THE RESPONSE OF MASSACHUSETTS BAY TO WIND STRESS

by

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Accepted by

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ABSTRACT

THE RESPONSE OF MASSACHUSETTS BAY TO WIND STRESS

by

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The effect of atmospheric stability on the wind stress coefficient (or drag coefficient), \( C_D \), of the commonly used quadratic law is demonstrated. The method of determining values for \( C_D \) is essentially a "wind set-up" method using Doodson filtered tidal records from Boston and Sandwich, Massachusetts and similarly filtered wind and barometric pressure data. The mean values for \( C_D \) for the three stability groups are: 1.10x10^{-3} for stable conditions, 1.40x10^{-3} for neutral conditions, and 1.84x10^{-3} for unstable conditions. The wind set-up method has been objected to for various reasons, but all these objections are fairly well answered for as far as this particular investigation is concerned.

Excellent correlation exists not only between Boston-Sandwich sea level differences and the component of wind stress along the longitudinal axis of the bay (which is necessary to carry out the above method), but also between Boston-Portsmouth sea level differences and the onshore component of wind stress. Filtering Boston, Sandwich, Portsmouth, and Eastport, Maine tidal records results in very similar non-tidal sea level curves even after pressure correction. This fact, combined with the several hour lag between changes in the onshore wind stress component and sea level changes at Boston, implies that the Gulf of Maine has an important effect on Massachusetts Bay.

Wind data at Boston is used for this study but it is corrected for the frictional effect of land using the result of comparisons with other wind stations around the bay. This has apparently not been done for many wind set-up investigations and is perhaps one reason why \( C_D \)'s from wind set-up methods have usually been larger than from other methods.

Wind stress was generally much greater in the winter of 1971 than in the summer, not only because of generally higher wind speeds, but also because of greater atmospheric instability and denser air. There is, however, reason to believe that wind-driven currents (in the upper
layer of water) are greater in the summer due to thermal stratification. Current data off Salem harbor generally support this contention and also indicate the existence of internal waves.

A sceptical look is also taken at the basic quadratic law and an alternative approach is suggested.
ACKNOWLEDGEMENTS

This study represents a further step in the continuing investigation of the Massachusetts Bay. Funding for this project was provided under the Sea Grant Project, DSR 81947, Grant No. 04-5-158-1 and by the Sloan Foundation through a student fellowship to Mr. Bruce Parker.

This study developed from the need for more understanding of the air-sea interaction mechanisms in the "Mass Bay" and elsewhere. The data analyzed herein came from many sources. For this reason many people were involved.

The authors wish to especially thank Dr. Erik Mollo-Christensen, who, though recovering from a serious illness somehow still found time for numerous stimulating and helpful discussions. They were greatly appreciated.

The authors would also like to thank Doug Brogan for key-punching so much wind and pressure data. The data used in this study came from various sources. The tide data and harmonic constants were supplied by Jack Fancher of the National Ocean Survey. Lt. Richard Moore and Lt. Gary Sundin, also from N.O.S., supplied current data and ran computer programs that the author had written during his employment there. Richard Boehmer of the Massachusetts Division of Natural Resources allowed the authors access to his tapes of Boston wind data. Brad Butman of Woods Hole supplied wind data for Boston Light Vessel. And Tech Sgt. Leo Rockwood was kind enough to loan the authors
personal copies of Otis Air Force Base weather data.
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1. INTRODUCTION

In a wide body of water such as Massachusetts Bay the drift currents due to wind stress are of the same order of magnitude as the tidal currents. Therefore, if for such an area one is to develop a mathematical model that describes transport mechanisms well enough to be predictive, an accurate representation of wind stress must be included. Since the strongest wind-driven currents are in the uppermost water layers, wind stress is especially important when considering the dispersion of buoyant pollutants such as hot water discharge from a nuclear power plant, or the drift of surface material such as oil spills. Wind stress on water, of course, also changes the sea level along a coast (by several feet during large storms), causes storm surges, and generates waves. On a larger scale it is a vital cause of ocean circulation and it affects atmospheric circulation as well.

Wind stress, however, is still not a well understood subject. Even a short survey of the latest papers on air-sea interaction reveal many theories and little agreement on the mechanisms involved in transmitting momentum and energy across the air-sea interface. Even the basic quadratic law used by everyone may not really be valid (as discussed in Appendix A) and, at the very least, there is much disagreement about the value of the coefficient used in this law and the experimental techniques used to determine it.
The final objective of this investigation is the determination of values for the wind stress coefficient and the demonstration of the effect of atmospheric stability on this coefficient. The method is essentially a "wind set-up" method using a year of filtered tidal records from Boston and Sandwich (at approximately opposite ends of Massachusetts Bay) and filtered wind and pressure data. The wind set-up method has been objected to for various reasons by various authors, but all of these objections have been fairly well answered for as far as this particular study is concerned.

In the process of achieving this main objective several other results of interest are obtained. In finding a correction for the Boston wind data (which had to be used because of quality considerations but generally had lower speeds than over the bay, due to the frictional effect of land), some idea of the uniformity of the wind field over the bay is obtained.

Filtering the above mentioned tidal records, as well as the tidal records from Portsmouth, New Hampshire and Eastport, Maine, results in an idea of the effect of the Gulf of Maine on Massachusetts Bay. Excellent correlation will not only be seen between Boston-Sandwich sea level differences and the component of wind stress along the longitudinal axis of the bay, but also between Boston-Portsmouth sea level differences and the onshore component of the wind stress. As a result of this study it is apparent that wind stress is generally greater in the winter than in the summer, but
there is reason to believe that wind-driven currents in summer (in the upper layer) may actually be greater due to the thermal stratification. This is also investigated.

Numerous computer programs were written in the course of this investigation to filter and analyze various data and to plot it on a CALCOMP plotter. They were not, however, deemed unique enough to warrant inclusion in this report.
2. THE WIND FIELD OVER MASSACHUSETTS BAY

2.1 Before the response of Massachusetts Bay to wind stress can be investigated it is necessary to have some idea of the actual wind field over the bay. The chart in Figure 1 shows the locations of several weather stations on or near the bay. All anemometers were located approximately ten meters above sea level.

For the analysis to be done in succeeding chapters one would prefer quality wind data at some location in the middle of the bay. Wind data from onshore stations will have smaller speeds for certain directions due to the frictional effects of the land. Directions may also vary slightly due to local topography or the placement of the anemometer. Also, local coastline configurations affect horizontal temperature gradients thereby making the land breeze-sea breeze slightly different at each location.

Boston Light Vessel is the only station located in open water. However, it is of limited use since observations were usually made only four times a day. Of the onshore stations Race Point and Gloucester (East Point) are the most open, being located on peninsulas. As will be seen, wind speeds measured at these two stations do tend to be greater than those measured on the other side of the bay. However, data from these stations were found to be of inconsistent quality. In fact, only two stations were found to be of consistently good quality for the entire year of 1971, Logan Airport in
FIGURE 1: Weather Stations on or near Massachusetts Bay.
Boston and Otis Air Force Base on lower Cape Cod. Otis Air Force Base is not directly on Massachusetts Bay, however, and it is also affected by sea breezes from the south.

Boston wind data must therefore be used for this investigation. Thus, a correction factor must be determined so that this data may approximately represent the wind over Massachusetts Bay. Such a correction factor will result from the comparison of the Boston wind data with wind data from the other stations around the bay. Some idea of the uniformity of the wind field over the bay will also be gained through these comparisons.

2.2 The initial comparison of wind data from around Massachusetts Bay is simply the determination of a "mean wind day" for the period of June through August 1971 for each location. In such an analysis vectorial averages are made of all observations for each specific hour of the day. The results are shown in Figures 2a and 2b. An arrow drawn from the center of the polar graph to any of the connected points represents the magnitude of the mean wind and the direction toward which it is blowing at that hour. The circled point at the center of each plot represents the overall mean wind for the entire period.

Vectorially subtracting out the overall mean wind from each hourly mean would center the connected points on the origin. Such a plot centered on the origin would represent the mean land breeze-sea breeze for that period of time. During this period of time a clearly observable land breeze-
FIGURE 2a: Mean wind day for June through August, 1971, for various locations.
FIGURE 2b: Mean wind day for June through August, 1971, for various locations.
sea breeze may in fact never have occurred on any day because of an overshadowing predominant wind. However, the unequal heating of land and sea in the course of each day still contributes a diurnal component that affects such a predominant flow. The determination of a "mean wind day" brings out this hidden diurnal component.

From the plots in Figures 2a and 2b it is evident that the land breeze-sea breeze is not bidirectional as the name would imply. The direction of the wind actually rotates around the compass during a 24-hour period. Such a rotation is apparently due to the Coriolis effect (A land breeze-sea breeze is bidirectional on the equator.). The results for Boston, Boston Light Vessel, Gloucester, and Race Point show clockwise rotations as one would expect from a Coriolis effect. The results for Portsmouth and Sandwich, however, show a counterclockwise rotation, and Otis Air Force Base shows a figure-eight result with both clockwise and counterclockwise rotations. Results for another period (May 20-June 23, 1973), not shown here, again give counterclockwise rotations for Portsmouth and Sandwich, as well as for Otis and Hyannis (on the south shore of lower Cape Cod, east of Otis), while the remaining stations were again clockwise. Apparently other factors can cause or modify the rotation of directions. It also seems apparent from the plots in Figures 2a and 2b that the local coastline configuration does have an effect on the land breeze-sea breeze. The results from the stations on lower Cape Cod are certainly complicated by
the fact that there is a large body of water on both sides of the Cape and that the Cape is of sufficient width to set up a second smaller scale horizontal temperature gradient. For more detailed theoretical information on the land breeze-sea breeze see Walsh (1973), Defant (1951), Haurwitz (1947), and Schmidt (1947).

The land breeze-sea breeze is not a year round feature in the Massachusetts Bay area. Mean wind days at Boston for other periods of 1971 are shown in Figures 3a and 3b. As would be expected the result for January through March shows no appreciable land breeze-sea breeze and the November-December result shows only a small diurnal component. Table I also lists the mean wind speed and direction for each month of 1971. The winds for the winter months are on the average greater than those for the summer months.

2.3 Perhaps the simplest way to compare wind data from various stations around the bay is by examining time series plots of speed and direction. Plots for several of these stations for the period of July 31 through August 29, 1971 are shown in Figures 4a through 4f. Some similarity can be seen in major speed and direction changes, which correspond to low wind frequencies. Higher frequency wind would not be expected to be very similar for stations some distance apart.

In this investigation, however, low frequency wind data will be used, for reasons to be explained in Chapter 3. Boston wind data will be filtered with a 39-hour Doodson filter (also to be explained in Chapter 3), so comparisons
FIGURE 3a: Mean wind day at Boston (Logan) for various periods of 1971.
FIGURE 3b: Mean wind day at Boston for Nov.-Dec., 1971.

TABLE I: Mean Boston Wind for 1971

<table>
<thead>
<tr>
<th>Month</th>
<th>Speed (Knots)</th>
<th>Direction Toward (Degrees true)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>9.6</td>
<td>102</td>
</tr>
<tr>
<td>Feb</td>
<td>3.7</td>
<td>100</td>
</tr>
<tr>
<td>Mar</td>
<td>7.1</td>
<td>109</td>
</tr>
<tr>
<td>April</td>
<td>3.6</td>
<td>127</td>
</tr>
<tr>
<td>May</td>
<td>0.8</td>
<td>81</td>
</tr>
<tr>
<td>June</td>
<td>2.5</td>
<td>49</td>
</tr>
<tr>
<td>July</td>
<td>3.8</td>
<td>47</td>
</tr>
<tr>
<td>Aug</td>
<td>3.4</td>
<td>82</td>
</tr>
<tr>
<td>Sept</td>
<td>0.7</td>
<td>105</td>
</tr>
<tr>
<td>Oct</td>
<td>0.1</td>
<td>67</td>
</tr>
<tr>
<td>Nov</td>
<td>3.9</td>
<td>125</td>
</tr>
<tr>
<td>Dec</td>
<td>5.8</td>
<td>115</td>
</tr>
</tbody>
</table>
Directions are toward which wind is blowing (degrees true).
FIGURE 4b: Time series plots of Massachusetts Bay wind data for July 31 - August 29, 1971. Directions are toward which wind is blowing (in degrees true).
FIGURE 4c: Time series plots of Massachusetts Bay wind data for July 31 - August 29, 1971. Directions are toward which wind is blowing (in degrees true).
FIGURE 4d: Time series plots of Massachusetts Bay wind data for July 31 - August 29, 1971. Directions are toward which wind is blowing (in degrees true).
FIGURE 4e: Time series plots of Massachusetts Bay wind data for July 31 – August 29, 1971. Directions are toward which wind is blowing (in degrees true).
FIGURE 4f: Time series plots of Massachusetts Bay wind data for July 31 - August 29, 1971. Directions are toward which wind is blowing (in degrees true).
should really be made with filtered wind data. (Such filtering will also remove the land breeze-sea breeze, since it has a 24-hour period, as well as the higher frequencies.)

Figures 5a through 5c show Doodson filtered wind data for Boston, Gloucester, and Race Point, for the period of July 31 through August 29. These low frequency plots are a good deal more similar than their unfiltered counterparts (Note that the vertical scale for speed is different for the filtered plots.). The changes in direction appear to be almost identical, although the actual directions may be different by 20°. Boston speeds are generally lower than those at Gloucester and Race Point, the amount depending on the wind direction. (When the wind is out of the west or north-west, however, Gloucester speeds are lower than Boston speeds, probably due to local topography and/or the placement of the Gloucester anemometer.) Generally, it would seem that the low frequency wind field over Massachusetts Bay is more uniform than the unfiltered wind field.

In Chapter 4 the 330°-150° True component of the Doodson filtered and squared wind will be of prime interest (i.e. the minor component of the low frequency quasi-wind stress). Therefore, comparisons should be made of this component for each station. Table II gives the following wind values for Boston, Boston Light Vessel, Gloucester, and Race Point, using data for June through August 1971: the vector mean; the absolute mean of the speeds and the major and minor components; and the absolute mean of the filtered wind speeds
FIGURE 5a: Time series plots of Doodson filtered wind data for Boston, Gloucester, and Race Point, for July 31 - August 29, 1971.
FIGURE 5b: Time series plots for Doodson filtered wind data for Boston, Gloucester, and Race Point, for July 31 - August 29, 1971.
FIGURE 5c: Time series plots for Doodson filtered wind data for Boston, Gloucester, and Race Point, for July 31 - August 29, 1971.
TABLE II. Comparison of Wind Stations.

<table>
<thead>
<tr>
<th>Wind Station</th>
<th>Vector Mean Direction</th>
<th>Absolute Mean of Components</th>
<th>Absolute Mean of Doodson Filtered Components</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Speed</td>
<td>Toward</td>
<td>Speed</td>
</tr>
<tr>
<td>Boston (Logan)</td>
<td>3.1</td>
<td>60°</td>
<td>8.4</td>
</tr>
<tr>
<td>Boston L/V</td>
<td>4.8</td>
<td>55°</td>
<td>9.1</td>
</tr>
<tr>
<td>Gloucester</td>
<td>3.8</td>
<td>13°</td>
<td>9.1</td>
</tr>
<tr>
<td>Race Point</td>
<td>5.3</td>
<td>33°</td>
<td>11.3</td>
</tr>
</tbody>
</table>

All speed in Knots. (1 Knot = .51 m/sec)
All directions in degrees true.
Major component is 240°-60° true.
Minor component is 330°-150° (longitudinal axis of Massachusetts Bay). (See Figure 1.)
and major and minor components. It should be remembered that the results for Boston Light Vessel are based on only four observations a day. The hours of these observations, i.e. 0100, 0700, 1300, and 1900, would appear from Figure 2a to give mean values that are probably slightly too small.

It was decided that the Race Point wind data, if it had been of consistently good quality, would have most closely represented the wind over most of Massachusetts Bay. Therefore, the Race Point result was used to determine a correction factor. The absolute mean of the minor component of the filtered wind at Race Point is approximately 1.63 times that at Boston. This value squared, i.e. 2.657, will be used in Chapter 4 to correct the Boston quasi-wind stress values. It should be mentioned that, if this value is in error, it is probably too small, as winds were generally weaker in the summer than in the rest of the year.
3. SEA LEVEL RESPONSE TO WIND STRESS

3.1 Two tide gauges are maintained in Massachusetts Bay by the National Ocean Survey (N.O.A.A., Dept. of Commerce), one at Pier No. 5 in Boston Harbor and one at Sandwich, Massachusetts, at the entrance to the Cape Cod Canal (See Figure 6.). Hourly tide data for the year of 1971 were acquired from N.O.S. for these two locations, along with data for Portsmouth, New Hampshire and Eastport, Maine. The latter two stations were considered in order to gain some idea of the effect of the Gulf of Maine on Massachusetts Bay.

Under nonhurricane conditions the greatest change in sea level along the New England coast is of course due to the astronomic tide. The mean tidal range varies from about 9 feet at Sandwich to about 21 feet at Eastport. The sea level change due to nontidal causes (mainly barometric pressure changes and wind stress) normally has less than a two foot range under nonhurricane conditions. (Two feet was in fact the maximum nontidal range for 1971, occurring on March 3rd to 5th as a result of moderately high winds and a record low pressure.).

The tide of course must somehow be removed from the data in order to see the nonperiodic effects on sea level. Two approaches will be tried. The simplest method is to eliminate the tide using a numerical filter. A more involved method entails doing a careful harmonic analysis of the data, then
FIGURE 6: Locations of N.O.S. tide stations.
predicting the tide using the resulting harmonic constants and subtracting these predictions from the actual tidal record.

3.2 Numerical filtering is simply the taking of weighted averages, the weights and lengths of the average being chosen to eliminate one or more specific frequencies. The standard way of representing such a filter is:

\[ y(t) = \sum_{i=-n}^{n} W_i x(t-i) \]

where \( x(t) \) is the input time series (i.e. the tidal record), \( W_i \) are the weights, and \( y(t) \) is the filtered result (i.e. the nontidal sea level). The above formula is applied to 2n+1 data points of the data record and a single result is obtained and assigned to the center data point. Then the filter shifts over one data point and another result is obtained. Continuing this process results in a new time series. The weights are chosen such that:

\[ W_i = W_{-i} \quad \text{and} \quad \sum_{i=-n}^{n} W_i = 1 \]

The simplest filter is a running mean, for which, in the above representation, all the \( W_i \)'s are equal to \( 1/(2n+1) \). Since the tidal frequencies with the greatest amplitudes are 1/12.42 hours (\( M_2 \)), 1/12.00 hours (\( S_2 \)), 1/12.66 hours (\( N_2 \)), and 1/24.84 hours (\( O_1+K_1 \)), a 25-hour running mean would eliminate a good portion of the tide from the time series.
For our purposes, however, such a filter leaves too much of the tide in the data. An excellent introduction to filtering may be found in Holloway (1958).

Numerous filters have been developed to eliminate tidal frequencies (See Godin (1972) and Groves (1955).), each having specific \( w_i \)'s and a certain length \( 2n+1 \). The filter chosen for this investigation was developed by Doodson and Warburg (1941) and is hereafter referred to as the Doodson filter. It is a 39-hour filter (i.e. \( 2n+1 = 39 \) for hourly data points) and has the following weights:

\[
W_i = \begin{cases} 
0 & \text{for } i = 0,5,8,10,13,15,16,18 \\
1/30 & \text{for } i = 2,3,6,7,11,12,14,17,19 \\
2/30 & \text{for } i = 1,4,9 
\end{cases}
\]

According to Groves (1955) the Doodson filter eliminates 99.79% of the diurnal frequencies (e.g. \( O_1, K_1, S_1 \), etc.), 99.94% of the semidiurnal frequencies (e.g. \( M_2, S_2, N_2, K_2 \), etc.), and 99.38% of the higher frequencies (e.g. \( M_4, M_6, M_8 \), etc.).

In using this or any other filter it must be emphasized that those nonperiodic contributions to the time series do not remain untouched. The nonperiodic result of using a filter is in reality a distorted version of the true nonperiodic contribution. Nonperiodic changes taking place within a period of 39 hours will be greatly reduced. Even a nonperiodic change over 48 hours will be reduced by 60%. Of course, the longer the duration of the change the less it will be distorted by the filter.
Because the nonperiodic contributions to sea level change are somewhat distorted by the filtering process, any other types of data with which the results of this filtering are to be compared must also be themselves similarly filtered. In other words, wind data and barometric pressure data must also be Doodson filtered before one tries to correlate them with nontidal sea level changes.

1971 tidal records for the four locations in Figure 6 were Doodson filtered. Results for the months of January and February are shown in Figures 7a through 7f. The similarity in these curves will be discussed in Section 3.4.

3.3 Because of the distortion brought about by filtering and because of the desire to look at nonperiodic changes of a shorter time scale, a second method of removing the tide was tried. Harmonic constants for Boston, Sandwich, and Portsmouth were obtained from N.O.S. and hourly predictions were made for 1971. These predictions were then subtracted from the actual hourly tide records for that year. If the harmonic constants are accurate the difference between actual and predicted tide should represent the sea level changes due to nontidal causes. This would be an approximately "real time" result with little distortion.

The results for the first twenty days in January are shown in Figure 8. The Boston result stands out immediately. There is obviously still some tide left in the data. A sinusoid of approximately 0.5 foot amplitude and 12.42 hour period is clearly discernible superimposed upon a slower
FIGURE 7a: Doodson filtered tide data and Boston barometric pressures (converted into sea level differences).
FIGURE 7b: Doodson filtered tide data and Boston barometric pressures (converted into sea level changes).
FIGURE 7c: Doodson filtered tide data and Boston pressures (converted into sea level changes).
FIGURE 7d: Doodson filtered tide data and Boston barometric pressures (converted into sea level differences).
FIGURE 7e: Doodson filtered tide data and Boston barometric pressures (converted into sea level changes).
FIGURE 7f: Doodson filtered tide data and Boston barometric pressures (converted into sea level changes).
FIGURE 8: Tidal record minus predicted tide for Portsmouth, Boston, and Sandwich (January 1-20, 1971).
variation similar to that in Figures 7a and 7b (Note, however, that the vertical scales in these figures are different.). Upon closer examination it was found that the range of the actual tide was generally about one foot greater than the predicted tide and that there was a slight phase shift. (See Figure 9.) The N.O.S. harmonic constants for Boston were obtained from a least squares harmonic analysis of five years of data, the most recent of which was 1940. Since that time Boston Harbor has been deepened and widened and this would account for the increased range and slight phase lag. Corroborating this is the fact that 1971 current data taken by N.O.S. in Boston Harbor have smaller tidal current velocities than data from a similar survey in 1952. The nineteen year wait between surveys is fortunate as that eliminates any possible astronomic differences (there being an 18.6 year cycle for the westward motion of the lunar node, which has an appreciable effect; see Schureman (1940) or Smart (1971)).

The harmonic constants for Sandwich and Portsmouth were based on least squares harmonic analyses of one year of data, 1971 and 1970 respectively. The results are considerably better than for Boston. The curves in Figure 8 are similar to those in Figures 7a and 7b. (Again note that the vertical scales are different.).

There was not sufficient time to carry out a careful harmonic analysis of the 1971 Boston data. Therefore, it was decided to use the "Doodsoned" results. Because of this,
all wind and pressure data used later will also be "Doodsoned."

According to Miller (1958) there are benefits to be gained by using the filtering method. For one thing filtering reduces "noise" in the data. It also minimizes short duration oscillatory phenomena caused by wind, such as long waves and storm surges. Groves (1955) also believes that a greater percentage of the tide is removed by filtering than by subtracting out predicted values. When comparing the Portsmouth curves obtained by the two methods it would seem that the reduction in nonperiodic changes by filtering may be less than had been expected (Although it would not be a problem if there was a great reduction, since wind and pressure data will also be filtered.). An additional benefit, as was mentioned in Chapter 2, seems to be that the filtered wind field (i.e. the low frequency wind field) over Massachusetts Bay is more uniform than the unfiltered wind field. Also, the main reason for wanting to see a "real time" result would be to see the effect of the land breeze-sea breeze. But such a 24-hour (S1) periodicity would probably be partially hidden by the K1 harmonic constant anyway, so that even the second method would be of little use (unless many years of data were analyzed).

3.4 The nontidal sea level curves in Figures 7a through 7f for Sandwich, Boston, and Portsmouth are very similar. Even Eastport which is 275 miles up the coast from Massachusetts Bay is very similar. Sea level barometric pressure
changes at Boston (Logan Airport) were converted into sea level changes (30.7 millbar pressure change equal approximately 1 foot of sea level change) and Doodson filtered. The results for January and February are also plotted in Figures 7a through 7f.

It is apparent that a large portion of the nontidal sea level changes at these four stations is due to similar barometric pressure changes. The filtered Boston tide data was then corrected for barometric pressure changes; the results for January, February, and March are shown in Figures 10a through 10c. These curves are still similar to the uncorrected curves. Sea level pressures from Otis Air Force Base near Sandwich were also converted and filtered and used to correct the filtered Sandwich tide data. Again the corrected curves were similar to the uncorrected curves. One would expect the same to be true for the other two locations. (It will be noticed from the Boston sea level and pressure curves in Figures 7a through 7f, that the sea level changes generally precede the pressure changes by a few hours, a result also gotten by Miller (1958). Thus, one would expect the corrected sea level curves to be of a shape similar to the uncorrected curves.).

The remaining similarity after pressure correction implies that there is still another common cause of the similar nontidal sea level changes. One would guess that the low frequency wind field might be similar near these four locations. Even if the low frequency wind field over the Gulf of
FIGURE 10a: Pressure corrected Boston nontidal sea level and $240^\circ-60^\circ$ component (referred to as "major" component) of filtered and squared wind.
FIGURE 10b: Pressure corrected Boston nontidal sea level and 240°-60° component of filtered and squared Boston wind.
FIGURE 10c: Pressure corrected Boston nontidal sea level and 240°-60° component of filtered and squared Boston wind.
Maine is not uniform, the Gulf of Maine responding as a whole to a nonuniform wind field should give similar sea level variations along its western shore. As will be discussed in the next section, the fact that the Boston pressure corrected nontidal sea level changes seem correlated with wind changes, but with a few hours delay, implies a large fetch, i.e. the Gulf of Maine. Thus, it would seem that the Gulf of Maine has an important effect on Massachusetts Bay nontidal sea level changes.

3.5 Wind data for Boston (Logan Airport) was Doodson filtered and squared. The component of this quasi-wind stress along a 240°-60° True axis (i.e. approximately perpendicular to the longitudinal axis of Massachusetts Bay) is also plotted in Figures 10a through 10c, along with the pressure corrected nontidal sea level at Boston. There does seem to be a correlation, but with a varying lag of at least a few hours (Similar results have been gotten by Miller (1958) and Butman (1974)). Such a lag implies that the wind is blowing over a large and probably varying fetch, i.e. all or part of the Gulf of Maine. The fact that the exact fetch is unknown and not constant would make it very difficult to accurately determine a wind stress coefficient from this data.

However, if the wind effect on Massachusetts Bay alone could be isolated, then the fetch would be known and constant, and a wind stress coefficient could be determined using a simple analytical model. There are two ways this might be
done using the filtered tide data from Sandwich, Boston, and Portsmouth.

Since Boston and Sandwich are at approximately opposite ends of Massachusetts Bay, the differences in nontidal sea level at these two locations should be correlated with the component of the wind stress along the longitudinal axis of the bay (i.e. the 330°-150° True axis). In other words, if the wind is blowing toward the Sandwich end of the bay one would expect the nontidal sea level at Sandwich to be higher than at Boston, and vice versa. The difference in nontidal sea levels at Boston and Sandwich should give an estimate of the longitudinal tilt of the water surface of Massachusetts Bay, corresponding to a certain wind stress in that direction.

If we make the rough assumption that the filtered Portsmouth tide data can represent the nontidal sea level at the open sea boundary of Massachusetts Bay, then the differences in nontidal sea level at Boston and Portsmouth should be correlated with the component of the wind stress along a 240°-60° True axis. In other words, it is assumed that if the wind is blowing toward Boston and Portsmouth, then the nontidal sea level at Boston will be higher than at Portsmouth due to the added fetch of Massachusetts Bay, and vice versa.

Figures 11a through 11c show plots of the differences in nontidal sea level at Boston and Sandwich and plots of the 330°-150° component of the filtered and squared wind,
FIGURE 11a: Difference in Boston and Sandwich nontidal sea levels and 330°-150° (called "minor") component of filtered and squared wind.
FIGURE 11b: Difference in Boston and Sandwich nontidal sea levels and the 330°-150° component of filtered and squared Boston wind.
FIGURE 11c: Difference in Boston and Sandwich nontidal sea levels and the 330°-150° component of filtered and squared Boston wind.
for January, February, and March. Figures 12a through 12c show plots of the differences in nontidal sea level at Boston and Portsmouth and plots of the $240^\circ-60^\circ$ component of the filtered and squared wind, for the same months. These uncorrected nontidal sea level differences are caused by both wind stress and barometric pressure differences between the locations. Since these locations are not far apart, the barometric pressure differences are usually small; but the sea level differences caused are still on the same order as those caused by wind stress. In January the winds were moderately strong, so the sea level difference curves correlate very well with the quasi-wind stress curves. Figure 13, however, shows a period in June and July when winds were weak. During this period the sea level differences between Boston and Sandwich correlate well with the pressure differences. In the next chapter, when a wind stress coefficient is actually calculated the nontidal sea level differences will be corrected for barometric pressure differences.

Even without this pressure correction the correlation is generally very good for both cases. There also seems to be a smaller lag (often zero) between wind stress changes and sea level changes. This would be expected if the sea level differences were indeed due only to wind stress on Massachusetts Bay alone. Butman (1974) has gotten similarly good results using 1972 Boston and Sandwich tide data filtered with a 30-hour Gaussian filter.
FIGURE 12a: Difference in Boston and Portsmouth nontidal sea levels and the $240^\circ-60^\circ$ component of the filtered and squared Boston wind.
FIGURE 12b: Difference in Boston and Portsmouth nontidal sea levels and the 240°-60° component of the filtered and squared Boston wind.
FIGURE 12c: Difference in Boston and Portsmouth nontidal sea levels and the 240°-60° component of the filtered and squared Boston wind.
FIGURE 13: Difference in Boston and Sandwich nontidal sea levels and the difference in Boston and Otis A.F.B. filtered sea level pressures (converted into sea level differences).
As expected, the Boston-Sandwich results are better than the Boston-Portsmouth results, the latter having the poorer assumptions to begin with. For one thing, the Boston-Portsmouth sea level differences should also be affected by the 330°-150° component of the wind stress, since one would expect little effect on Portsmouth but a definite effect on Boston. The assumption that the 240°-60° component of wind stress will affect Boston and Sandwich in the same manner, is also not really true. The difference in effects, however is on a smaller scale, being due mostly to Coriolis effects during strong onshore winds. This will be discussed in Chapter 4.

The Boston-Sandwich results for 1971 will be used in Chapter 4 to determine a wind stress coefficient and to investigate the effect of atmospheric stability on wind stress.
4. THE DETERMINATION OF A WIND STRESS COEFFICIENT AND
THE EFFECT OF ATMOSPHERIC STABILITY

4.1 The usual starting point for calculating wind stress
from wind speed for use in the equations of motion is an
equation of the form:

\[ \tau_s = C_D \rho_a U^2 \]

where \( \tau_s \) = the wind stress
\( \rho_a \) = the density of air
\( U \) = the wind velocity at a
specific height above
sea level (usually 10m)
and \( C_D \) = the wind stress coeffi-
cient, or drag coeffi-
cient.

This equation was borrowed from fluid flow over a solid sur-
face with the assumption that, relative to the air, the water
is practically solid. The legitimacy of this equation and
\( C_D \) will be discussed in Chapter 5. Suffice it to say for the
moment that \( C_D \) does not seem to be constant for the air-sea
situation. For flow over solids the only parameters usually
considered as affecting the value of \( C_D \) are roughness and
Reynold's number, both of which are usually constant for a
given situation. For the air-sea situation, however, rough-
ness is not well defined nor constant. \( C_D \) seems to depend
in some undecided way on wind speed, for some unagreed upon
reasons. \( C_D \) also seems to depend on the stability of the
atmospheric boundary layer (as will be demonstrated by this
investigation), so that temperature differences between air
and water are important. The humidity of the air may also have an effect. More of this will be discussed later.

4.2 A rather simple derivation of the equations of motion including wind stress, $\tau_\omega$, results from the following diagram:

![Figure 14](image)

Here hydrostatic pressure is assumed. Balancing forces with change in momentum (in the x-direction) one obtains:

$$\rho_w(h+\eta)\frac{Du}{Dt} = \rho_w(h+\eta)fv \Delta x + \left[ \frac{1}{2}\rho_w g(h+\eta)^2 - \frac{1}{2}\rho_w g(h+\eta+\eta \frac{\partial \eta}{\partial x} \Delta x)^2 - \tau_{b_x} \Delta x + \tau_{s_x} \Delta x \right]$$

where $\rho_w = $ the density of water

and $g = $ the acceleration due to gravity.

This leads to:

$$\frac{Du}{Dt} = fv - g \frac{\partial \eta}{\partial x} + \frac{\tau_{s_x} - \tau_{b_x}}{\rho_w(h+\eta)}$$  \hspace{1cm} (1)
And similarly for the y-direction:

$$\frac{Dv}{Dt} = -fu - g \frac{\partial \eta}{\partial y} + \frac{\tau_{xy} - \tau_{xv}}{\rho_w (h + \eta)} \quad (2)$$

The equation of continuity would be:

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = - \frac{\partial \eta}{\partial t} \quad (3)$$

It should be remembered that \( u \) and \( v \) are depth integrated velocities, i.e.

$$u = \int_{-h}^{\eta} u' dz \quad \text{and} \quad v = \int_{-h}^{\eta} v' dz$$

where \( u' \) and \( v' \) are velocities at any depth.

In equations (1) and (2),

$$\tau_{sx} = \tau_s \cos \theta = C_b \rho_a U^2 \cos \theta \quad (4)$$

$$\tau_{sy} = \tau_s \sin \theta = C_b \rho_a U^2 \sin \theta \quad (5)$$

where \( \theta \) = the angle of the wind with the x-axis. For simplicity the effects of atmospheric pressure have not been explicitly included. Instead, all water elevations will be assumed corrected for atmospheric pressure changes in the manner mentioned in Chapter 3.

A simple one dimensional approach will be taken, so only motions in the x-direction will be considered. The equations reduce to:

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} = - g \frac{\partial \eta}{\partial x} + \frac{\tau_{sx} - \tau_{bx}}{\rho_w (h + \eta)}$$

and

$$\frac{\partial u}{\partial x} = - \frac{\partial \eta}{\partial t} .$$
(By this assumption we are essentially ignoring Coriolis effects.)

Assuming steady state conditions and uniform and constant shear stresses, we have:

\[
-u \frac{\partial u}{\partial x} = -g \frac{\partial \eta}{\partial x} + \frac{\tau_{sx} - \tau_{bx}}{\rho_w (h + \eta)}
\]

and

\[
\frac{\partial u}{\partial x} = 0.
\]

So the advection term disappears, leaving:

\[
\frac{\partial \eta}{\partial x} = \frac{\tau_{sx} - \tau_{bx}}{\rho_w g (h + \eta)}
\]

i.e. the slope of the water surface is directly proportional to the surface stress minus the bottom stress.

Next the case of an enclosed rectangular channel of constant depth is considered.
The boundary conditions of \( u = 0 \) at \( x = 0 \) and \( x = L \), and the continuity equation \( \frac{\partial u}{\partial x} = 0 \), lead to the fact that \( u = 0 \) everywhere along the channel. It should again be remembered that this is a depth integrated velocity. Actually, near the surface there will be a current going with the wind and in the rest of the depth a return current caused by the end wall.

![Velocity profile](image)

**FIGURE 16:** Velocity profile for water movement in a closed rectangular channel (from Baines and Knapp, 1965.)

The value of \( \tau_{b\alpha} \) depends on the magnitude of the return current. As will be seen later, this return current is very small and therefore \( \tau_{b\alpha} \) may be assumed negligible. As shown in Figure 15, to the approximations made, the uniform constant wind stress on the water surface is balanced by the pressure gradient of the piled up water.

Integrating equation (6) the displacement of the water surface, \( \gamma \), at either end can be found. Neglecting \( \gamma \)...
relative to \(h\), one obtains:

\[
\eta = \frac{\tau_s}{\rho_w gh} x + c_1
\]

\(c_1 = \text{Constant}\)

At \(x = 0\),

\[
\eta_0 = c_1
\]

and at \(x = l\),

\[
\eta_l = \frac{\tau_s}{\rho_w gh} l + c_1
\]

Then the total difference in water level at the ends of the channel is:

\[
\eta_t = \eta_l - \eta_0 = \frac{\tau_s l}{\rho_w gh}
\]  

(7)

It is apparent that the total water level difference between the ends of a rectangular channel is directly correlated with the wind stress. To a first approximation Massachusetts Bay may be considered a closed rectangular basin (The effect of the eastern open sea boundary will be discussed later, along with other assumptions made.). See Figure 1.

Substituting equation (4) into equation (7) and rearranging one obtains:

\[
C_D = \frac{\eta_t \rho_w}{f \alpha} \frac{gh}{U^2 \cos \theta}
\]

(8)

Thus, values for the wind stress coefficient, \(C_D\), can be determined from the total difference in sea level, \(\eta_t\).

Determining \(C_D\) in this manner is usually referred to as the "wind set-up method," the "slope method," or the "tilt method." In the past, several objections have been raised
about its validity, which will be discussed in Section 4.5.

At this point four basic assumptions have been made which will be discussed in Section 4.5 as to the degree of their validity: (1) the basic one dimensional assumption, which really assumes that \( v = 0 \) and thus that there will be no Coriolis effect; (2) the steady state assumption, i.e. that an equilibrium is reached; (3) the assumption that \( \tau_b \) is negligible; and (4) the assumption that the open boundary has little effect on the results (which is related to assumption (1)).

It was felt that a more exact relationship than equation (8) might be obtained if bottom topography were considered. Two methods were tried, one assuming a linearly sloping bottom and the other (using a method developed by Keulegan (1953)) considering the bottom profile of mean cross section depths.

However, these methods resulted in negligible differences from the results using a simple one dimensional rectangular basin.

Therefore, equation (8) will be used with the Boston-Sandwich data to estimate \( C_D \).
4.3 When wind speeds are measured a good distance above the water surface (as they usually are, the standard being 10 meters) wind stress at the water surface should theoretically depend on the stability of the layer of air above the surface, i.e. on the degree of stratification of the air over the water. If the temperature of the air increases with height, then volumes of air displaced upwards will be heavier than the surrounding air and will tend to sink back down. Likewise, volumes of air displaced downwards will be lighter than the surrounding air and will also tend to return to their original elevations. In this "stable" situation turbulence is reduced and the transfer of momentum and energy across the air-sea interface is decreased. If the air temperature decreases with height, then displaced volumes of air tend to be accelerated and turbulent transfer is increased for this "unstable" situation. If air temperature is approximately constant with height, displaced volumes of air tend to remain at their new positions, neither hindering nor helping turbulent transfer for this "neutral" situation.

We would expect, for the same 10m wind speed, the greatest wind stress for unstable conditions and the least for stable conditions. Such an increase or decrease in wind stress due to stability will be represented by an increase or decrease in $C_D$ in the quadratic formula $\tau_s = C_D \rho \alpha U^2$.

At the present time there is limited experimental support
for such stability effects on wind stress. Roll (1952)*, studying sea states, found that for a given wind speed, mean wave heights were greater during unstable conditions than during neutral conditions. Brown (1953) and Fleagle (1956) have reported similar results. More recently DeLeonibus (1971), using an eddy correlation method, found that $C_D$ on the average decreased with increasing stability (e.g. $C_D = (1.3+6)10^{-3}$ for unstable; $C_D = (1.2+.4)x10^{-3}$ for neutral; and $C_D = (0.8+.4)x10^{-3}$ for stable conditions; where the wind speed was measured at 7.5m above the surface instead of the usual 10m). Smith (1970) attributes his wider scatter at lower wind speeds to stability effects, but found no clear correlations. Brocks (reported in Kraus (1972)) using a wind profile method also got wider scatter at lower speeds. Decreasing scatter in the data with increasing wind would be expected since stratification is decreased by strong winds.

The only investigation to consider stability effects using a wind set-up method was done by Darbyshire and Darbyshire (1955) at Lough Neagh in Northern Ireland. They found that, at a given wind speed, the set-up for unstable conditions was approximately twice as great as the set-up for stable conditions. Decon and Webb (1962), however, felt that these results were obscured by the fact that the wind

*Reported in Deacon and Webb (1962); paper in German.
measurements used were taken over land a couple of miles from the lake shore. Darbyshire and Darbyshire apparently felt this less of a problem, since they state that the country around Lough Neagh is flat. There still must be some frictional effects, however.

In the present investigation the stability at the time of an observation will be classified according to the following criteria based on air and water temperature differences:

- **Unstable** \( (T_a-T_w) < -3^\circ F \)
- **Neutral** \( -3^\circ F \leq (T_a-T_w) \leq 3^\circ F \)
- **Stable** \( (T_a-T_w) > +3^\circ F \)

where \( (T_a-T_w) \) = the difference in air and water temperatures.

This division is somewhat arbitrary. Darbyshire and Darbyshire used \(+2^\circ F\) as the boundaries for neutral conditions, which seemed a slight bit too narrow. It was of course not possible to use a gradient Richardson number classification of stability, since wind speeds and temperatures were not known at two elevations above the water surface.

4.4 The differences in nontidal sea level at Boston and Sandwich (examples of which were plotted in Section 3.5) were corrected for sea level pressure differences between Boston and Otis Air Force Base (which is not far from Sandwich and was used because the Sandwich weather data proved to be of questionable quality). The assumption was then made that these sea level differences were due solely to changes in the \(330^\circ-150^\circ\) component of the wind stress.
(This assumption is discussed in Section 4.5.).

Table III gives 51 maximum sea level differences taken from the 1971 data and the corresponding maximum 330°-150° component of the filtered and squared wind at Boston (which will be corrected according to the result of Chapter 2). The table also gives \((T_a - T_w)\) and a stability classification according to the criteria of Section 4.3. All temperatures were averaged over a 48-hour period centered at the time of the maximum sea level and wind stress changes. Water temperatures were only available at Sandwich, but since these varied so slowly they were deemed acceptable even though the rest of the Sandwich weather data were of questionable quality. The air temperatures used were from Otis Air Force Base.

Figure 17 shows a plot of the sea level differences versus the 330°-150° component of the filtered and squared wind. For convenience three straight lines were drawn through the three stability groups of data points, even though there were not necessarily good straight line fits. Although there is a certain amount of scatter there does seem to be evidence for increased set-up with unstable conditions and decreased set-up with stable conditions. It will also be noticed that all the stable data points are below the neutral line and all the unstable data points are above the neutral line. (The greatest scatter is in the neutral data points, which may be partially a result of the definition of this group). It should be remembered that this is essentially low wind
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<td>DATE</td>
<td>(Feet)</td>
<td>$u^2$ (Knots)$^2$</td>
<td>$+\ or\ -$</td>
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<td>STABILITY</td>
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<td>-</td>
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<td>- 9</td>
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<td>- 4</td>
<td>U</td>
<td>3.17</td>
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</tbody>
</table>

$+\ means\ Boston\ higher\ than\ Sandwich$

$-\ means\ Sandwich\ higher\ than\ Boston$

$S =\ Stable$

$N =\ Neutral$

$U =\ Unstable$
WIND SET-UP vs WIND SPEED SQUARED

- ○ UNSTABLE
- ▲ NEUTRAL
- □ STABLE

FIGURE 17: Maximum differences in Boston and Sandwich corrected nontidal sea levels versus $330^\circ$-$150^\circ$ component of filtered and squared Boston wind.
data, where stability would have its greatest effect. At higher wind speeds any stratification of the atmospheric boundary layer will be quickly destroyed.

Equation (8) will be used to determine values of $C_D$:

$$C_D = \frac{n_t \rho_w gh}{\rho_a \ell} \frac{1}{U^2}$$

where $n_t$ = the difference in corrected nontidal sea level at Boston and Sandwich

$\rho_w$ = the density of sea water

$\rho_a$ = the density of air

$g$ = the acceleration due to gravity

$h$ = the mean depth of Massachusetts Bay

$\ell$ = the distance between Boston and Sandwich along a 330°-150° axis.

$U^2$ = the 330°-150° component of the filtered and squared wind at Boston.

$n_t$ will be in feet and usually on the order of .1 foot. $h$ is approximately 138.6 ft. $\ell$ is approximately 39 nautical miles, or $237,120$ ft. $g$ is 32.2 ft/sec$^2$. $U^2$ is in (knots)$^2$ and must be multiplied by 2.853 to be converted to (ft/sec)$^2$. It must also be multiplied by 2.657 so as to represent the wind over the bay. $\rho_w$ is approximately 1.99 slugs/ft$^3$ and varies little enough with temperature so as to have a negligible effect.

$\rho_a$ cannot be considered constant, however. As can be seen in Table IV, the density of dry air at 0°F is approximately 18% greater than that at 80°F. From equation (8) it is evident that this can mean an 18% difference in wind stress (at the same wind speed). Most investigators have used a constant value for $\rho_a$. In a case such as Keulegan (1953), where his data covers all parts of the year this would increase the scatter in his results for $C_D$. If humidity
effects on $\rho_a$ are also considered there could be even greater differences. Moist air is lighter than dry air ($\rho_{\text{water vapor}} = 0.622 \rho_{\text{dry air}}$) and warm air can hold more moisture than cold air. Thus, the difference in $\rho_a$ for a dry winter day and a humid summer day might be greater than 18%. (Humidity effects, however, are not numerically considered here.)

**TABLE IV: Variation of Air Density with Temperature**

<table>
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<tr>
<th>°F</th>
<th>$\rho_a$ (in slugs/ft$^3$)</th>
</tr>
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<tbody>
<tr>
<td>0</td>
<td>0.00268</td>
</tr>
<tr>
<td>20</td>
<td>0.00257</td>
</tr>
<tr>
<td>40</td>
<td>0.00247</td>
</tr>
<tr>
<td>60</td>
<td>0.00237</td>
</tr>
<tr>
<td>80</td>
<td>0.00228</td>
</tr>
</tbody>
</table>

With the above values, equation (8) reduces to:

$$C_D = 0.00494 \frac{\eta_t}{u^2 \rho_a}$$

The results of using this formula are given in Table III and plotted in Figure 18 versus wind speed. There is evidence for a stability effect on $C_D$. Not only does the mean value for $C_D$ increase with decreasing stability, but all the stable data points are below the lowest unstable data point. There also seems to be a decreasing scatter with increasing wind speed (although there are too few data points to really tell).
\[ C_D \text{ vs WIND SPEED} \]

\begin{figure}
\centering
\includegraphics[width=\textwidth]{figure18.png}
\caption{Wind stress coefficient versus wind speed for the three stability groups.}
\end{figure}
The mean value of $C_D$ for all the data is $1.50 \times 10^{-3}$. There have been so many investigations concerned with determining $C_D$ that there is no problem finding someone who has obtained a similar value. (For summaries of many of these results see Wilson (1960), Deacon and Webb (1962), Neuman (1956), Francis (1954), and Wu (1969).) However, this value falls roughly in the center of the range and is close to those values chosen by most reviewers as being most probably correct. More interesting is the fact that this value of $C_D$ is close to values gotten by wind profile and eddy correlation methods. Many authors have objected to the wind set-up method for low wind speeds, and indeed most previous results have been a good deal greater than the results of other methods. (See Deacon (1957) and Neumann (1956, 1959).)

Comparison of the results of this study with the results of DeLeonibus (1971), which is probably the best investigation to date concerning stability, yielded a surprising if not a purely coincidental similarity. The unstable results for both studies are plotted in Figure 19. Although the magnitudes of $C_D$ are considerably different, the shapes of the variation of $C_D$ with wind speed are similar. Considering that two completely different methods were used and that a number of assumptions were made along the way this similarity is surprising. The results for neutral and stable conditions were not similar, however.
Unstable Results Compared With Those of DeLeonibus (1971)

This study (Wind speed measured at 10 m)
DeLeonibus (Wind speed measured at 7.5 m)

FIGURE 19: Unstable wind stress coefficient versus wind speed, for this study and DeLeonibus (1971).
The lowest of the stable data points in Figure 18 were from the end of a long stable period in July. The possibility is only mentioned that humidity may be playing a part here, either in decreasing $\rho_a$ (which if not taken into account would lead to a too small $C_D$) or in adding to the stability. The affect of a humidity profile on atmospheric stratification has yet to be studied to any great extent. (But see Monin, 1967.).

4.5 At the end of Section 4.2 the assumptions made in using this simple one-dimensional approach were listed. They will now be discussed as to their degree of validity. Also, several objections raised by various authors about the wind set-up method will also be discussed. Although one or more of these assumptions or objections may have affected the magnitude of calculated $C_D$'s, it is believed that since the same method was used to obtain all values, the results concerning stability are still valid.

(1) The basic one-dimensional assumption of course is an approximation. Ignoring motion in the $y$-direction essentially means ignoring any Coriolis effect on the difference in sea level at Boston and Sandwich. Only during times of strong winds from the east would this seem to have a significant adverse effect on the results since the higher wind speeds and greater fetch would make the Rossby number a good deal smaller. In this case there will be a tendency for the water to move to the right, thus raising the sea level at
Boston more than at Sandwich. There were a couple of cases where the correlation was poor between sea level difference and the 330°-150° component of wind stress. And coincidentally these occurred during times of strong easterly winds. For example, notice the poor correlation between Boston-Sandwich sea level differences and the 330°-150° component of wind stress on February 8th in Figure 11b, and then notice the 240°-60° component of wind stress in Figure 10b.

(2) It has frequently been questioned as to how often the steady state requirement is actually met in the real physical situation. (Deacon and Webb, 1962; Francis, 1954.) All data for this investigation have been filtered, however, and low frequency changes in sea level and wind (often spanning a couple of days) are most likely to reach an equilibrium condition.

(3) The assumption that the bottom stress, $\tau_b$, is negligible due to the small return current has been verified experimentally by a number of investigators (Van Dorn, 1953; Francis, 1953; Baines and Knapp, 1965) for depths much less than Massachusetts Bay. The deeper the body of water the more negligible $\tau_b$ becomes.

(4) The open east sea boundary of Massachusetts Bay makes the continuity equation used before only an approximation. Since it is movement along the longitudinal axis of the bay that is of concern here, it is hoped that the
open boundary will have little effect. Movement in this direction should not appreciably affect the flow of water into and out of the bay. The open boundary of course also allows for the effect discussed in paragraph (1). There is also the question as to how much the northern arm of Cape Cod shelters Sandwich from water movement due to a $240^\circ$-$60^\circ$ component of wind stress. Considering the low frequency of the data it is doubtful that there is an appreciable adverse effect. Considering the shallowness of the lower half of Cape Cod bay, the opposite might be expected anyway, i.e. more of a set-up at Sandwich due to $240^\circ$-$60^\circ$ wind stress than at Boston.

Objections raised by various authors concerning the wind set-up method include the following:

(5) Wind also causes surges and seiche movement which may mask wind set-ups (Francis, 1954; Deacon and Webb, 1962). However, the filtering of the data in this investigation should eliminate that problem.

(6) Most wind set-up studies for large bodies of water have had to rely on wind data from onshore weather stations, so that measured wind speeds were lower than actual wind speeds over the water, due to the frictional effects of the land. (Deacon, 1961.) Such was also the case for this study, since Boston wind data had to be used. However, the frictional effects of the land were demonstrated and a correction factor was calculated so that the Boston data might approximately represent wind over the bay.
(7) Since tide gauges are in relatively shallow water there can be a definite influence on water level due to either "wave set-up" (as a result of breaking) or "wave set-down" (a consequence of increased radiation stress and thus decreased hydrostatic stress when waves run into shallow water and do not break).

(Stewart, 1961; Longuet-Higgins and Stewart, 1964). Since the Boston Harbor gauge, at Pier No. 5, and the Sandwich gauge just inside the entrance of the Cape Cod Canal, are both in at least 40 feet of water, it is questionable how important this effect is. It is also possible that the filtering of the tide data may reduce whatever effect there might be. Stewart's view is that the water must be deep enough so that the phase speed of the waves is not significantly affected by the bottom. This would be more of a problem for high wind data than for the low wind data of this investigation.

(8) Ursell (1956) felt that if the depth of the basin were not uniform the wind driven vertical circulation produced a non-negligible effect. He felt that the momentum change due to a change in water velocity, \( \frac{du}{dx} \), could not be neglected. In the derivation in this study, however, this term was not simply neglected, it dropped out due to continuity. Since continuity is only an assumption for this study, it is pertinent to mention that when Hidy and Plate (1966) calculated the necessary correction from experimental
data in their wind-wave tunnel the modification for \( \frac{du}{dx} \) amounted to less than 1% of the total shearing stress. (Of course a wind-wave tunnel has a more uniform depth than Massachusetts Bay, so the result would not be as small for the bay.)

(9) The departure from homogeneity of the water may also affect the results of a wind set-up study. According to Deacon and Webb (1962) the density differences associated with a horizontal temperature gradient can cause a slope of the water surface (e.g. a 6°C difference over 40km of a 20m deep body of water can lead to a 1 cm (.03 ft.) difference in sea level). Horizontal temperature gradients do not seem great enough in Massachusetts Bay, however, for this to have a serious effect. Francis (1954) is more concerned with the effects of vertical thermal stratification. While it is true that wind-driven water movements are more confined to the upper layers, this does not mean that the depth used in equation (7) should be the depth of the thermocline (This would have the same effect as increasing the heretofore negligible value for \( \tau_b \); which if included in equation (7) would have been added to \( \tau_s \), instead of subtracted, because the return current flows in the negative x-direction.). However, it does seem possible that the thermocline might absorb enough momentum and energy from the surface to affect wave growth on the surface, which might in turn affect the wind stress.
(10) One last item to consider, although the effect is probably negligible, is the fact that the anemometers are stationary while the water surface moves up and down about nine feet over a tidal cycle. Thus, different parts of the wind profile are being measured at different times. At this distance above the surface this probably does not mean much difference in wind speeds. Even if it did, the wind data was filtered for tidal frequencies.

There are certainly many ways in which the results of this investigation might have been adversely affected. But the fact that the mean value calculated for $C_D$ was in the accepted range and close to values obtained by other methods (e.g. wind profile and eddy correlation methods), implies that such effects were of small importance. More important, the demonstration of the effect of atmospheric stability is still valid, since the same method was used to calculate all $C_D$'s.
5. COMMENTS ON WIND DRIFT
AND CONCLUSIONS

5.1 As mentioned in Chapter 1, for a body of water the size of Massachusetts Bay the drift current due to wind is on the same order of magnitude as the tidal current. Thus, wind stress cannot be ignored in any mathematical model that is to describe transport mechanisms well enough to be predictive. Since the strongest wind-driven currents are in the uppermost water layer, wind stress would be especially important when considering the dispersion of buoyant pollutants such as hot water discharge from nuclear power plants, or the drift of surface material such as oil slicks.

In the previous chapter values of the wind stress coefficient were found for different atmospheric stabilities. It was also postulated that due to greater air density, usually greater atmospheric instability, and usually greater wind speeds, the wind stress in winter was on the average much greater than in the summer. (The 1971 data was in agreement with this.) This does not mean, however, that one should necessarily expect greater wind drift currents (in the upper layers) in the winter. One must also consider the stratification of the body of water. Massachusetts Bay, for example, is very stratified during the summer and has a distinct thermocline. One would expect such stratification to tend to confine more of the momentum received from the wind in the upper layers. The result should be a smaller amount of water (i.e. that above the thermocline) moving
faster. If such is true summer wind drift currents could be greater than expected.

This is exactly the result found by Doebler (1966). Using Lightship current and wind data off Delaware and North Carolina he found greater summer currents than winter currents even though summer wind stress was less. Gonella (1971), using data collected over several years on board the Bouée Laboratoire in the Mediterranean Sea, has also gotten the same result and thus demonstrated what he refers to as "the screen effect of the thermocline on the momentum flux."

No reliable winter current data is available for Massachusetts Bay. The National Ocean Survey spent the summers of 1971 and 1972 in the area between Boston and Cape Ann. Most of the current data and STD data from those years were acquired for this study. Although there is no winter data with which to compare, so as to determine the effects of stratification on drift currents, it was felt that some idea might be gotten by looking at current stations near the coast, where frictional effects might destroy the thermocline. Temperature profiles collected by Bumpus (1974) for Massachusetts Bay did seem to confirm that the thermocline does break down at the edges of the bay, the water being more homogeneous there.

N.O.S. current stations having three or more depths were analyzed. A mean nontidal drift was determined for each depth of each station, representing the entire duration of the station (usually seven days). Fourteen stations off
Salem Harbor are shown in Figure 20. The nontidal drifts are given in Table V. It will be noticed that for stations closer to shore (e.g. Stations 6, 11, 17, and 19) the current speeds for the three depths are closer in magnitude than the speeds for the three depths at stations further out. It will be noticed that for these outer stations (e.g. Stations 12, 13, 14, 16, 18, and 20) there is a sudden drop in current magnitude between the 25 foot depth and the bottom depth. It will also be noticed that with most of these outer stations the 25 foot depth shows a stronger current than the 15 foot depth. Since the average depth of the thermocline is around 25 or 30 feet, it is felt that these higher speeds were probably due to internal waves, which are known to exist in Massachusetts Bay (Halpern, 1971). Generally, the data from these fourteen current stations do seem to indicate that stratified conditions in Massachusetts Bay result in greater wind-driven currents in the upper layers and smaller currents in the lower layers.

5.2 The main result of this investigation has been the demonstration of a definite atmospheric stability effect on the wind stress coefficient (or drag coefficient), $C_D$, and some indication of the magnitude of this effect (for wind measured at 10 meters above sea level). The mean values of $C_D$ for the three stability groups were:

- $1.10 \times 10^{-3}$ for stable conditions,
- $1.49 \times 10^{-3}$ for neutral conditions,
- $1.84 \times 10^{-3}$ for unstable conditions.
FIGURE 20: N.O.S. Current Data Off Salem Harbor (Station Locations)

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Locations of current stations are shown in Figure 20.
The mean value for the whole set of data, $1.50 \times 10^{-3}$, is within the range of values presently considered acceptable by most authors. It is also lower than most wind set-up results for low winds and closer to the results of other methods (e.g. wind profile and eddy correlation methods). This is not to imply that these latter methods should be considered the standards of comparison (as many would have one believe). In fact, an objection is raised as to their validity in Appendix A, and other objections by several authors are referenced. There have been numerous objections to the wind set-up method, and all of these have been fairly well answered for as far as this particular study is concerned. The filtering of tide, wind, and pressure data has solved many problems.

Of course the determination of $C_D$ was only possible because the correlation between Boston-Sandwich sea level differences and the component of wind stress along the longitudinal axis of Massachusetts Bay was very good. Even the correlation between Boston-Portsmouth sea level differences and the onshore component of wind stress was surprisingly good. It also became apparent from this study that the influence of the Gulf of Maine on Massachusetts Bay should not be underestimated. The nontidal sea level curves for the four tide stations ranging from Sandwich, Massachusetts to Eastport, Maine were very similar (even after taking barometric pressure into consideration) and the several hour lag between wind changes and sea level changes implies a large fetch.
For 1971 it was found that wind stress was generally much greater in the winter than in the summer. This was not only a result of greater wind speeds, but also greater atmospheric instability and colder, denser air (the latter factor has been overlooked by many investigators). This greater wind stress in winter has caused greater nontidal sea level changes in winter (the maximum range was about two feet for 1971), but there is reason to expect greater wind-driven currents (in the upper layers) in the summer due to thermal stratification.

Some basic observational evidence for the rotation of the sea breeze was also presented, but no explanation for the counterclockwise rotation at the stations on lower Cape Cod was attempted. The frictional effect of land on wind speeds was also demonstrated and was corrected for in the wind set-up technique. This has apparently not been done by many investigators using this method and is perhaps one reason why the results of wind set-up methods have often been higher than the results of other methods.

A sceptical look has also been taken at the basic quadratic law (in Appendix A) and an alternative approach suggested.

5.3 This study, having touched on many subjects, can be continued in many directions. First, it is worth repeating this study on several more years of data (which should be available, depending on how long the Sandwich tide gauge has been running; at least two and half more years of data exist
and possibly more; and hopefully a few large storms might have occurred). Perhaps with more data the shape of the $C_D$ versus wind speed curves could be better defined for each stability.

It would also be worthwhile acquiring a year's worth of reliable current data somewhere in the bay, with sensors at several depths. Among other benefits this would give some idea of the effect of seasonal stratification on the wind driven circulation of the bay.

Though there are technical difficulties to be sure, a method of determining $C_D$ from current measurements should be developed. (e.g. Gonella, 1971, uses a current profile method.)

The ideas in Appendix A may also lead to some interesting results if pursued.
G. I. Taylor proposed in 1916 that the quadratic law, which had been used for fluid flow through a pipe, could also be applied to air flow over land. He also originated what is now referred to as the "wind profile method," when he used observations of wind speed at different heights above the ground to calculate the drag coefficient. The drag coefficient proved to be constant for similar roughnesses and Reynold's numbers, and the quadratic law worked well in describing quantitatively the transfer of momentum from the atmosphere to the land.

The quadratic law was later applied to air flow over a water surface with the assumption that, relative to air, water was approximately solid. It soon became apparent, however, that the drag coefficient for the air-sea situation was not a constant. Even above the experimental scatter of the numerous techniques developed to calculate $C_D$ (all of which can be objected to for a number of reasons; see Kraus, 1967, and Stewart, 1961), there was a definite variability in the results. At this point some investigators looked for a replacement for the $U^2$ law ($U =$ wind speed), some favoring a $U^3$ law, and others something in between. Most kept the basic $U^2$ law and simply allowed their "constant," $C_D$, to be a function of $U$. There is still such a lack of agreement on this subject that the big argument still seems to be whether $C_D$
decreases or increases with $U$.

There are of course several important differences between the air-sea situation and the air-land situation. Whereas the land, being solid, does not move (noticeably) as a result of receiving momentum from the air, the sea, not being solid, does move as a result of having momentum imparted to it. And it is in fact the movement of the water, not the air, that we are trying to describe when we put a wind stress term in the equations of motion. (It seems that part of the problem with the theories and experimental "laws" used in air-sea interaction, is the fact that most were actually derived to describe the movement of the less dense material at an interface.)

Equally as important is the fact that the interface itself moves and changes shape constantly. And a great deal of the momentum from the air goes into changing the shape of this interface, i.e. into generating and maintaining waves. (See Stewart, 1967, 1973.)

The idea of "roughness" is also not well defined for the air-sea situation. Some investigators have assumed that the waves created by the wind could be considered roughness. They then admitted that a problem existed in that roughness was a variable, but since they could approximately predict sea state according to wind speed this became their justification for trying to correlate $C_D$ with wind speed. The great range of roughness lengths from a smooth sea to a storm sea
apparently did not bother them, perhaps because the quadratic law seemed to work well on land with roughnesses from grass fields to city buildings (which is not necessarily dynamically similar). When looking at the air-sea interface with respect to its effect on the air layer, it is not unreasonable to consider the waves as roughness. But when looking at the effect on the water layer, we cannot avoid looking at motions on the same scale as this so-called "roughness," i.e. at the wave motion (and the resulting Stoke's transport). It would therefore seem more logical to keep "roughness" on a small scale (i.e. that of smooth water, or perhaps up to and including capillary waves) no matter what the sea state, and then to look at how the waves may change the "effectiveness" of this stress. (A clearer meaning of this last phrase will be attempted shortly.) In addition, as mentioned by Kraus (1972), since there is no level at which mean wind speed is zero (the interface itself having a mean drift) the roughness length cannot be associated in any direct way with the mean square surface height.

Perhaps at some stage there should also be some agreement on what we really want the surface stress term in the equations of motion to represent. Should \( \tau_s \) represent all momentum imparted from the air to the water, or perhaps just that resulting in a mean drift of water particles? Apparently, a large amount of the momentum received from the air goes into waves. And what can happen to these waves? They can move along in some "equilibrium" type situation
imparting momentum to the body of water in the form of a Stoke's type transport. They can decay and impart more momentum to the water, or they can grow and impart less. They can reach shore where they may be reflected or they may cause a set-up (if the water is shallow and they break) or a set-down (if the water is shallow and they don't break). Or they can leave the area all together and head out to sea. In all these cases, how much of the wave's momentum is received by the body of water in a form that might contribute to currents or wind tides? (And for example, how much momentum is carried by storm surges, which can travel for hundreds of miles?). Is it reasonable to treat the problem as though all the momentum received from the air can be included in the equations of motion? And are methods that calculate $C_D$ using only parameters measured in the atmospheric layer really legitimate?

Consider the very simple wave profile shown in Figure 21. As first suggested by Jeffreys (1925) (and as used off and on in other theories over the years, e.g. Munk (1955)), there can be a separation of the air flow at the crests of developed waves, if they are steep enough (the question is, how steep is that). In such a case only part of the wave profile "feels" the direct wind. Aside from the possible flow separation, the slope of the wave affects how much of the momentum from the air is received in the form of a pressure and how much as a true shear stress. Continuing this very basic geometric approach, it will also be seen from
Figure 21, that part of the true shear stress will be in the horizontal direction and part in the vertical. The same will be true of the pressures. It is assumed that horizontal momentum imparted to the body of water as a mean drift will come only from the horizontal component of the true shear stress and the Stoke's transport of the waves.

With this in mind consider the following sequence of events. Beginning with a smooth water surface, the wind blows and imparts momentum to the body of water predominantly in the form of a shear stress. Eventually waves will begin to grow (by whatever mechanism, Phillips or Miles, or a combination thereof). As wave slopes slowly increase, less of the momentum will be transmitted in the form of a shear stress and less of this shear stress will be in the horizontal direction. At the same time, however, as the waves grow, Stoke's transport will slowly increase. Also, at some point
the waves may become steep enough for a sheltering effect to take place due to air flow separation at the crests. This will further reduce the momentum transmitted in the form of a shear stress. Thus, as the waves grow there will be three factors affecting the amount of horizontal momentum imparted to the body of water as a mean drift: (1) the decreasing horizontal shear stress (due to increasing wave slope); (2) the increasing Stoke's transport (increasing with increasing wave slope); and (3) the degree of sheltering (also a function of wave slope). (Only the actual true shear stress being felt by small portions of the water surface will obey the quadratic law, and in this case the $C_D$ would be a constant). The three factors just listed have a dependence on wave slope (or wave steepness), which in turn depends on the wind in some complicated way, usually described spectrally and/or statistically.

If all of the above is approximately true, and for some reason we decide to still use a quadratic law to describe the entire process, what would this imply about the $C_D$ used? For one thing it is unlikely to be a simple function of wind speed. It is even possible that it might decrease and then increase at some point (if plotted versus wind speed), depending on the relative varying influence of the three factors described above. (This is not said as justification of the shape of the curves in Figure 19.) It is also likely that there would be a lower bound to $C_D$, corresponding to a smooth water surface, and an upper bound, corresponding to a
"fully developed" sea state. (Of course, if wind speeds got so high that waves were literally torn apart into spray, a different situation resembling sand transport might result. This of course is not considered in postulating an upper bound for $C_D$).

A few additional comments should also be made. There does seem to be evidence that Stoke's transport is a component of the total drift and that it does increase with wave growth. It is also very possible that it accounts for most of the drift under "fully developed" sea conditions. (See Kenyon, 1969, and Bye, 1967.)

It is, as of the present time, undecided as to whether capillary waves should be considered as part of the roughness or as having some small slope effect.

Relating to an earlier comment about methods using only atmospheric parameters to determine values for $C_D$, it is felt that some technique to determine $C_D$ (or some other value in some other type of equation) must be developed based on measuring drift current.

It is of course possible that, after combining the geometric considerations described here, some accurate Stoke's transport theory, and some reliable spectral or statistical wave prediction theory, we will not have a better quantitative description of wind stress than the quadratic law with a carefully calculated $C_D$. But somewhere along the way some valuable insight might be gained into air-sea interaction processes.
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