strong wind gradient imposed by the presence of a low-level jet at or below 100 m height during about 80% of the time is exerting strong influence on the turbulence regime by suppressing low-frequency fluctuations. The dissipation length scale was found to equal a new length scale \( L_{or} \) which has the vertical velocity standard deviation in the numerator and the sum of the wind gradient and the Brunt-Väisälä frequency in the denominator. It was also found that the 10 m drag coefficient \( C_{m0} \) was a strong function of this scale, with small values of \( L_{or} \) giving almost zero \( C_{m0} \) values. Integrating the locally found wind gradient down to the surface resulted in roughness length values several orders of magnitude less than that valid for a dynamically smooth flow. With turbulence measurements at only one height, 10 m, no conclusive explanation of this anomaly could be given. Observed significant wave height during the experiment was, however, found to be a factor of 2.5 lower on the average than predicted by current theory.

1. Introduction

BALTEX (Baltic Sea Experiment) is one out of five GEWEX (Global Energy and Water Experiment) regional experiments. As is evident from its name, it is devoted to the study of the energy and water balance of the Baltic Sea region, more precisely the water shed area of the Baltic Sea [GKSS-Research Center, 1995]. It is unique among the GEWEX experiments in that it includes an oceanic component. Thus the study of the water and energy balance of the Baltic Sea itself is a major topic within BALTEX. With this background in mind, it is highly relevant to note that 5 years' worth of temperature profile data taken on a 145 m high tower situated close to the coast on the island of Gotland fight in the middle of the Baltic Sea proper (site B, Näsudden, Figure 1) show that more or less stable stratification occurs during as much as 2/3 of all time; see Smedman et al. [1997] for details. This renders the study of situations with stable stratification particularly relevant.

Previous studies of situations with stable stratification over the Baltic Sea, notably those by Smedman et al. [1995, 1997], indicate that situations which exhibit considerable deviations from conventional Monin-Obukhov theory often prevail during stable conditions. No doubt this has important consequences for the parameterization of turbulent fluxes in such conditions in numerical models of all scales ranging from general circulation models to mesoscale models. As will be evident from the present paper, the problem of very stable flow contains even more facets than those reviled in the previous papers referred to above.

The situation which is conductive to formation of stable boundary layers over the Baltic Sea is that with warm land surfaces surrounding a much colder sea, which is typical for springtime and summertime conditions in the area. Similar situations are found in other areas in high latitudes with large water bodies surrounded by land on all sides. One of the first studies related to the formation of stable internal boundary layer over cold water is that of Csanady [1974], who studied the surface stress over Lake Ontario. Additional references to previous studies are given by Smedman et al. [1997]. Their paper is an attempt to systematize observations of the development of stable internal boundary layers from several field experiments in the Baltic Sea area and from numerical simulations. The conclusion is that during conditions with not too large temperature difference between land and sea a two-stage process is operative: first the layer immediately above the surface becomes stable, but after a certain transport time a mixed layer capped by an inversion develops. These two subregimes differ very much in terms of the stress on the underlying water surface.

In the case of a land/sea surface temperature difference larger than about 12°C a different regime occurs: the degree of turbulence remains so low in the boundary layer over the cold water surface that a mixed layer never develops, irrespective of travel distance. Smedman et al. [1997] coined the term "quasi-frictional decoupling", in order to distinguish this flow regime from cases with true frictional decoupling, which sometimes occur in situations with decaying surface waves [Volkov, 1970; Makova, 1975; Chambers and Antonia, 1981; Smedman et al., 1994]. The present paper is a study devoted to the turbulence regime of such a situation.

The analysis of the present paper is based on intensive measurements carried out in the area east of the island of Gotland (Figure 1, site A) during the time period May 29 to June 15, 1995. The conditions during most of this experiment were such that "quasi-frictional decoupling" applies. The synoptic situation was rather persistent, with a high-pressure ridge extending from Russia toward the northeast and a weak low-pressure area to the southwest. Thus the general flow over the Baltic Sea was from the south-east or from the south, with a typical speed of 7 - 8 m s⁻¹ at
2. Sites and Measurements

The main measuring site is the island Östergarnsholm, situated about 4 km east of the big island of Gotland, Figure 1. Östergarnsholm is a low island with no trees. A 30 m tower has been erected at the southernmost tip of the island. The base of the tower is situated at about 1 m above mean sea level. The distance from the tower to the shore line in calm conditions is only a few tens of meters in the sector from northeast over south to southwest. In highwave conditions there is an impression of the tower standing in the water. The seafloor immediately outside the peninsula has an approximate slope of 1:7, which is optimal in terms of presenting minimum disturbance to incoming surface waves (K. Kahma, Finnish Marine Research Inst.; personal communication, 1996). At about 6 km from the peninsula the depth is 50 m, reaching below 100 m further out. The approximate sector 60°-220° is characterized by more than 150 km undisturbed upwind over water fetch. Thus the surface wave field for winds from this sector is governed mainly by wind forcing and wave age.

During the 1995 campaign the 30 m tower was instrumented with slow response ("profile") sensors of in-house design for temperature [Högström, 1988] and for wind speed and direction [Lundin et al., 1990] at the following heights above the tower base: 7, 11.5, 14, 20, and 28 m. In addition, humidity was measured at 6 m above the tower base. Turbulent fluctuations were recorded with a Institute of Meteorology, Uppsala University turbulence instrument at 9 m. This instrument is basically a three-axial hot-film instrument mounted on a wind vane, with additional sensors for temperature and humidity (wet bulb sensor). The instrument has been carefully tested in a big wind tunnel [Högström, 1982 and Bergström and Högström, 1987] and in the field [Högström, 1988 and 1990]. "Profile data" were recorded at 1 Hz, and turbulence data were recorded at 20 Hz.

Wave height and direction as well as water temperature was measured with a directional wavemeter buoy (run and owned by the Finnish Marine Research Institute) moored 5 km south-southeast of the tower at a site with a water depth of about 40 m. The buoy relayed data to a receiver station at the Gotland coast some 5 km west of the buoy.

The measurements during the intensive measuring field campaign also included numerous radiosoundings, pilot balloon measurements, and airborne measurements, which taken together gave a detailed picture of the flow regime. An account of these data is given by U. Högström et al., (A case study of two-dimensional stratified turbulence, submitted to Journal of Atmospheric Sciences, 1997) (hereinafter referred to as Högström et al., submitted manuscript, 1997).

3. General Characteristics of the Flow Situation

The most characteristic feature of the meteorological regime studied here is the very strong stable stratification in the lowest few hundred meters in general and the layer next to the sea sur-
face in particular (Figure 2). In spite of the long over water transport distance, more than 150 km, there is thus no sign of a transition to a mixed layer regime, as discussed by Smedman et al. [1997].

Another characteristic feature of this particular flow regime is the frequent occurrence of low-level jets at 100 m above the water surface or less (Figure 3). In fact, such low-level jets, with more or less pronounced wind maxima, were present in 43 out of a total of 51 pilot balloon wind profiles. Although most of these are daytime measurements, the phenomenon is also present in the few night profiles that were actually taken. The mechanism for the jet was identified as an analogy in space to the well-known nocturnal jet over land, as discussed in detail by, for example, Smedman et al. [1995]. Thus an inertial oscillation is initiated as a result of the sudden reduction of friction that occurs when air flows from heated, rough land over the cold, smooth sea. Because the mechanism of the jet is frictional decoupling, it may at first seem strange that this jet occurs at all times of the day. However, as clearly shown by measurements as well as by simulations, the thermal contrast between land and water was strong enough during the nights that the phenomenon was effective also then.

As shown in Högström et al. (submitted manuscript, 1997), the originally three-dimensional turbulence encountered over the upwind land area is gradually transformed into two-dimensional turbulence as the air becomes stabilized. This stratified turbulence undergoes spectral upscaling, which results in a characteristic mesoscale spectrum with a well-defined -5/3 range for spectral scales from kilometers to hundreds of kilometers, as predicted by Lilly [1983]. In the recordings of temperature and wind on the 30 m tower, this stratified two-dimensional mesoscale turbulence results in the kind of fluctuations exemplified in Figures 4 and 5. Corresponding traces of wind direction (not shown here) show the same rapid time variation but almost no height variation.

![Figure 2](image1.png)  ![Figure 3](image2.png)

**Figure 2.** Vertical profile of potential temperature (θ) measured on May 31, 1995, at 0108 local standard time.

**Figure 3.** Vertical profiles of wind speed and direction (WD), measured on May 31, 1995, at 0618 local standard time. The different symbols are data from three different balloons, released in succession during about 20 min.
To summarize, the mean flow environment for the boundary layer studied here is a flow of high static stability, the Brunt-Vaisala frequency, $N = (g/\rho_0) 1/z^2$ taking typically values in the range 0.03 to 0.07 s$^{-1}$. The vertical wind gradient is also large, of the order of magnitude 0.05 to 0.1 s$^{-1}$. Another general characteristic is rapid variations with time in the mean flow field, caused by the stratified quasi-two-dimensional turbulence, discussed above.

4. Bulk Coefficients

4.1. Drag Coefficient and Apparent Roughness Length

The general definition of the drag coefficient $C_p$ is

$$C_p(z) = \left[ u^2 U(z) \right]^{1/2}$$

(1)

where $u$ is the friction velocity and $U(z)$ the mean wind speed at height $z$ above the water surface, thus making $C_p$ a function of $z$ as well.

To get an idea of the overall drag of the flow, it is pertinent to start looking at the geostrophic drag coefficient $C_g$. Then $U = G$, the geostrophic wind speed, and $z$ refers loosely to the top of the boundary layer. Periods with several pilot balloon measurements have been chosen, and $G$ has been taken as the measured wind speed at about 500 m. For hours in between the pilot balloon measurements, $G$ values have been interpolated. In this way, slightly more than 50 hours of simultaneous measured (or interpolated) $G$ - values and $u$ - values, derived from eddy correlation measurements of the alongwind component $u$ and the vertical wind $w$, have been used to derive values for the geostrophic drag coefficient $C_g$. The mean value is

$$C_g = 6.7 \times 10^{-4}$$

The variation of individual $C_g$ - values is large: from virtually 0 to about $3 \times 10^{-5}$, but there is no systematic variation with $G$.

A "typical" value for $C_g$ can be derived with a formula by Swinbank [1974]:

$$C_g = 0.0123 R_o^{0.14}$$

(2)

where $R_o = \text{surface Rossby number} = G/f \rho_0$, where $f$ is the Coriolis parameter and $\rho_0$ is the roughness length. The observed mean value for $G = 8$ m s$^{-1}$ and $f = 1.2 \times 10^{-4}$ s$^{-1}$. Assuming dynamically smooth flow,

$$\rho_0 = \frac{v}{9u^2}$$

(3)

where $v$ is the coefficient of viscosity, gives $\rho_0 = 2.5 \times 10^{-4}$ m and, with $u = 0.08$ m s$^{-1}$, $R_o = 2.7 \times 10^{-4}$. Equation (2) finally gives $C_g = 6 \times 10^{-4}$, which is about 10 times larger than the value derived directly from the measurements.

Most commonly, $C_g$ is taken to refer to the wind at 10 m above the water surface. The data set chosen for the analysis of section 5 contain 84 half-hourly data sets, representative for the meteorological situation studied here. The following mean value is obtained for $C_g$:

$$C_g = 0.34 \times 10^{-3}$$

with a standard deviation of $0.28 \times 10^{-3}$ and no systematic variation with wind speed. As discussed in section 6, $C_g$ is, however, a function of a local scale of the flow. In the wind speed range encountered here, $2 < U < 8$ m s$^{-1}$, the $C_g$ value expected during neutral conditions is about $1.2 \times 10^{-3}$ [see, e.g., Geernaert and Plant, 1990], that is more than 3 times larger than that measured. For an "ordinary" stable boundary layer an estimate of the effect of atmospheric stability on the drag coefficient can be obtained by application of Monin-Obukhov similarity. For the wind profile

$$\frac{\partial U}{\partial z} = \frac{1}{u^2} \frac{\partial \theta}{\partial z}$$

(4)

where $L = \text{Monin-Obukhov length} = -u^3 T_0 / (g k w \theta' \theta')$ with $T_0$ a reference temperature, $k$ the von Kármán constant $= 0.40$ and $w \theta'$ the kinematic heat flux $(\theta$ is virtual potential temperature).
For ϕ_m the following linear form is generally obtained:

\[ ϕ_m = 1 + Az/L \]  

(5)

Over land the best estimate for A = 5.3 [cf. Högström, 1996]. Figure 6 shows measured data for ϕ_m at Östergarnsholm plotted against z/L. Note that this plot is made up of flux measurements at 9 m and wind gradients referring to the same level only. Although the scatter is considerable, a linear trend is clearly discernable. The line drawn by eye has a slope of A = 3, that is significantly smaller than typically obtained over land. This result will be further discussed in section 6. Below it will be used to help derive an apparent z_0 value for the present flow situation.

Assuming that (5) is valid for the entire atmospheric layer below 10 m, and inserting (5) into (4) and integrating from z = z_0 to z = 10 m, gives

\[ U_{10} = \frac{u_{10}}{k} \left( \ln \left( \frac{10}{z_0} + A \frac{10}{L} \right) \right). \]

From (1) then the following expression is obtained:

\[ C_{m_{10}} = \left[ \frac{k}{\ln \left( \frac{10}{z_0} + A \frac{10}{L} \right)} \right]. \]

(6)

As mentioned above, the data set gives the following mean value for C_{m_{10}}:

\[ C_{m_{10}} = 0.34 \times 10^{-3}. \]

From Figure 6 it can be seen that the corresponding mean value for the quantity A10/L = (1 - ϕ_m) for this data set is about 3.5. Inserting this value and the above value for C_{m_{10}} in (6) results in z_0 = 10^{-7} m. This value should be compared with the z_0 value obtained with (3) for the dynamically smooth case, 2.5x10^{-5} m, which is thus 250 times larger. As a surface cannot be smoother than smooth, it must be concluded that for the present stable flow situation the ϕ_m relation (5), with A = 3 cannot be valid throughout the boundary layer, that is, from the measuring height for which Figure 6 is valid (about 10 m) and right down to the surface; see section 6.

The reduction of C_{m_{10}} noted above should be manifested in the surface wave height as well. That this is indeed the case is illustrated clearly in Figure 7, which shows measured significant wave height on the ordinate against the corresponding wave height calculated with a method outlined by Kahrna [1986], which is based on state-of-the-art knowledge in the field. For the combination of fetch in excess of 150 km and a 10-m wind speed less than about 8 m s^{-1}, the formula simply becomes

\[ H_s = 2.65 \times 10^{-2} U_{10}^2. \]

(7)

From Figure 7 it is seen that the measured significant wave height is on the average about a factor 2.5 lower than that derived with (7). In view of the quadratic dependence of H_s on the wind speed, and recognizing that the stress at the surface is the direct wave-forcing agent, this result can be interpreted as a reduction of effective C_o with the same factor, in general agreement with the above findings. This result is an indication that the flux divergence between the surface and the 10 m measuring height is probably not the main factor responsible for the small C_o value observed, see section 6 for a further discussion.

4.2. The Bulk Coefficient for the Sensible Heat Flux

The following relation defines the bulk coefficient for the sensible heat flux (or the Stanton number) C_o:

\[ H/(\rho c) = C_o U_{10} (T_{10} - T_s) \]

(8)

where H is the sensible heat flux, ρ the density of air, c, the heat capacity of air at constant pressure, U_{10} the wind speed at 10 m above the water surface, T_{10} the temperature at 10 m and T_s the water surface temperature. Figure 8 shows C_o/1.2x10^{-3} plotted against the 10 m wind speed. The asterisks represent measurements with T_{10} - T_s < 2°C and the circles represent measurements.
with $T_{m} - T_{s} > 2^\circ C$. Also shown are two curves from Geernaert and Plant [1990], which were derived from integration of nondimensional temperature profile Monin-Obukhov functions similar to (4) and (5). The two curves represent air/surface temperature differences of $2^\circ C$ and $4^\circ C$, respectively. Dashed lines have been drawn by eye to tentatively represent the trend of the two data groups. By taking points midway between these lines, representative measured values for $T_{m} - T_{s} = 2^\circ C$ can be extracted and compared with the corresponding curve from Geernaert and Plant [1990]. It is found that the ratio $C_{n}/C_{nGp}$, where $C_{n}$ is the value from this experiment and $C_{nGp}$ the corresponding value from Geernaert and Plant [1990], is less than 0.15 for $U = 4$ m s$^{-1}$, around 0.4 for 6 m s$^{-1}$ and about 0.7 for 8 m s$^{-1}$. Thus for light winds in particular the exchange of sensible heat is strongly suppressed in the present flow regime.

5. Turbulence Characteristics

5.1. Spectral Characteristics

Figures 9, 10, and 11 show examples of spectra from the experiment. The top graph of each figure gives energy spectra for the $u$, $v$ and $w$ components; the bottom graph gives the cospectrum between $u$ and $w$.

The continuity equation

\[ \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0. \]  

(10)

Assuming further that $\partial u/\partial y$ is small compared to the other terms of the continuity equation gives, after integration from the surface to the measuring height $z$,

\[ w = -z \frac{\partial u}{\partial x}. \]  

(11)

Figure 9. Example of measured spectra during a 30-min period with low turbulence. Energy spectra of the (a) longitudinal (pluses), lateral (crosses), and vertical (circles) wind components and (b) $u$-$w$ cospectrum.

Figure 8. The Stanton number $C_{n}$ divided by its neutral value $1.2 \times 10^{-3}$ plotted as a function of mean wind speed. The solid lines are from Geernaert and Plant [1990] and represent theoretically derived values for $T_{m} - T_{s} = 2^\circ C$ and $4^\circ C$ respectively. Asterisks represent measurements with $(T_{m} - T_{s})$ values < $2^\circ C$, and circles represent conditions with $(T_{m} - T_{s})$ values > $2^\circ C$. The dashed lines are drawn by eye to fit the two data sets.
A further, notable result of Figure 9 is the straight line decrease of the spectral $u$ component over the entire spectral range in this representation. The slope of the line is about -1.0, which thus differs from the -2/3 slope found for an extended mesoscale range in Högström et al. (submitted manuscript, 1997). This slope was found to be consistent with the prediction of a theory for stratified turbulence by Lilly [1983]. Although Figure 9 clearly shows that the motions in this case are quasi-horizontal down to at least the frequencies is a common phenomenon in many of the spectra of this study. In calculations of the stress, these positive low-frequency contributions are disregarded. As noted below, this part is in fact very small for the more typical cases of this study (compare Figures 10 and 11).

Inserting this in (9):

$$\frac{\partial \tilde{u}}{\partial \tilde{z}} = -\frac{\partial \tilde{u}}{\partial \tilde{x}}\left(\frac{u - z \frac{\partial \tilde{u}}{\partial \tilde{x}}}{u^{3/2}}\right)$$

(12)

With $z = 10$ m and typical measured values; $\tilde{u} / \tilde{x} = 0.1 \text{ s}^{-1}$ and $\tilde{u} = 4.0 \text{ m s}^{-1}$, shows that the terms in parenthese in (12) is positive. Then for positive $\tilde{u} / \tilde{x}$ the mesoscale $u$ value is bound to decrease with time, so that the mesoscale deviation of $\tilde{u}$ from its hourly mean is negative. From (11), $\tilde{w}$ is also negative, so that the product $\tilde{u} \tilde{w}$ is positive. The same sign is obtained for the product when $\tilde{u} / \tilde{x}$ is negative. A positive tail at low fre-
frequency 1 Hz, Lilly's theory is probably not valid for the smallest scales of motion.

The curves in Figure 10 are very different from those of Figure 9. The energy spectra for the three components are mutually similar for $n > 10^3$ Hz, and the two horizontal components show a pronounced spectral gap. The cospectrum of $u$ and $w$ has a very marked negative peak in a narrow spectral range. In comparison with this negative contribution to the covariance, the positive contribution at low frequencies is very small.

Comparing the $u$ spectral energy levels of Figures 9 and 10 in the approximate spectral region between $n = 10^3$ and $10^4$ Hz reveals that the level is about a factor 10 higher in the former case compared to the second. This is the more notable in view of the fact that the wind speed is much higher in the case shown in Figure 10, 4.3 m s$^{-1}$, compared to 2.6 m s$^{-1}$ in the case of Figure 9. Thus it is reasonable to introduce the term low-frequency spectral energy suppression for the case shown in Figure 10. A similar phenomenon was noted by Smedman et al. [1995] in the case of a wind maximum at low height above the surface of the sea.

Figure 11 shows spectra with well-developed turbulence like Figure 10 and a well-defined high-frequency -2/3 tail, indicating an inertial subrange. There are, however, two notable differences between the spectra of Figures 10 and 11: (1) The maximum energy of the $u$, $v$ and $w$ spectra is about the same in the two cases, but there is no clear spectral minimum in the case of Figure 11. Consequently, the spectral energy level for frequencies below approximately $10^3$ Hz is appreciably lower in Figure 10 compared to Figure 11. (2) The cospectral minimum in Figure 11 is at much lower frequency than in Figure 10.

In Figure 12 has been plotted the ratio between the spectral energy at spectral maximum and the corresponding spectral energy at the minimum against the wind gradient, $dU/dz$. Although the scatter is large, in particular for large wind gradient values, there is a clear increasing trend for the ratio with increasing gradient. Smedman et al. [1995] offered no explanation for the phenomenon of low-frequency spectral energy suppression. The result has, however, attracted the attention of a plasma physicist, D. E. Newman of Fusion Energy Division, Oak Ridge National Laboratory, Tennessee, who, together with colleagues, has put forward the idea that shear can inhibit eddy transport not only in plasmas but also in geophysical flows. The role of stable, sheared flow in suppressing transport has been examined extensively in toroidal plasma confinement devices [Burell, 1994]. This phenomenon is characterized by the simultaneous formation of strong shear, steep gradients of density and temperature, a marked decrease in fluctuation activity, and an increase in the time for which heat and particles are confined in the devices. While the dynamics and forcing mechanisms of the plasma turbulence are different than those of geophysical turbulence, the basic notion of shear suppression of transport remains the same. As examples, the strong meridional gradients of volcanic aerosol in the subtropical stratosphere are shown to coincide with regions of strong wind shear, and the strong ozone gradients near the edge of the stratospheric winter polar vortex are also found to correlate with layer with enhanced wind shear [Hartman et al., 1989; Ware et al., 1995]. The idea of shear suppression in geophysical flows is novel and still controversial. The results presented here tend to lend strong experimental support to the idea, which will be pursued in future collaboration between this group and that of D. E. Newman.

5.2. Turbulence Length Scales

Turbulent flow can be conveniently characterized in terms of several length scales (see Hunt et al. [1988] for an overview). It is convenient to start with the integral scales:

$$I_0 = U / \rho(t) dt.$$  \hfill (13)

Here $U$ is the mean wind speed, and $\rho(t)$ is the time autocorrelation function. $I_0$ can be determined for the three velocity components separately: $I'_u$ defining the alongwind scale for the $u$ component and $I'_v$ and $I'_w$ the corresponding alongwind scales for the lateral and vertical components, respectively. For an exponentially decaying correlation function, the wavelength where the velocity spectrum has its peak value is $\lambda_m = 2\pi I'_u$. The dissipation length scale, expressed in terms of the vertical velocity component [Hunt et al., 1988] is

$$L_v = \sigma_v / \varepsilon$$  \hfill (14)

where $\sigma_v$ is the standard deviation of the vertical velocity, and $\varepsilon$ is the energy dissipation.

A plot (not shown here) of $I'_v$ against $L_v$, where the integral length scale was obtained from spectral peak data and $\varepsilon$ from inertial subrange estimates, shows that $L_v = 0.7 I'_v$, which is not very different from the relation obtained by Smedman et al. [1995] in another study of stable flow over the Baltic Sea: $L_v = I'_v$, but which contrasts to the result obtained by Kaimal [1973] in the stable surface layer over land, which gave $L_v = 2 I'_v$. The following ratios between the various velocity component length scales are obtained on the average: $I''_u : I''_v : I''_w = 1 : 5 : 4$, in general agreement both with the results of Smedman et al. [1995] and Kaimal [1973].

Hunt et al., [1988] argue that the dissipation length scale $L_v$ is governed mainly by the ratio of the vertical velocity standard deviation and the velocity gradient. The height of the boundary layer $h_0$ and the height above the surface $z$ also enter the expression, so that

$$\frac{1}{L_v} = \frac{1}{h_0} + \frac{A_y dU}{L dz} + \frac{A_y}{z}.$$  \hfill (15)
This relation, with an $A_s$ value close to that suggested by Hunt et al. [1988], that is, $A_s = 0.5$, was found to fit the data from the stable marine boundary layer regime described by Smedman et al. [1995] quite well. When $L$ derived from the present data set is plotted (not shown here) against $\sigma_j (dU/dz)$, it is found that the parameter $A_s$ must be increased by a factor of 4 in order to fit the data. This suggests, not unexpectedly, that the present turbulent regime differs from that of an "ordinary" stable boundary layer. Some atmospheric studies quoted by Hunt et al. [1988], notably that by Hunt et al. [1985], suggest that, during certain conditions, the buoyancy length scale, $L_B = \sigma_j / N$, is the relevant scale. Here $N$ is the Brunt-Väisälä frequency:

$$N = \sqrt{g \frac{\partial \theta}{\partial z}}.$$  

If $L$ is plotted against $L_B$ (not shown here), a clear correlation is obtained, with a best fit line with a slope much larger than unity again. In view of the above findings, it is natural to test a combination of the velocity gradient scale and the buoyancy length scale:

$$\frac{1}{L_{UT}} = \frac{A_1}{\sigma_j (dU/dz)} + \frac{B}{\sigma_j / N},$$

Figure 13 shows the dissipation length scale $L_\epsilon$ plotted against $L_{UT}$ with $A_1 = B = 1.0$. Disregarding the few data for which $L_{UT}$ is larger than about 1.5 m, it is found that $L_{UT} = L_\epsilon$. It is concluded that for the present, special stable flow regime, the velocity gradient and the potential temperature gradient both determine the dissipation length scale. In the following, when $L_{UT}$ is referred to, it is implicitly implied that $A_1 = B = 1.0$.

5.3. Scaling of Turbulent Fluxes

In section 4 it was demonstrated that the following relation for the nondimensional wind gradient $\phi_h$ is approximately valid at the measuring height 9 m:

$$\phi_h = 1 + 3 z/L.$$  

This result is further discussed in the concluding section 6.

Figure 14 shows a plot of the nondimensional temperature profile $\phi_h$ against $z/L$ for the measuring height 9 m. Here

$$\phi_h = 1 + 3 z/L.$$  

but certain points, representing relatively low $z/L$ values, deviate strongly, having values in the range 10 to 40. Also this plot will be discussed in section 6.

Another form for the relation between the sensible heat flux and the potential temperature gradient is the following [Hunt et al., 1988]:

$$-w' \theta' = \gamma \sigma_j \frac{\partial \theta}{\partial z},$$

where $\gamma$ is a dimensionless parameter and $L_B$ is the buoyancy length scale defined earlier. A plot (not shown here) of $\gamma$ against the Richardson flux number,

$$Rf = \frac{g w' \theta'}{-T_0 w' w' \frac{\partial U}{\partial z}}$$

agrees in general well with that obtained from the large-eddy simulation by Mason and Derbyshire [1990] and with that of data.

Figure 13. The dissipation length scale $L_\epsilon$ plotted against the length scale $L_{UT} = \sigma_j (dU/dz + N)$. The line drawn has the slope 1:1.

Figure 14. Nondimensional potential temperature gradient $\phi_h$ plotted against $z/L$. The straight line has been drawn by eye to fit the data, with the assumption that $\phi_h = 1$ for $z/L = 0$. 

$$\phi_h = 1 + 3 z/L.$$
from the marine stable boundary layer field study by Smedman et al. [1995].

6. Discussion and Conclusions

The analysis presented in the previous sections shows that the stress on the water surface as well as the sensible heat transfer is strongly suppressed. Nevertheless, the turbulence measurements at 10 m above the water surface show conclusively that turbulence there is continuous and fully developed most of the time. In fact, the sample spectrum shown in Figure 9, which has very little turbulence and no negative contribution to the $u$-$w$ cospectrum, is an extreme case. The nondimensional wind and temperature gradients appear to adhere to Monin-Obukhov scaling, with both $\phi_m$ and $\phi_h$ increasing linearly with $z/L$. The slope of these lines is, however, much less than that observed in "ordinary" stable boundary layer data. It is possible that this reduction is due to the shear suppression effect discussed in section 5.1.

As discussed in the previous section, it is not possible to reconcile the measured wind speed and stress at 10 m through integration of the wind profiles obtained from the $\phi_m$ function (18), because it implies a physically unrealistic $z_v$ value. To explain this result, two possibilities are evident. (1) There is considerable flux divergence in the air layer between the water surface and 10 m, implying that the stress is considerably greater, about a factor of 2 is needed, than that measured at 10 m. (2) The gradients are systematically much larger in the layers below 10 m than at that height.

The wave height measurements referred to in section 4 indicate a response of the sea surface in general agreement with the observed reduction in stress, which would thus be an argument against hypothesis (1). If this is so, some evidence must be presented in favor of hypothesis (2). Data from another stable marine boundary layer study by Smedman et al. [1995] show $\phi_m$ and $\phi_h$ values very much in excess of those predicted by the Monin-Obukhov expressions for "ordinary" stable boundary layers. This anomalous result was found to occur invariably in cases with a low level-jet at low height. Based on data from the same experiment, it was shown by Bergström and Smedman [1994], that for cases without excessive shear measured $\phi_m$ and $\phi_h$ values adhered quite well to the "ideal curves," that is, linear expressions with the same values of the slope factors as obtained over land. Smedman et al. [1995] showed that the occurrence of anomalously high $\phi_m$ and $\phi_h$ values is in full agreement with the finding in that paper that (15) was indeed applicable, with the constant $A_s = 0.5$. Although this is of course pure speculation, it could be argued that the regime with anomalously high $\phi_m$ and $\phi_h$ values applies in the present case in some layer below 10 m height. Thus the scaling expressed by (17) should not be valid closer to the surface. In line with this interpretation, it is possible that the strong, very high $\phi_h$ values in Figure 14 at relatively low $z/L$ values are a manifestation of the proximity of this regime. One could possibly also interpret the few deviating high $\phi_m$ values in Figure 6 for $z/L < 1$ in a similar way.

Figure 15 displays a very puzzling result. The drag coefficient $C_D$, computed as the ratio of the measured stress and wind speed squared at 10 m, has been plotted against the scale $L_{usr}$ defined in (17). It is seen that very low $C_D$ values are obtained for small $L_{usr}$ values and that $C_D$ increases linearly to $L_{usr} = 1$, where it levels out. It is reasonable to consider also this result in terms of the alternatives (1) and (2) discussed above. Thus with alternative (1), the very low $C_D$ values for $L_{usr} < 1$ could be due to a flux divergence that increases in relative magnitude with decreasing $L_{usr}$ values. This would mean that, for small $L_{usr}$ values, local turbulence production and transport ability at the measuring height 10 m is strongly suppressed but that the magnitude of the stress would increase toward the surface of the sea. Alternative (2) would imply that the strong flux suppression manifested for low $L_{usr}$ values in Figure 15 is actually active throughout the entire atmospheric layer below 10 m.

Whichever of the alternatives (1) or (2) comes closest to the truth, in fact a combination of both is equally possible, Figure 15 expresses a significant result, that is, that a large wind gradient, in particular in combination with a large temperature gradient (which is needed to produce small $L_{usr}$ values), tends to suppress the local shearing stress. It is argued that this is a manifestation of the shear suppression hypothesis that was introduced and discussed in section 5.1.

In conclusion, fundamental features of the very stable flow situation under consideration in this paper have been elucidated, but there are still some unresolved issues. Their solution would require additional turbulence measurements at lower heights.

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