Gulf Stream Transport Variability at Periods of Decades

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ABSTRACT

Variability of sea level on the offshore side of the Gulf Stream has been estimated with a wind-forced numerical model. The difference in sea level between the model and coastal tide gauges therefore provides an estimate of variability of the Gulf Stream. These results can be compared with direct measurements of transport; the agreement is surprisingly good. Transport estimates are then made for sections offshore of four major tide stations along the U.S. East Coast. When data since World War II are used, the spectrum of sea level at the coast appears to peak at periods of ~150–250 mo. The difference signal (ocean minus coast), however, which the authors interpret as transport variability, has a weakly red spectrum. Power decreases at somewhat less than f^{-1} at periods just less than ~500 months but decreases strongly at periods less than ~150 months. The low-frequency variability arises primarily from the influx of open ocean Rossby waves. The large variance at low frequencies suggests that measurements of the transport of western boundary currents do not have many degrees of freedom; measurements made many years apart may vary substantially because of this localized variability. Sea level at the coast is coherent over long distances, but the incoming Rossby wave radiation from the open ocean has a relatively short north–south scale. These results emphasize that transport measured at one location along the coast may be incoherent with transport at locations only ~200 km away. As a result, measurements at one location will in general not be representative of transport along the entire coast.

1. Introduction

The transport of a major western boundary current is one of the fundamental measurable parameters of ocean circulation. Early work with absolute deep velocity measurements, such as by Warren and Volkmann (1968), showed that, even with a great deal of careful effort, the resulting uncertainties were large. The groundbreaking studies of the Florida Current by Richardson and his colleagues (Schmitz and Richardson 1968; Richardson et al. 1969; Niiler and Richardson 1973) found an annual cycle amid the large variability, but several years of data are necessary before an annual signal emerges.

In order to understand the processes behind time-varying signals, it is helpful to know the signal's variance and an estimate of the uncertainty of the measurements. The purpose here is to explore the long-term variability of Gulf Stream transport by using the available long data records of wind and sea level. Surprisingly, we find that the spectrum of Gulf Stream transport—at least to the extent that we are able to estimate it—continues to increase in power out to periods of several decades. For our calculations here, the limiting data are winds. The open ocean winds since the late 1940s are fairly well known, but data gaps during the two world wars make calculations that use data prior to 1946 much more difficult.

The method we use here seems new but straightforward. We are able to compute the height of sea level on the offshore side of the Gulf Stream from a wind-driven model, as described in the next section. Sea level at the coast is well known from measurements at tide gauges. The difference signal between these two datasets can then be used as a measure of the variability of Gulf Stream transport. Comparisons between our estimates of variability and direct transport measurements show that the underlying assumptions are met quite well.

The low-frequency variability in sea level at the coast has been known for decades (e.g., Hicks et al. 1983) although the similar signals in the open ocean have not been known for quite so long (e.g., Levitus 1990). The obvious question arises: does sea level at the coast merely go up and down in close agreement with the fluctuations offshore? If so, there would be no implications for changes in the transport of the boundary current. If these variations are not in phase, however (which is our principal result), the transport of the boundary current will fluctuate. If we are to understand the variability in

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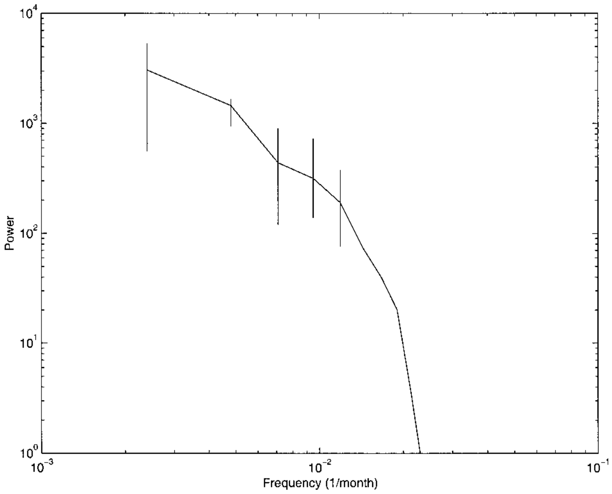

FIG. 1. Spectrum of modeled ocean fluctuations estimated from the mean of raw periodograms offshore of Fernandina, Florida; Charleston, South Carolina; Norfolk, Virginia; and Lewes, Delaware. The steep rolloff at periods shorter than 60 mo is caused by a lowpass filter applied to the winds that force the model. The vertical bars show the range of the power in each of the four raw periodograms that form the mean. The model output is 480 months long, beginning in 1950 (see text) and is forced with COADS winds.

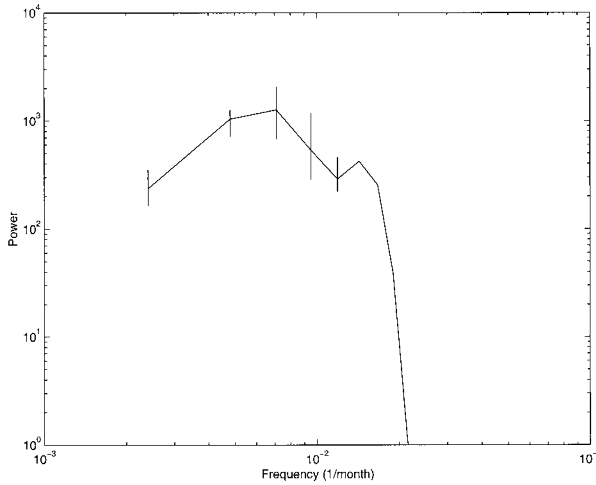

FIG. 3. Spectrum of sea level fluctuations estimated from the means of raw periodograms at Fernandina, Charleston, Norfolk, and Lewes, as in Fig. 1. The steep rolloff at periods shorter than 60 mo is caused by a low-pass filter applied to the winds that force the model. Vertical bars show the full range of power in each of the four raw periodograms that form the mean.

transport of the Gulf Stream, it appears essential that we understand these fluctuations, both in space and in time.

2. Wind-forced model results

In a series of papers (Sturges and Hong 1995; Sturges et al. 1998; Hong et al. 2000) we have shown that, at least within the North Atlantic subtropical gyre, the height of sea level computed with a simple wind-forced model agrees remarkably well with observations of several kinds. Sturges and Hong (1995, Fig. 4) found that model calculations of Rossby wave signals agree surprisingly well with low-frequency sea level observed at the Bermuda tide gauge, over a range of 16 cm, and with excellent phase agreement. Hong et al. (2000, Fig. 6) similarly found that the model results agree well with a tide gauge in Puerto Rico. Sturges et al. (1998, Fig. 7) found that the model variability agrees with sea level as estimated by dynamic height from historical hydrographic data over a range of ~20 cm. Indeed, Hong et al. (2000, Fig. 13) find that sea level on the U.S. East Coast can be modeled surprisingly well using only the wind forcing over the open ocean. The model is forced with COADS winds (e.g., Slutz et al. 1987). The mean curl at each grid point is removed; our results, therefore, speak only to the variability, not to the mean flow. Al-

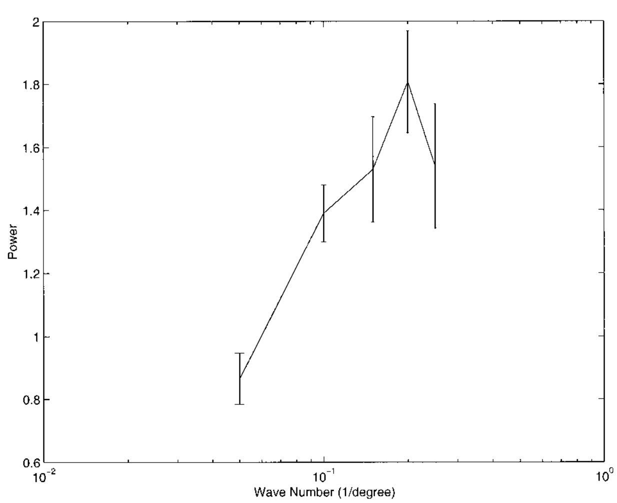

FIG. 2. The N–S wavenumber spectrum of model output along 66°W from 18° to 38°N in the variance-preserving form. The vertical bars show 90% confidence limits.

TABLE 1. Transport, 10^6 m^3 s^{-1}, in the Miami–Bimini section of the Florida Current. The 1964–68 data are averaged from the dropsonde measurements of Niiler and Richardson (1973), Richardson et al. (1969), and Schmitz and Richardson (1968). Data from 1969–74 are from (daily) Cable data (Jupiter–Settlement Pt.) provided by J. Larsen, NOAA/PMEL. The standard deviations listed are not of the measurements, but apply to the yearly means (see text).

Year	Mean	Std. dev.
1964	33.0	1.7
1965	33.5	1.0
1966	31.8	0.8
1967	26.4	1.9
1968	29.8	1.3
1969	31.1	0.5
1970	32.4	0.4
1971	32.5	0.3
1972	33.8	0.3
1973	32.4	0.2
1974	32.1	0.3

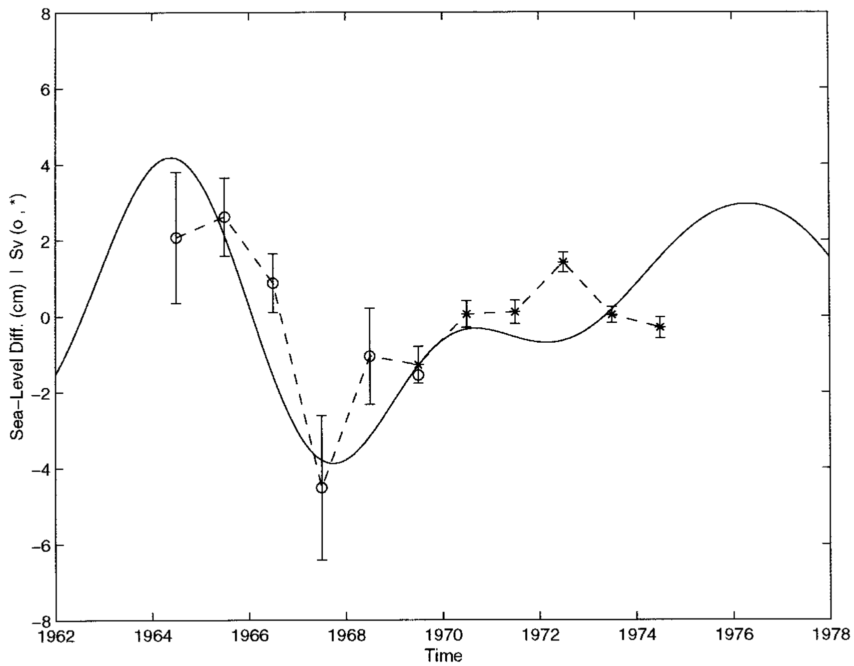

FIG. 4. Comparison between two estimates of variability in Gulf Stream transport offshore of Miami, Florida. The solid curve shows the difference between the Miami tide gauge and model output offshore. Open circles correspond to measurements by the dropsonde method; asterisks, transport by cable voltage (Jupiter–Settlement Pt.). Measured values show the mean of all measurements available within a full year; no low-pass filtering is done. The variability in transport is shown; mean of each set is removed (see Table 1). Miami tide gauge has the mean and trend removed.

though we are using model output from our own work, other wind-forced models can give similar results; see, for example, Ezer et al. (1995).

The open ocean variability at long periods is not small, but can be ~25% of the total signal. The range, peak to peak, is nearly 20 cm at 18°N, and the change in sea level across the stream there is ~75 cm.

Figure 1 shows the mean spectrum of modeled ocean variability on the offshore side of the stream at four locations: Fernandina, Florida; Charleston, South Carolina; Norfolk, Virginia; and Lewes, Deleware (specific details of the model output at these locations will appear below). The steep rolloff at periods shorter than 60 mo is from the filter used on the wind data that forces the model. An important feature of Fig. 1 is that the spectrum decays at least as steeply as $\sim f^{-1}$ at all periods shorter than ~500 mo, the longest periods we can compute. Because of the limited (and uncertain) number of degrees of freedom here, the error bars show the full range of the four individual raw periodograms.

The modeled variability at periods of roughly 5 yr and longer has a characteristic pattern in the open ocean: a series of wind-forced long Rossby waves is seen to cross the Atlantic; at any instant in the main anticyclonic gyre there are usually several highs and lows in the north–south direction, but only one in the east–west direction. Thus, the wavenumber spectrum in the N–S direction peaks at shorter wavelengths than in the E–W.

Figure 2 shows the wavenumber spectrum of modeled sea level variability in the ocean along 66°W. We have model output every month; this figure is the result of averaging nearly 500 individual values. A spectrum along 62°W looks essentially the same. There are, of course, very few independent estimates. Although the wind power peaks in the vicinity of ~100 mo, the ocean's response (e.g., Fig. 1) has a red spectrum. The power is concentrated at a N–S wavelength of ~500 km for *energy at these long periods.* We therefore conclude that, because these signals make up the dominant energy that is impinging upon the offshore side of the Gulf Stream, the N–S decorrelation distance is approximately one-quarter wavelength, ~125 km. While this is an important result, it also implies that the four values used for the average in Fig. 1 are essentially independent.

3. Sea level at the coast

By contrast with the open ocean signals, the spectrum of sea level at the coast peaks at a frequency that is resolvable with the data used here. Figure 3 shows a spectrum composed of the mean of the periodograms of tide gauge signals at (as before) Fernandina, Charleston, Norfolk, and Lewes. The peak is broad and in the range of 150–250 mo; the power decreases to longer periods at all stations. The spectra in Fig. 3 are com-

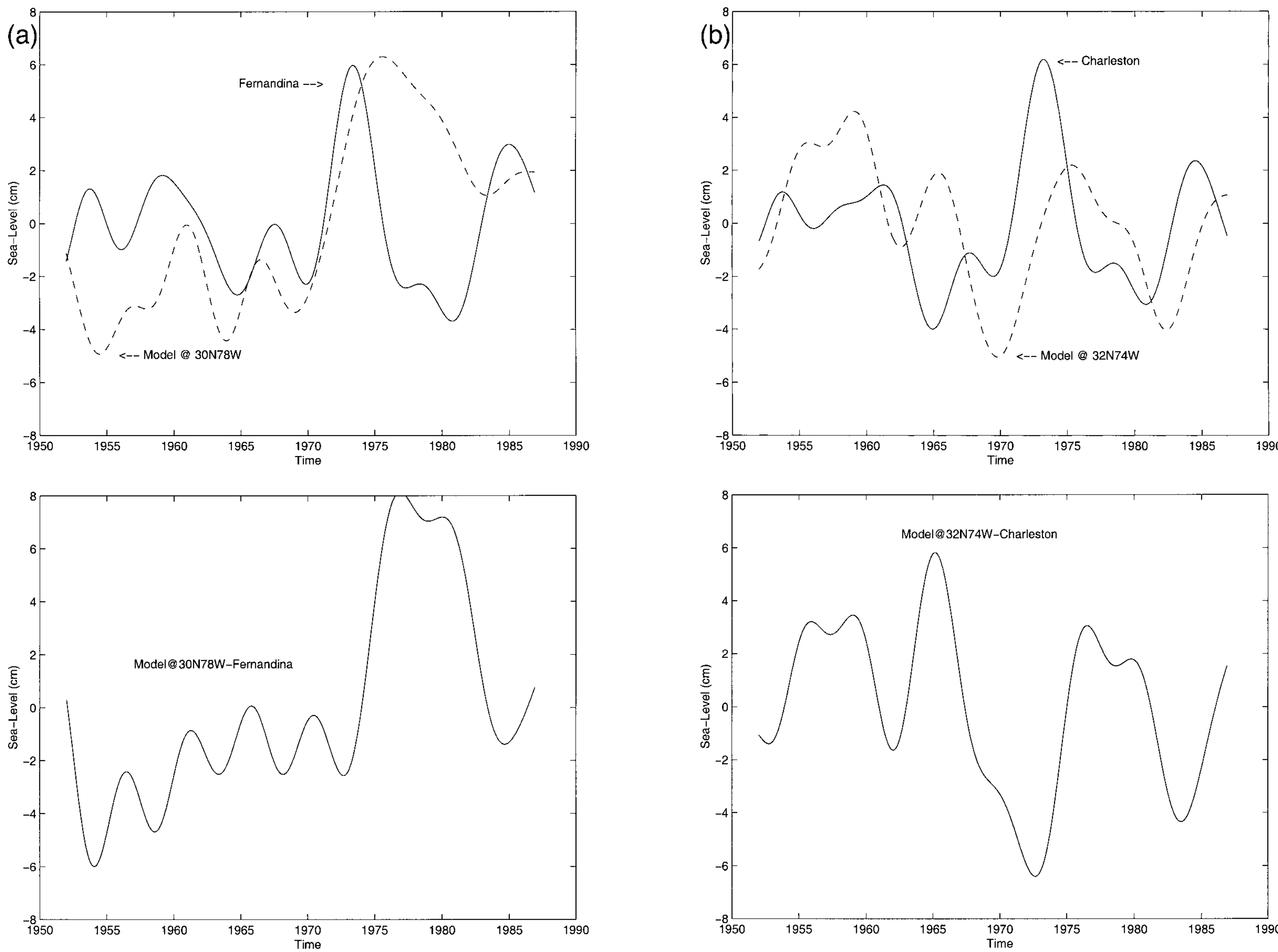

FIG. 5. (top) Observed (coastal) sea level (a) at Fernandina and (b) at Charleston and nearby model output on the offshore side of the Gulf Stream. Both signals have been low passed to suppress variability at 3 years and pass signals at 5 years. (bottom) The difference between the upper two signals, intended to represent variability in transport of the Gulf Stream. Note that both panels have the same scales.

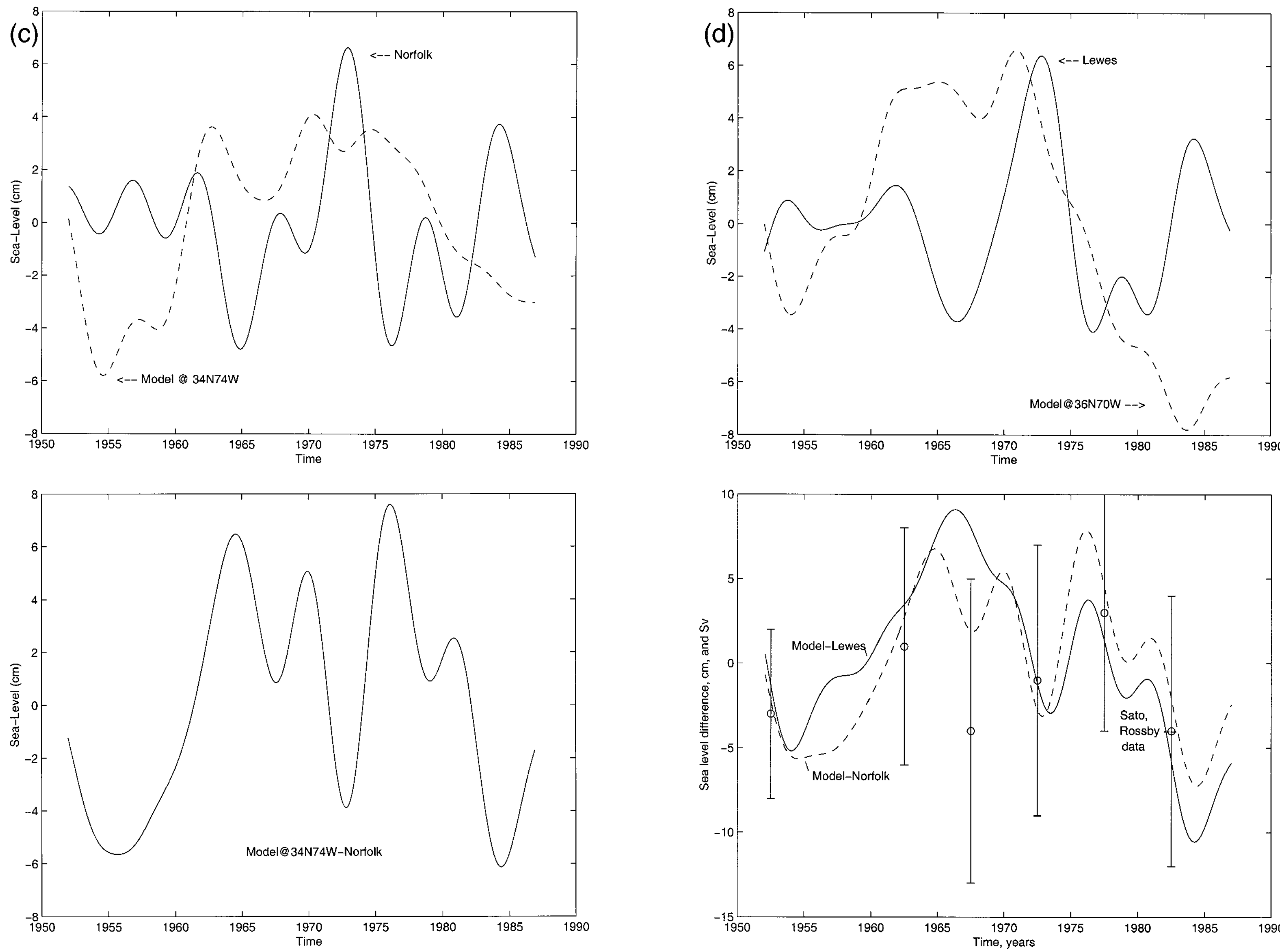

FIG. 5 (*Continued*) (top) Observed (coastal) sea level (c) at Norfolk and (d) at Lewes, Delaware, and nearby model output on the offshore side of the Gulf Stream. Both signals have been low passed to suppress variability at 3 years and pass signals at 5 years. (bottom) The difference between the upper two signals, intended to represent variability in transport of the Gulf Stream. In (c) note that both panels have the same scales. In (d) the difference signal from 5c at Norfolk is repeated (dashed), and the data points and vertical error bars are from the geostrophic transport calculations of Sato and Rossby (1995).

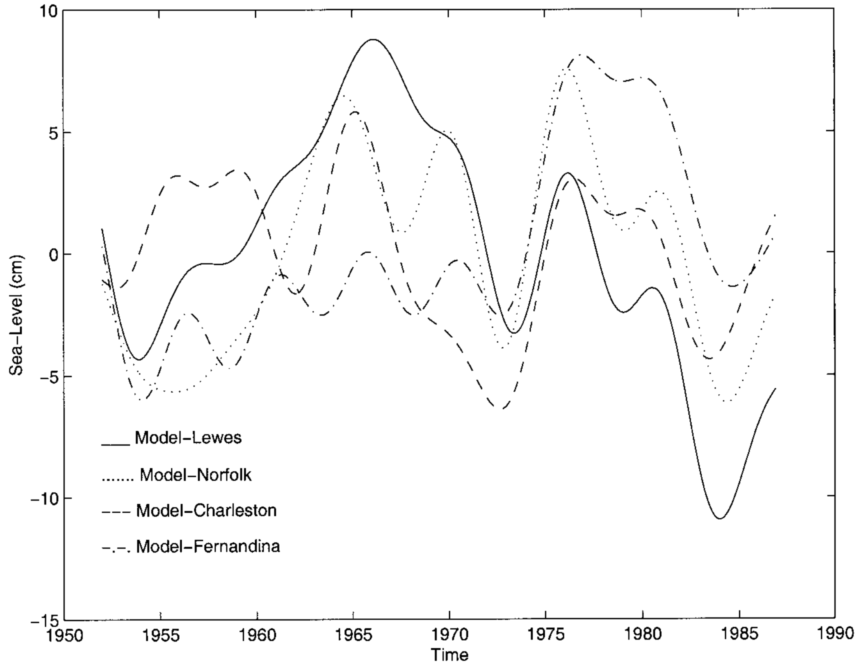

FIG. 6. A composite of the differences in height across the stream at all four of the locations shown in Fig. 5.

puted, using records beginning after World War II, to coincide with the wind data. Spectra computed with the available longer records, however, give similar results. There is thus a contrast in that the coastal stations show a power peak at intermediate periods, whereas the offshore variability continues to rise to longer periods.

4. Transport variability from sea level

Knowing the difference in height of the sea surface between two points allows us to calculate the geostrophic velocity at the sea surface. In order to determine the transport, however, we need to know the interior baroclinic structure. If we assume that the changes in baroclinic structure are small, then knowledge of changes in sea surface slope is equivalent to knowledge of changes in transport.

The foundation for believing that changes in baroclinic structure are small is based on one of the principal findings of our previous studies: the modeled height of the sea surface agrees remarkably well with observations at every place where we have been able to make meaningful comparisons. This is found to be true both with hydrographic data and with tide gauges. The basis of the model is that the ocean's vertical structure changes imperceptibly over the timescales of our calculation, as specified initially from the observed mean density profile. The only changes to this structure are the result of (linear) Rossby waves. The baroclinic structure of our model is fixed; therefore, since the model output agrees so well with observations, changes in sea surface slope can be concluded to be equivalent to changes in the total vertically integrated transport.

5. Comparison with observed transport

Before we present our primary results, it is useful to show that these calculations really do tell us about variations of Gulf Stream transport. It is possible to make

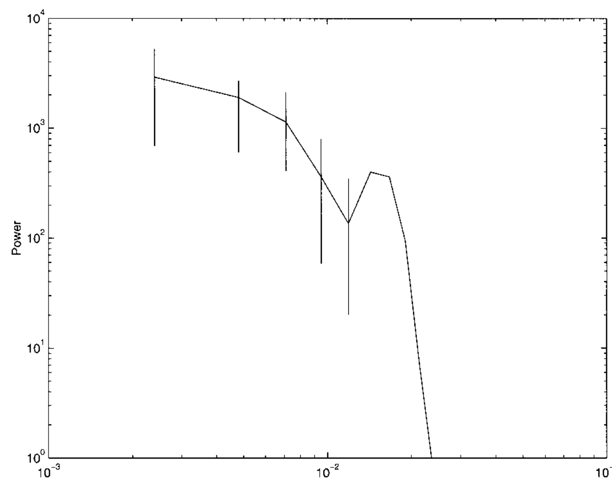

FIG. 7. Spectrum formed from the mean of the four raw periodograms of the difference signals across the stream offshore of the four tidal stations studied here: Fernandina, Charleston, Norfolk, and Lewes, as shown in Fig. 6. The error bars show the full range of the raw periodograms.

limited comparisons between our results and actual measurements of transport, although at these low frequencies the measurement requirements are severe. First, absolute measurements of transport are relatively scarce. Second, because our model computes only the variability and not the absolute transport, we can make meaningful comparisons only with measurements that are long in time at the same place.

Figure 4 is a comparison between our results at Miami, Florida, with direct measurements. Those prior to 1970 are from the direct observations by dropsondes (see Table 1); the later data are from the cable measurements, kindly supplied by J. Larsen (e.g., see Larsen 1992). For the dropsonde data, all values available within a calendar year are averaged together. The error estimates in Fig. 4 are intended to show the standard error of the mean. They are made with the assumption that monthly dropsonde values are independent and that the (daily) cable values are independent after 10 days. The model has zero mean; the transport measurements have been offset to allow agreement in the mean. The agreement between the sea level height differences and observations is quite good and gives confidence that these height differences give information about transport variability.[1]

The results of Sato and Rossby (1995) are also available to allow comparison with direct measurements, although their results give geostrophic transport values without the benefit of absolute velocity measurements at depth. Comparisons with their data, which are also very encouraging, are shown (below) in Fig. 5d.

6. Results

Figures 5a–d show the results at Fernandina, Charleston, Norfolk, and Lewes. These use the model output at a location estimated to be appropriately ''offshore'' at 30°, 32°, 34°, and 36°N. The upper panels show both sea level observed at the coast and from a point in the model on the offshore side of the Gulf Stream; the lower panels show the differences. The offshore location in the model can be chosen without regard to the side-to-side meanders of the stream. The model (see Sturges and Hong 1995; Sturges et al. 1998) computes only the westward propagating (forced) waves, and has an open western boundary. The model computes only the open ocean variability and does not contain the return flow (the Gulf Stream).

The fundamental result here, from all these locations, is that the variability offshore is substantially out of phase with that at the coast; the difference signal has roughly the same amplitude as the individual signals. While this result may seem plausible, it could not have been known with confidence beforehand. The changes from south to north along the coast tend to be gradual from one location to the next. The north–south variability on the offshore side was described earlier (Fig. 1). Because the changes on the offshore side vary so much from one location to the next, the effect on the *differences* is that the changes in total height (or slope) across the stream, and hence transport, vary markedly.

Figure 5a, for the section off Fernandina, shows that a section made during ~1954 would find anomalously low transport, while one made during the late 1970s would find high transport, relative to the long-term mean. Off Charleston, Fig. 5b the transport appears to have a sharp minimum in ~1973 as well as in ~1984. Off Norfolk, the highs in ~1965 and ~1977 are qualitatively similar to those off Charleston, but quantitatively quite different. Note that the low in ~1956 off Norfolk, by chance, coincides with a high off Charleston. Similar variability is seen in the section off Lewes except that the peak amplitude is larger. Measurements made in ~1966 would give substantially different results from those made in ~1984. The peak-to-peak difference here is ~20 cm, which is ~20% of the total height difference across the stream.

Figure 5d, lower half, also shows the variability of transport as determined by Sato and Rossby (1995, their Fig. 10). They considered a region extending from ~35° to ~37°N on the offshore side and ~36° to 40°N on the inshore side. Because their N–S range of values is larger than is appropriate for a single location with our model, the difference signal for Norfolk is repeated (dashed), along with that for Lewes. No adjustments have been made to our results or to theirs for this comparison. The data point for the late 1960s agrees only at the upper error limit with the model results, although all the other points agree quite well. Their Fig. 10, however, shows that the data point for the 1965–70 interval is based on only approximately seven station pairs, whose range is ~30 Sv (Sv $\equiv 10^6$ m^3 s^{-1}), so the lack of better agreement here may not be critical. Perhaps because of the large latitudinal range and the 5-yr block averaging, the range of their observed values is only from -4 to $+3$ Sv, while the model results suggest that the range could be as large as $\sim\pm 10$ Sv.

Figure 6 combines the transport variability at all four sections shown in Fig. 5. Transport is low in the early 1970s and 1980s *at all locations* but high in between. During the interval from the early 1950s until the early 1970s, however, the signals drift in and out of phase

[1] An attempt was made to make a similar comparison using the more recent cable measurements (now available at the NOAA Atlantic Oceanographic and Meteorological Laboratory, Miami). Because the tide gauge used in the comparisons of Fig. 4 went out of service, the introduction of a different tidal datum for the new gauge causes a new (and essentially arbitrary) offset to be introduced into the comparisons. Furthermore, the calibration of the newer cable data is most reliable after 1991 (D. Wilson 1999, personal communication). As a result, the more recent comparisons are not sufficiently reliable to be useful. It would also seem desirable to be able to make comparisons with the transport observations reported by Johns et al. (1999). But, again, the shift in the tide gauge data introduces sufficient ambiguity to seriously weaken the strength of the comparison.

with each other, so it is difficult to draw general conclusions during that time. The differences between the transports at neighboring locations are striking.

Figure 7 shows the mean of the periodograms of the differences computed across the Gulf Stream at the four gauges used here. The spectrum is red, but not strongly so; the power decays somewhat less steeply than f^{-1} in the last two low-frequency bands. Nevertheless, a disconcerting feature of Fig. 7 is that the spectrum continues to rise out to the longest periods we can compute. The differences in transport at the locations selected here are consistent with the short north–south scale of the open ocean model results.

7. Conclusions

The transport variability of the Gulf Stream is estimated here as the difference between observed coastal sea level and modeled sea level on the offshore side of the stream. The estimates we have made of transport variability agree surprisingly well with the available transport measurements from historical data, both off Miami (Fig. 4) and downstream from Hatteras (Fig. 5d). An important finding here is that the variations offshore *are not in phase* with the variations at the coast. During some time intervals the transport is anomalously high at almost all locations, yet during other times the locations all have seemingly random phases. The variability from one location to the next is as large as the variation at a single location. It is risky to make statements about coherence from one location to the next because there are so few degrees of freedom in these records. Still, it appears that at locations only a few degrees of latitude apart, the signals are not reliably correlated. This conclusion is consistent with the short N–S length scale of the incoming low-frequency Rossby wave energy.

The spectrum of sea level variability at tide gauges has a barely resolvable peak in the vicinity of 150–250 mo. The variability in the open ocean just offshore, however, has a spectrum that continues to rise out to the longest periods available to our wind-forced model. The difference between these two, which is one of the primary results here (the unassuming Fig. 7), has a spectrum that continues to rise out ~500 mo, the longest periods we can determine. This result is based on the differences across the Gulf Stream offshore of four long-term gauges. Between periods of 500 and 150 mo, the spectrum decays at a rate somewhat less than $\sim f^{-1}$, suggesting that the power is only ''barely red'' at these longest periods; we may speculate that the spectrum may be leveling off. Nevertheless, the spectrum falls off steeply for periods shorter than ~150 mo.

Figure 2 shows an estimate of the N–S horizontal wavenumber spectrum in the open ocean. If we assume that the majority of the variability at the coast is forced by the incoming long-wave Rossby wave energy, we might assume that the north–south scale will be similar to that shown in Fig. 2. It turns out, however, that the tide gauges are highly correlated along the coast at the lowest frequencies, as can be seen from inspection of Fig. 5. Even though they are coherent, there are substantial and irregular phase shifts from one gauge to the next. The difference between coastal sea level at Fernandina and Lewes arises largely from these phase shifts, so the difference between them is as large as the individual variability at a single location.

Our model includes only the variability that is directly wind forced. Because the downstream increase in transport arises from nonlinear effects in the recirculation region, the actual variations in the stream may differ from the results of our simple model. It is hard to imagine that the nonlinear effects would decrease the variability.

The transport of the Gulf Stream is found to have substantial variance at the same long periods as the open-ocean wind curl. A bothersome implication is that we should be slow to draw conclusions about the low-frequency behavior of the stream from a set of measurements based on only a few years' duration. In other words, if the spectrum peaks at periods of ~150–250 mo, observations of longer duration are required before we can begin to believe that the variability of this signal has been measured well.

The N–S wavenumber spectrum along the western side of the ocean shows that the power peaks in the vicinity of ~500 km, for a decorrelation distance of ~125 km. The actual transport of the Gulf Stream, and presumably other boundary currents, is modified with these relatively short N–S scales. As a result, direct measurements of transport at one location may be representative of only a short segment of the boundary current and may not be representative of the transport along the entire coast unless these N–S variations are accounted for.

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