Wind Forcing of the Atlantic Thermocline along 32°N at Low Frequencies

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ABSTRACT

The Bermuda tide gauge record extends back to the early 1930s. That sea level fluctuations there are highly coherent with dynamic height from hydrographic data has two interesting implications. First, it should contain information about the low-frequency circulation of the Atlantic. Furthermore, because dynamic height contains information on heat storage, it might, on the limited timescales accessible in the record, also contain clues about climate.

A simple model of wind forcing of the Atlantic from the African coast to Bermuda uses the Levitus mean density data to estimate the long Rossby wave speed as a function of longitude. Sea level and thermocline variability estimated this way are in remarkably good agreement with observations at periods of more than a few years duration. The peak-to-peak sea level signal is ~18 cm, which is nearly 25% of the slope across the Gulf Stream at this latitude. The model results suggest that the variability is largest somewhat to the east of Bermuda; fluctuations of ~10 cm extend as far east as ~35°W.

One surprising result is that at the longest periods in the COADS data, the wind curl has a double-peak structure in longitude. That is, there is a significant amount of power on the eastern side of the ocean as well as near Bermuda. Therefore, it is essential to use the full horizontal resolution of the wind data; using the mean curl across the Atlantic turns out not to be a good way to estimate thermocline variability. One might wonder if the wind data are reliable at these long periods were it not for the good agreement between the results and observed sea level. The power in wind variability increases out to ~500 months, although with little statistical reliability. Sea level variability however, appears to peak at somewhat shorter periods. Although it is pushing the resolution of the data, this result is a limitation imposed by the basin width scale. The power in the *model* ocean's response to wind forcing is nearly an order of magnitude larger during the first half of the record (1952–69) than during the second (1970–86).

It is likely that significant changes in buoyancy forcing by the atmosphere are coherent with changes in wind. Nevertheless, these results suggest that the variability in sea level—and so in deep temperature—can perhaps be accounted for without invoking changes in stored heat of the deep ocean.

1. Introduction

It is becoming increasingly clear that our understanding of the ocean has progressed beyond the point of trying to estimate the "mean circulation." We now have data of sufficiently long duration to make it possible to inquire into the low-frequency variations of quantities such as the transport of boundary currents and the shape of the main thermocline. Our interest here was also sparked by an attempt to understand the apparent long-term rise of sea level on the U.S. Atlantic coast, a signal badly contaminated by low-frequency variability.

The long record at Bermuda seemed a good starting point for understanding these low-frequency signals. Many people have studied the tide gauge data at Bermuda and the hydrographic observations nearby. It was shown by Roemmich (1990), following the earlier work of Schroeder and Stommel (1969) and Wunsch (1972), that dynamic height compares very well with tide gauge data at periods as long as the eddy band. The energy in these fluctuations peaks near 6–8 months. There are also fluctuations at much longer periods, however; at roughly 30–200 mo, the variability in sea level makes it difficult to determine the rate of rise of sea level accurately, both at Bermuda and along the east coast of the United States. Therefore, it was our hope to understand the wind forcing in this frequency band in order to reduce the effective noise level in the signal associated with sea level rise. Furthermore, since sea level differences are a measure of velocity in the ocean, we hoped to learn about the low-frequency variability in the Gulf Stream by studying the signals at Bermuda and the U.S. east coast. In this paper, we concentrate on results between the African coast and Bermuda, postponing the more complex issue of Gulf Stream variability.

The frequency range available is limited by the duration of data in wind and sea level, but it seemed worthwhile to investigate the range of periods out to

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approximately 100–200 mo. Though fluctuations of this duration can barely be resolved, they are important because there is a substantial amount of power in that band.

Here we examine the relationship between sea level at Bermuda and wind forcing across the Atlantic. To do this we use a linear, continuously stratified model separated into vertical modes (e.g., see Gill and Clarke 1974; Picaut and Sombardier 1993). Our calculations, other than in the use of continuous vertical stratification, are similar to those of Kessler (1990), who had very good results with an analogous calculation in the Pacific.

Though it is widely believed that sea level fluctuations at low frequencies are driven primarily by the winds, there is ample reason to suspect that forcing by variations in other parameters could be large enough to mask wind effects. Climate-scale heating of the deeper waters has been reported by Roemmich and Wunsch (1985) and by Levitus (1992). Substantial changes in salinity or ice coverage have been documented as well (e.g., see Chapman and Walsh 1993; Dickson et al. 1988; Levitus 1989, 1990; Manning 1991). And studies of climate changes and the attendant effects on sea level, such as the recent papers by Gordon et al. (1992), have appeared in almost every journal in recent years. The recent work of Deser and Blackmon (1993) has shown, as one might have suspected, that large-scale changes in atmospheric circulation over the Atlantic are associated with decade-scale variations in sea surface temperature.

Low-frequency energy in the North Atlantic has been shown to propagate as long Rossby waves; Price and Magaard (1986) found baroclinic Rossby waves at periods substantially longer than annual. Rossby waves have also been found in studies of the Panulirus data at Bermuda by Frankignoul (1981).

We chose here to begin by studying the effects of wind. We expected to find some degree of correlation of sea level variability at Bermuda to wind; presumably the remaining signal would be the result of other physical factors that could be addressed later. What we had not anticipated, however, was that the coherence between wind and sea level would be so high. Though many previous studies have shown that we might expect to find substantial signals above the main thermocline, the indications have been that the buoyancy, or thermohaline signal, was substantial. Talley and Raymer (1982) show evidence of substantial changes in temperature and salinity that one would assume are related to buoyancy forcing. The maps of energy levels shown by Price and Magaard (1986) suggest that there is forcing in regions of higher latitude, rather than the relatively "local forcing" considered here, although whether by wind or by eddy formation in the Gulf Stream region near approximately 40°–50°W is not clear. The recent work of Liu (1993a,b) appeared after our work was completed; it is a more general approach to this problem. The horizontal wind forcing we observe is significantly different from the pattern assumed by Liu (1993b), but the general relevance remains.

2. Data

The tide gauge data at Bermuda were obtained from the National Ocean Survey (NOS) of NOAA; we have used monthly mean values. Sea level has been adjusted to constant atmospheric pressure by the inverted barometer effect. Sea level data are available at most major U.S. gauges since the 1920s. Although sea level data at Bermuda began in 1932, there is a break from 1937–43.

COADS winds (e.g., Slutz et al. 1985) are used for wind stress. Mayer and Weisberg (1993) have given a description of this dataset from an oceanographic perspective. The data are available as mean monthly winds on a 2° by 2° grid. COADS data, in general, are available prior to World War II, but with often scanty observations; we have used only the data after 1945. Wind data at meteorological stations near the coast are available from the NOAA/National Climatic Data Center, Asheville, North Carolina; we have been unable to obtain data prior to 1945 in digitized form.

To determine the curl we used a standard Green's theorem method, integrating the wind components around 4° boxes, beginning at Bermuda, along 32°N to the coast of Africa. The point of the integration was to reduce the noise while maintaining reasonable resolution. The wind curl was then low-pass filtered to remove power at periods shorter than 24 months and to pass power at periods longer than 60 months.

The speeds of long Rossby waves for our model were calculated for the baroclinic modes. Temperature and salinity data are available from the National Oceanographic Data Center in several forms: on a CD-ROM disc (NODC 1992) and in an averaged dataset made available from S. Levitus, which we used. The time series of long-term hydrographic data near Bermuda (Anonymous 1988), sometimes called the Panulirus data, is also available in the format of the "Live Atlas" of Luyten and Stommel (1988). The tide gauge data (and all data used here) are available for use by others via an "anonymous ftp" account, over Internet, on atlantic.ocean.fsu.edu.

3. Method

A model of wind forcing

We concentrate here on forcing at periods longer than annual, at which the response will be primarily from long, nondispersive, baroclinic Rossby waves. Price and Magaard (1986) found that waves at these long periods traveled within a few degrees of due westward. Krauss and Wuebber (1982) concluded that first-mode Rossby waves would travel nearly westward at the annual period, and our interest is in periods substantially longer. The results of Liu (1993b), however, suggest that, even at decadal scales, there would be a significant amount of N–S wave propagation. The distribution of wind forcing, however [as shown by Mayer and Weisberg (1993)], is much more limited meridionally than Liu assumed; the region of strong forcing extends only to 36° or 38°N. Because the forcing at higher latitudes is weak, it is unlikely to propagate into the region of interest. Moreover, the region of interest has a weak mean westward flow; Chang and Philander (1989) have shown that westward flows inhibit meridional probagation. As we concentrate on the ocean at 32°, we restrict our attention to E–W propagation alone.

After separation of the linearized equation for a stratified ocean into vertical modes, one finds that the low-frequency, large-scale wind-driven response is described by

$$\frac{\partial p_n}{\partial t} - \mathrm{Cr}_n \frac{\partial p_n}{\partial x} = B_n \,\mathrm{curl}(\tau/\rho_0 f), \qquad (1)$$

where

$$B_n = -\frac{c_n^2}{f^2} b_n$$

$$b_n = \int_{-H_{\mathrm{mix}}}^{0} F_n dz / H_{\mathrm{mix}} \int_{-H}^{0} F_n^2 \mathrm{d}z.$$

Here, p is pressure, t time, x the horizontal coordinate (positive to the east), τ wind stress, f the Coriolis parameter, H the total depth of the ocean, H_{mix} the depth of the surface mixed layer, and ρ_0 is a constant representative water density: Cr_n is the speed of mode n long Rossby waves, given by

$$\mathrm{Cr}_n = \beta c_n^2 / f^2. \qquad (2)$$

For each mode c_n, we calculated the long gravity wave speed from the mean hydrographic data of Levitus at 32°N. The long Rossby wave speed of each mode along 32°N is calculated from Eq. (2). The wave speed in our model is allowed to vary slowly with longitude, in keeping with the observed density distribution. We have not fine tuned the wave speeds for an optimum fit in the results. For the first mode, C_R increases from ~2 cm s^{-1} off Africa to ~2.3 cm s^{-1} near Bermuda; H_{mix} was taken to be 100 m. We show our primary results using the first-mode calculation; results from using higher modes are described in section 5.

One of our concerns was the choice of depth in the region of irregular bathymetry near the Mid-Atlantic Ridge; we, therefore, used the depth to which the mean density data extended in that region. However, Picaut and Sombardier (1993) have shown that for the first few modes it is more important to give an accurate specification of stratification than of depth. Moreover, the main thermocline is much deeper near Bermuda than in the low-latitude Pacific, where they worked, so the standard levels in the Levitus data give proportionally better resolution.

4. Results

a. Distribution of wind forcing across the North Atlantic

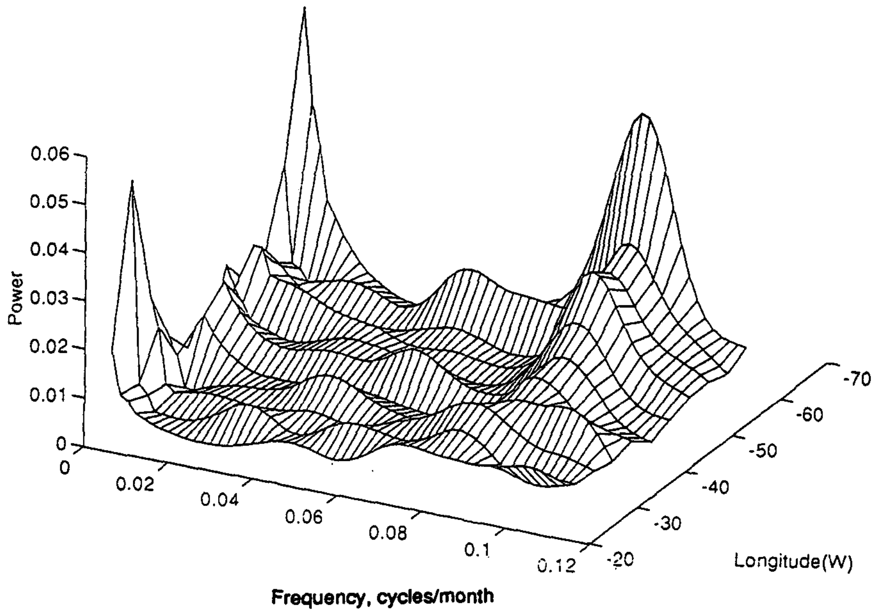
Figure 1 shows the spectral content of the wind curl along 32°N from the COADS data. Although we forced the model only with low-passed values, we show the unfiltered values in order to make comparison with the annual variability more readily apparent. The upper part of the figure shows a perspective view, with the annual term removed; the power in the annual signal is shown below.

The data were first multiplied by a 50% cosine bell taper, or "data window," to reduce leakage. The raw spectral estimates were then smoothed by five Hanning passes, decreasing gradually at the lower frequencies until the raw periodogram values are shown at the longest period, 512 months. This unusual way to show spectra is necessary in order to show the power at the longest periods; it should be remembered that the model is forced with (low passed) data that have not been "smoothed" in frequency (as is the ocean). Note that the scales in Fig. 1 (upper part) are linear, so this is equivalent to a variance-preserving plot.

At the longest periods available in the data, Fig. 1 shows that the power in curl has maxima on opposite sides of the ocean. This feature appears to be a new result. The mean curl also has similar maxima. At frequencies close to annual, however, the power shows a single maximum, suggestive of a first-EOF mode, as described earlier by Ehret and O'Brien (1989).

There is a conspicuous amount of power in Fig. 1 near the annual period. It is important to recognize that this spread is an authentic feature of the wind forcing; in some years the strongest winds are in December and in other years in February, etc. As is appropriate for such calculations, the power at exactly 12.0 months was removed from the raw periodogram before smoothing and does not appear on this diagram. (The identical result can be obtained by removing the annual cycle explicitly before taking the FFT.) The large amount of power that remains shows that the near-annual power is spread over a wide frequency band. The annual power, as shown in the lower part of Fig. 1, has a single maximum near Bermuda. This maximum, at a period of 12.0 mo, is larger by a factor of ~4 than the maximum at the longest period, 512 mo. However, since the spectrum is red, the large value at 512 mo does not represent a peak, so the comparison may be misleading. That is, the power could continue to increase toward lower frequencies but we cannot see it in the data.

There is some evidence, however, that the sea level power may level off at shorter periods, as shown in Fig. 2. The record at Bermuda, begun in 1932, was inter-



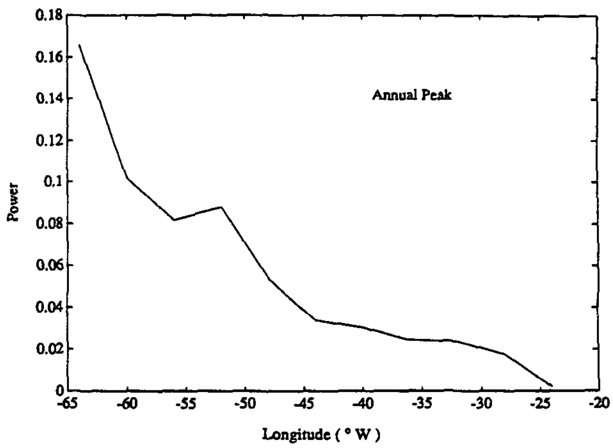


FIG. 1. Spectrum of wind curl along 32°N, computed from integrating the COADS wind stress components around 4° boxes. The axis that goes into the page is longitude, as viewed from the coast of Africa. The lowest frequency is 1/(512 mo). Upper panel shows the power smoothed (in frequency) by five Hanning passes, with the amount of smoothing decreasing to lower frequencies (see text). A 50% cosine bell data window was used to reduce leakage; the power at 12.0 mo was removed before smoothing. Lower panel shows power from the raw periodogram at (12.0 mo)$^{-1}$, also computed from 512 months of data.

rupted between 1937–43. The spectrum of this gappy record cannot be computed by ordinary methods. The spectrum of the continuous part of the record is shown (solid curve) in Fig. 2. Using the technique described by Sturges (1992), however, we find (the dotted part at the lowest frequencies) that the power goes down in the lowest resolvable frequency bands ($\sim$1/750 mo). It will be argued in the next section on physical grounds that this decrease is to be expected. Thus, we may speculate that the Bermuda sea level spectrum may not be red at frequencies immediately lower than this.

It also should be clear that variability in wind curl at periods of $\sim$500 mo has no statistical reliability, in the usual sense. These long-period variations could be completely random. Nevertheless, if these fluctuations occur in the winds, then that is the forcing of the ocean. The statistical reliability may be moot, yet the physical connection between forcing and response is based on a priori physical principles, not on a relationship inferred from a statistical result.

b. *Model results*

The primary question we address is whether sea level computed from wind forcing compares favorably with observations. To help understand the computed sea level from our model, we first show some results from forcing by simple wind fields. Figure 3 compares observed sea level with results of two calculations, first by forcing the model with the mean wind curl across the Atlantic (solid curve), and second, using only the local Ekman pumping at Bermuda (dashed curve). Clearly, both forcing terms are important; the amplitudes are similar, the appropriate frequencies are present, but obviously neither term, individually, does a very good job.

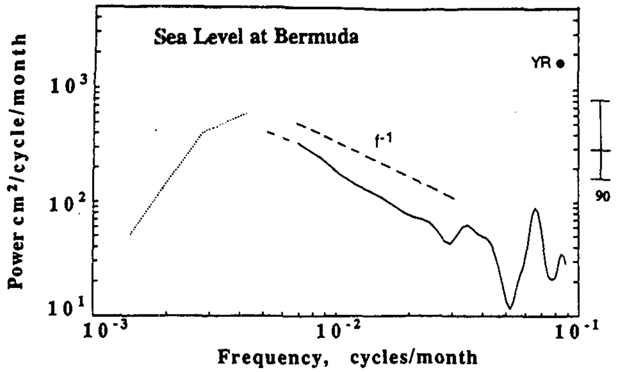


FIG. 2. Spectrum of sea level at Bermuda. Full curve shows a calculation from the continuous portion of data beginning in 1944, smoothed by three Hanning passes, except for the last point shown (dashes). The dotted segment at low frequencies shows an estimate based on a technique for using gappy data (see text) that extend to 1932.

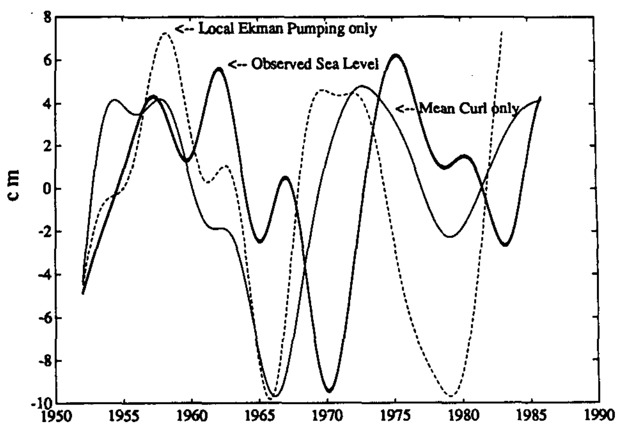


FIG. 3. Comparison of observed sea level at Bermuda (thick solid line) with the results of the model calculations: by forcing with mean curl along 32°N in the Atlantic (thin solid line) and by local Ekman pumping only, near Bermuda (dashed line).

Figure 4 shows a comparison between observed sea level and the model, using the full longitudinal resolution of the wind field and the contribution from only the first vertical mode. The comparison is remarkably good; it is clearly essential to include the full resolution of wind to obtain optimum results. It is appropriate to emphasize that the wave speeds were computed independently of Fig. 4; they were allowed to vary slowly in longitude, as computed from the density field, but were not tuned to optimize the agreement. The matching between thermocline displacement and sea level at Bermuda could perhaps be considered a free parameter, but the resulting value is essentially what one would use in a two-layer model.

It would be appropriate to try to estimate the errors associated with Figs. 3 and 4. One estimate of the "noise" associated with a dataset can be determined from the spectrum, which usually falls to some minimum value (or "noise floor") at high frequencies. This minimum is assumed to be an upper bound on the random noise in the signal; the area associated with that value, times the Nyquist frequency, is the noise variance. That method, using monthly sea level data at Bermuda, suggests a noise of order 3 mm. One suspects that this error, however, is small compared with issues associated with the small number of degrees of freedom. The bulk of the power in Fig. 4 appears to be at periods of roughly 10–15 years, but the frequency spectra of both sea level and of curl are red. We cannot say reliably whether the spectrum continues to rise toward longer periods; this issue is addressed below in another context. Because digitized records for most wind signals are not available prior to 1948, we have restricted our work thus far to the period since that date; work with earlier data is proceeding.

We have compared the model thermocline variations with the displacements of isotherms from 10° to 20°C near Bermuda using the Panulirus data. The agreement with a single isotherm is not quite so good as with the integrated signal, as might be expected.

It is instructive to compare the spectra of the wind curl and observed sea level to see at what frequencies the power is found. The mean curl has a slight bump of power in the 4–6-yr band and is red at longer periods. The spectral decay in power in sea level (from Fig. 2) is $\sim f^{-1.5}$ at periods less than $\sim$200 mo, whereas the decay in wind curl is $\sim f^{-4}$ out to periods as long as can be detected in the data. The point that the spectrum of the mean curl is considerably "more red" than the spectrum of sea level at the lowest frequencies may be an important clue to the ocean's response; we return to this point in the discussion, section 5.

1) HIGHER MODES

The results shown in Fig. 4 are from the solution using only the first mode. Since the sea level fluctua-

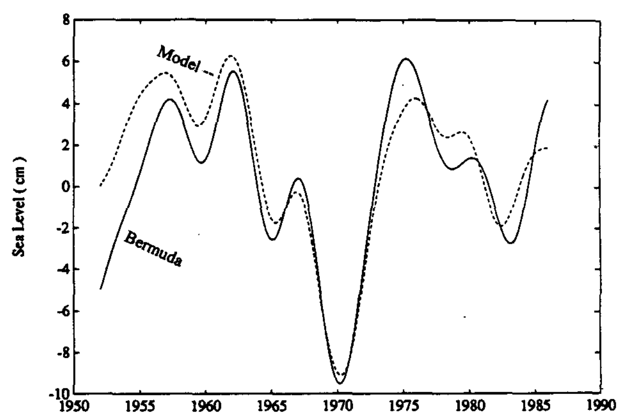


FIG. 4. Comparison between observed and computed sea level, using the full longitudinal resolution of the wind field. Sea level, adjusted for local atmospheric pressure, is shown on the left-hand Y axis; thermocline fluctuations on the right-hand Y axis.

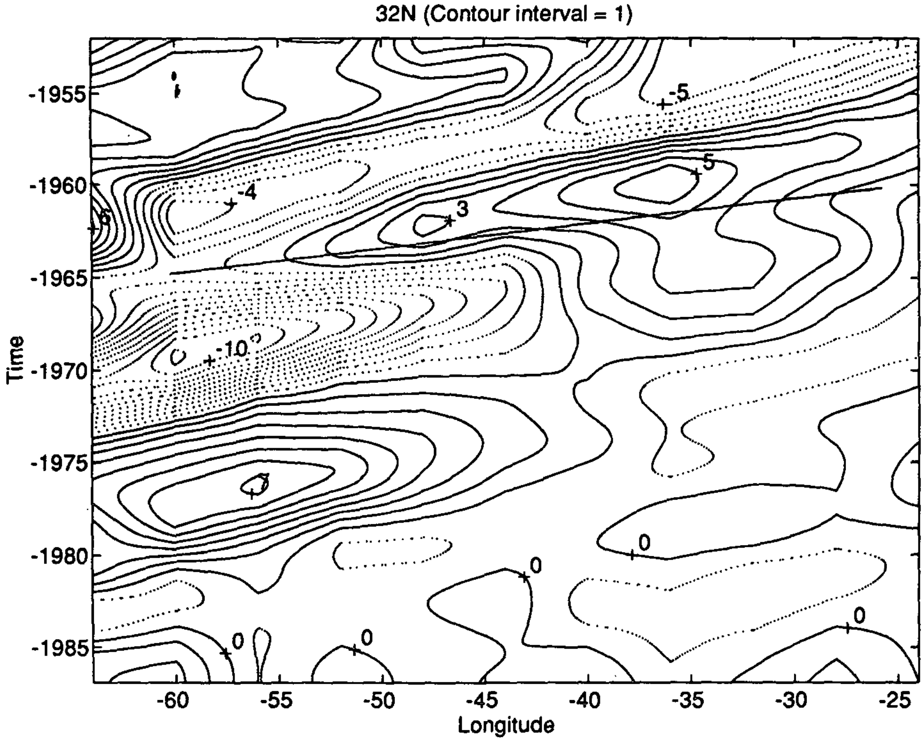


FIG. 5. Model results for sea level fluctuations along 32°N as a function of time. Heights are in centimeters. The diagonal line shows the slope of the mean speed of long Rossby waves in the model, 2.2 cm s^{-1}.

tions are coherent with wind at periods as short as ~3 yr, at these periods one suspects that it may be necessary to include higher modes in order to make better estimates of the wave speeds.

To see whether the higher modes could improve the result shown in Fig. 4, we added vertical modes, first 2, and then 3, to the calculation and computed rms difference values between their relative contributions and observed sea level. Although the addition of mode 2 changed a few details, the agreement did not improve significantly and was worsened by the addition of mode 3.

To test the possible blocking influence of the Mid-Atlantic Ridge, we also tried a calculation in which the second-mode forcing was begun (in the E–W direction) only to the west of the ridge. This calculation did not improve on the result shown in Fig. 4.

It should be noted, however, that these additional calculations involving higher modes cannot be done very accurately because they essentially involve only the first and last terms in Eq. (1). In our calculations we simply used the data available and have not presented any results for the first mode (Fig. 4) before the arrival of the mode-1 wave at the left edge of the domain. Although the second-mode contribution on the African side begins immediately, because the higher modes travel so slowly, it would require an additional ~20 years of wind data (earlier than the data used here) to permit the second term in (1) to become effective at Bermuda at the earliest times of Fig. 4, and those data are not available. For the test with second-mode forcing beginning only at the ridge, it requires only ~10 years for that mode to arrive; as the results did not improve after the arrival of second-mode signal, it is clear that its addition did not help the calculation.

2) RESULTS ACROSS THE WIDTH OF THE OCEAN

The only place for a test such as in Fig. 4 is at Bermuda, yet the model computes the variations at all longitudes; the complete variability in sea level (from the model calculation) is given in Fig. 5. It is interesting to find that the maximum displacement is not found at Bermuda, at the left edge of the diagram, but slightly to the east, between 55° and 60°W. The variability is still surprisingly large as far to the east as ~38°W. The variability on the eastern side of the Atlantic is remarkably reduced in the last half of the interval in the diagram, a point to which we will return. We note that the "beta spiral" cruises were carried out in 1980–81, when the spatial variability was considerably smaller than in previous years.

Figure 6 shows frequency spectra derived from the results in Fig. 5. In this view, the figure is reversed from

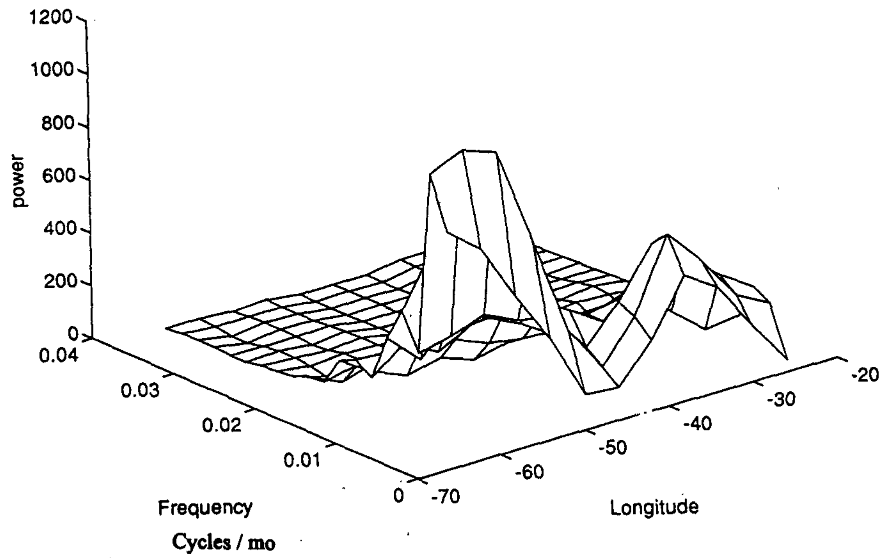


FIG. 6. Frequency spectra, as a function of longitude, of the fluctuations shown in Fig. 5. The lowest frequency is $(512\ \text{mo})^{-1}$; the values were smoothed as in Fig. 1.

that shown in Fig. 1 (for wind), to allow the low-frequency side to be seen more clearly. The remarkable feature here is that the frequency band containing the highest power is not the lowest frequency in the data, as was the case with the wind, but at the next higher frequency, $(256\ \text{mo})^{-1}$. That is, because of the way the ocean responds to the wind forcing, the power in the sea level fluctuations does not mirror the wind input in a one-to-one manner, nor would it be expected. We return to this issue in the section 5 but point out that this result, in the model, is consistent with the observed sea level spectrum of Fig. 2.

To investigate the wind forcing further, we constructed a diagram (not shown) of the wind curl forcing, similar to Fig. 5. Whereas the propagation of Rossby waves is obvious in Fig. 5, none is evident in wind forcing. It is well known that at higher frequencies weather systems propagate from west to east; at these low frequencies there is no consistent sense of propagation in either direction. There are intervals where the phase lines tend toward propagation in one direction for a few years, but these can be of either sign, so that no general pattern emerges.

The effect of higher power in the western part of the ocean, as seen in Fig. 5, is also evident here in Fig. 6. Note that this figure is linear in frequency and in power, so the visual height of the diagram is an accurate indication of the power in the signal. It should also be remembered that the wind power was low-pass filtered to remove power in the region where the spectrum appears so flat in Fig. 6.

3) WAVENUMBER SPECTRUM

Under the further assumption that the fluctuations shown in Fig. 5 have relevance to processes occurring in the ocean, we determined the wavenumber spectrum, shown in Fig. 7. It is computed from each time slice in the results of Fig. 5, after imposing a 50% cosine taper, and then averaged in time. The difference between the 1952–69 and 1970–86 intervals is striking. The duration of these intervals is arbitrary; they are merely the first and last halves of the data. This large difference should give us pause when we try to interpret the results from cruise data as representing anything approaching "mean conditions."

The wavenumber spectra are clearly red. The low power at wavelengths less than 1000 km is to be expected, as the wind forcing was filtered to remove pe-

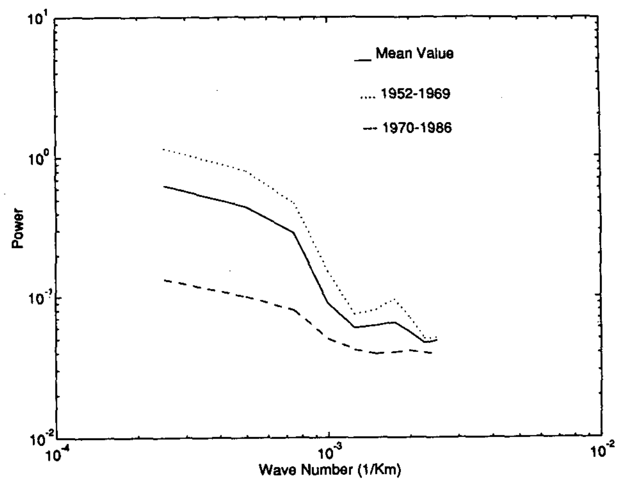


FIG. 7. Wavenumber spectra of the model fluctuations along 32°N. The values were computed along each individual time strip (from Fig. 5) after applying a 50% cosine data window and then averaged over the time intervals shown.

riods shorter than 2 yr. At the longest wavelengths, recall from Fig. 5 that some features have one sign for nearly the full width of the basin. Thus, the wavelengths of the process may be longer than the "width of our array," meaning that the longer waves are poorly resolved. This result, of course, is in keeping with the limitation imposed by the finite width of the basin, as discussed below.

5. Discussion and conclusions

For this study, we have tried to determine to what extent the low-frequency fluctuations in the depth of the thermocline and tide gauge data are forced by large-scale winds over the ocean. We used a simple wind-forced model and found that if we use the full longitudinal resolution of the wind, our model results compare surprisingly well with observations. We must be careful to qualify the interpretation of this result. It does not mean that there are no significant changes in surface heat flux or buoyancy, for the changes in sea surface temperature and salinity are well documented (e.g., see Gordon et al. 1992). Our result implies that the forcing by wind curl is sufficient, in this simple model, to account for the observed variability in thermocline depth. Though one can imagine that we might, for example, have found that the effect of wind forcing was small, making it necessary to invoke sources and sinks of heat or salinity, nothing of the sort was found.

One interesting result is that at the longest periods in the data, the (COADS) wind forcing has a zonal double-peak structure in the vicinity of 32°N. While this feature is clearly contained in the data, it has little statistical significance because it is at the limit of the frequencies available in the data. We conclude that this must be a reliable feature, however, because our results agree with observations only when this aspect of the wind is included. Another feature worth noting is that our good comparison between model and data is obtained with a completely zonal calculation, supporting our initial assumption that the effects of N–S wave propagation at these periods could be neglected.

Although we used a model that employs vertical modes to account for realistic stratification, the entire result is contained in the first mode; nothing was gained by adding modes 2 or 3. However, Fig. 4 clearly shows a less than ideal match; the errors may be related to changes in the heat content of the ocean, to errors in wind data, to some neglected features of our simple model, or some combination of these.

In comparing the scales of the wind forcing with the size of the ocean basin, there is an obvious mismatch. Not only do the wind patterns not propagate to the west, but the peak power is at frequencies that are "too low." A Rossby wave having a period of $\sim$500 mo (as in the wind curl, Fig. 1) at 2.2 cm s^{-1} has a wavelength of $\sim 26 \times 10^3$ km, or five times the width of the ocean. Examining Fig. 5, however, we see that there are events that have a single sign for most of the width of the ocean. If the wind forcing happens, by chance, to be of a consistent sign and phase to cause the response to increase across the full width of the ocean, so that there is a maximum in the west and a zero in the east, this pattern will form a quarter of a wavelength. That is, the period associated with a wavelength four times the distance from Africa to the offshore side of the Gulf Stream can be logically taken as the longest period to be expected from a wavelike response at this latitude. This argument suggests that a period of $\sim$350 months would be the longest wave consistent with the ocean's response. While this is *consistent* with the observed data at Bermuda (Fig. 2) and from the model (Fig. 6), two cautionary remarks are in order. First, we are pushing the limits of the data rather far. Second, this limitation only applies for a wave-like response; no such limitation is imposed on merely local forcing. One expects that purely local forcing, however, would be much less likely to develop the large amplitudes that could arise from consistent, 5-yr long forcing over the width of the ocean.

Slope of sea level between Bermuda and the United States

Once we are able to compute the variability of sea level at Bermuda from wind forcing, an interesting question comes to mind: What can this tell us about historical variability of Gulf Stream transport? The answer is tantalizing but elusive. Though our results so far are inconclusive, we sketch our preliminary results here because the question is important.

Sea level at the coast is obviously affected by processes on the continental shelf. Nevertheless, the sea level difference between Bermuda and the U.S. East Coast has traditionally been thought of as an approximate measure of the pressure gradient across the Gulf Stream [e.g., see section 4.3.1, Fofonoff (1981)]. The fluctuations in the slope between Bermuda and Charleston are remarkably large; the peak-to-peak changes of 20 cm (at these low frequencies) are roughly one-quarter of the total signal. The presentation by Levitus (1990, his Fig. 9) uses less low-frequency smoothing and the peaks are larger. The sense of the signal is that in the 1955–65 era the slope across the stream was larger than normal and that in the early 1970s the slope was smaller. This result suggests that the transport of the stream dropped significantly between the mid-1960s and the early 1970s and then increased.

Cross-spectra between sea levels at Bermuda and Charleston (not shown) show that sea level on the East Coast is highly coherent and consistently out of phase with Bermuda by 70°–90° at the longest periods we can resolve (this is significantly different from zero, with error bars of only $\sim$15° at 90% confidence). Whether we should interpret this as a physical delay (25 months at 100-mo periods) or as a phase shift, we cannot say. The mean wind curl over the ocean is also coherent with the sea level difference, suggesting (to no one's surprise) that the low-frequency variability in the Gulf Stream at these periods is also wind forced. However, the higher-resolution winds, rather than the mean wind curl, are found here to be more highly coherent with sea level than is the mean curl.

The appropriate test of the effectiveness of the Bermuda–Charleston sea level difference as a good measure of Gulf Stream transport is to compare this difference with transport data. Since the late 1960s observations of transport in the Florida Current have been made by several methods [see, e.g., the summary by Schott et al. (1988)]. Sanford and Larsen (1985) found a remarkably high correlation between the cable results and direct observations of transport by conventional means. Because the longest available set of continuous data is composed of the cable measurements, we chose to use them for comparison with sea levels.

Maul et al. (1984) have shown sea level differences across the Florida Current to be a better measure of transport and to compare better with cable voltages than sea level at the coast used alone. A comparison between cable voltages and sea level differences across the Florida Straits has been made by Mayer and Maul (1991), who studied the variability at the annual period and the first two harmonics. Noble and Gelfenbaum (1992) compared variability in bottom pressure and sea level at coast, again at the seasonal scales. The variability within the Florida Straits, at the annual period, appears to be coherent with the local winds as well, as Lee and Williams (1988) found. These comparisons are encouraging but may not give robust results, as it is well known that everything appears to be coherent with everything else at the annual period.

We compared the transport of the Florida Current at ~27°N derived from "cable voltages" (e.g., see Larsen 1992) with the sea-level slope from Bermuda to Charleston, simply by plotting both signals and making visual comparisons; the records are not long enough for meaningful cross spectra. The "correlation to the eye" of the energetic variations appears quite good at some times and poor at others. Various attempts to improve our results led to little; we suspect that one large source of error in our comparisons is the slope of sea level across the wide shelf off Charleston. We are not aware of any existing work on this subject, and our attempts to reduce this source of variability will be reported separately in more detail. One other source of significant error is that part of the transport that is not directly wind driven but is contributed from the flow from across the equator, making up for deep thermohaline flow. For a discussion of this issue, see, for example, Hautala et al. (1994) or Schmitz et al. (1992).

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