Wind-Forced Transport Fluctuations of the Florida Current

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ABSTRACT

Volume transport fluctuations of the Florida Current (Gulf Stream), generated within the Straits of Florida by local meridional wind stress, are investigated. A simple coastal response model was applied to the Straits of Florida and forced by along-channel winds only. The predicted volume transports were in good agreement with transport estimates derived from moored current meters and cable voltages for synoptic scale winter winds in the period band from 4 to 10 days. Good agreement was also found for the annual transport cycle for the two years of available data, suggesting that the seasonal change in along-channel wind forcing provides a significant contribution to the annual transport cycle of the Florida Current.

1. Introduction

Volume transport fluctuations of the Florida Current (FC) have been observed within the Straits of Florida on time scales ranging from tidal to annual. Tidal transport variations are typically on the order ±1.5 Sv (1 Sv ≡ 10^6 m^3 s^{-1}) (Mayer et al. 1984). Seasonal changes are estimated at ±4 Sv and are asymmetrically distributed about a mean northward transport of about 32 Sv; i.e., maximum values occur during summer and minimum in fall (Niiler and Richardson 1973; Molinari et al. 1985; Larsen and Sanford 1985; Leaman et al. 1987). Between the tidal and annual periods a near continuum of transport fluctuations exists with amplitudes as large or greater than the seasonal signal (Lee et al. 1985). Many past studies have been concerned with FC variations in the 2–15 day period band, where current variability was explained by east–west meanders of the FC axis (Duing 1975; Brooks 1979) and atmospheric forcing (Duing, Mooers and Lee 1977). Within the Straits of Florida meandering motions have been associated with the generation of wind-forced continental shelf waves (Schott and Duing 1976; Brooks and Mooers 1977a and 1977b); whereas, north of the Straits instability theory has been used to account for the meanders (Niiler and Mysak 1971; Orlanski 1969; Orlanski and Cox 1973; Luther and Bane 1985).

Recent time series measurements of FC transport fluctuations between Florida and the Bahamas at 27°N, using a variety of observational techniques (Molinari et al. 1985b), indicate a strong connection between transport variations in the 2–10 day period band and local meridional wind stress (Lee et al. 1985). However, the physical mechanism responsible for these wind-forced transport changes is uncertain. The exact cause of the FC seasonal cycle is also not understood. There is observational evidence suggesting that the wind stress curl over the western Caribbean may force the seasonal cycle (Schott and Zantopp 1985). However, numerical model studies of the North Atlantic for a homogeneous ocean with real bathymetry and forced only by the wind stress curl over the Caribbean, produced an annual cycle of the FC that was in phase with observations, but with an amplitude of only ±0.5 Sv (Anderson and Corry 1985). The wind stress curl over the topography of the northwest Atlantic had the greatest effect on the annual cycle of the FC in the barotropic model with a seasonal amplitude of ±1.0 Sv. The full baroclinic model of Anderson and Corry (1985) driven by seasonal wind stress fields results in an annual cycle of the FC that was in phase with observations, but with an amplitude of ±2.0 Sv, which is still about a factor of 2 less than observed (Leaman et al. 1987).

2. Model

In this paper we examine a simple analytical model of FC transport fluctuations forced by local meridional wind stress only to account for the part of the spectrum that is coherent with the wind on weekly and annual time scales. The model was originally developed for the case of an alongshore wind blowing over a long, straight, coastal ocean (Csanady 1982). The model does not include the mean transport or baroclinic fluctuations. For a constant alongshore wind stress imposed suddenly at $t = 0$, parallel to the y axis over a semi-infinite shallow sea with a flat bottom in the domain $x \leq 0$ bounded by straight coasts at $x = 0$, L_x the governing transport and continuity equations are

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$$fV = c^2 \frac{\partial \eta}{\partial x} \tag{1}$$

$$\frac{\partial V}{\partial t} + fU = \tau_{ys} \tag{2}$$

$$\frac{\partial U}{\partial x} = -\frac{\partial \eta}{\partial t} \tag{3}$$

where U and V are the vertically integrated velocity components in the x and y directions respectively, f is the Coriolis parameter, η the free sea surface elevation, τ_{ys} the alongshore surface wind stress and $c = \sqrt{gH}$, with g the acceleration of gravity and H the total water depth, a constant. In this model, bottom friction is neglected, $\partial\eta/\partial y = 0$, for there are no cross-shore barriers and $\partial U/\partial t = 0$ as well as $\partial V/\partial y = 0$.

The governing equation for sea level is obtained by cross-differentiation and subtraction of equations (1) and (2) with use of (3):

$$\frac{\partial^2 \eta}{\partial t^2} - c^2 \frac{\partial^2 \eta}{\partial x^2} + f^2 \eta = 0. \tag{4}$$

The motion starts from rest so that the boundary conditions at the coasts, $U = 0$, from Eqs. (1) and (2) are

$$\left.\begin{aligned} fV &= c^2 \frac{\partial \eta}{\partial x} \\ \frac{\partial V}{\partial t} &= \tau_{ys} \end{aligned}\right\} \quad \text{at} \quad x = 0, L_x$$

which can be rewritten as

$$\left.\begin{aligned} V &= \int_0^t \tau_{ys} dt \\ \frac{\partial \eta}{\partial x} &= \frac{f}{c^2} \int_0^t \tau_{ys} dt \end{aligned}\right\} \quad \text{at} \quad x = 0, L_x.$$

Csanady (1982) has shown that the nonoscillating part of the solution to (4) is

$$\eta = \frac{\tau_{ys}}{c} t e^{x/R} \tag{5}$$

for time scales longer than a day; i.e., $ft \gg 1$, where $R = c/f = \sqrt{gH}/f$ is the barotropic Rossby radius of deformation.

The transport components associated with the nonoscillatory part of the solution are found from (1) and (2) substituting for η from (5):

$$V = \tau_{ys} t e^{x/R} \tag{6}$$

$$U = \frac{\tau_{ys}}{f}(1 - e^{x/R}). \tag{7}$$

3. Application of the model to the Straits of Florida

Csanady's (1982) solutions apply for the domain of $x \leq 0$; i.e., negative x. Therefore, when x is small compared to R, the alongshore transport will simply be accelerated by alongshore wind stress, $V = \tau_{ys}$ and $U = 0$, the coastal constraint. When x is large compared to R, then pure Ekman flow will result, $U = \tau_{ys} f^{-1}$ and $V = 0$. There exists an intermediate region given by $0.1 \leq |x/R| \leq 2.0$ where both U and V are significant. In the Straits of Florida at 27°N the mean depth $H = 460$ m, width $L = 92$ km, and $R = 670$ km. Thus $|x/R| = 0.14$ so that both U and V should be significant throughout the channel except very near the coast but with V somewhat larger. In this regard the flow response to alongshore wind forcing in the Straits of Florida can be considered as the combination of solutions for east coast and west coast semi-infinite seas. Along-channel winds should simply accelerate along-channel transport according to (6) and (7) by generating cross-channel surface Ekman transports that due to the combined influence of convergence and divergence on the sides of the channel will support an opposite cross-channel "adjustment drift" (Csanady 1982) below the surface Ekman layer that accelerates the along-channel flow. Sea level will adjust geostrophically to balance the developing along-channel flow according to (5), resulting in cross-channel surface slopes in geostrophic equilibrium with along-channel transport, Eq. (1).

The total wind-driven volume transport T_w, through the Straits can be estimated from (6) by

$$T_w = \int_0^L \tau_{ys} t e^{x/R} dx. \tag{8}$$

Estimates of T_w as a function of along-channel wind speed and duration are shown in Fig. 1. As expected from Eq. (8) wind-driven volume transports in the Straits of Florida are strongly influenced not only by

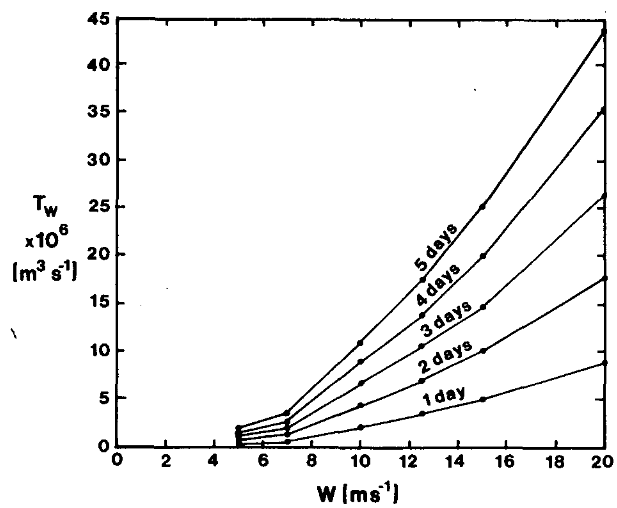

FIG. 1. Estimates of wind forced volume transports (T_w) through the Straits of Florida at 27°N as a function of along-channel wind speed (W) and wind duration.

the speed of the wind but also the duration. Typical winter along-channel wind events in this region have speeds ranging from 10 to 15 m s^{-1} and last from 1 to 3 days before decreasing or reversing. Therefore, we should expect to observe wind-forced transport fluctuations ranging from 2 to 15 Sv.

It is unlikely that these wind-driven transports will ever reach a state of frictional equilibrium because owing to the large depths the frictional adjustment times, $t_f = H/r$, in the Straits of Florida would be on the order of 10 days for a bottom resistance coefficient (r) of 0.05 cm s^{-1}. A 5 Sv wind forced transport fluctuation would require a 12 cm s^{-1} cross-sectional averaged current, $\langle v \rangle$, considered together with the bottom resistance coefficient, gives a quadratic bottom drag coefficient, $C = r/|\langle v \rangle|$ of 4×10^{-3}, which seems high. If C is computed from Manning's formulation $C = gn^2/H^{1/3}$ we get approximately 1×10^{-3} with Manning's number (n) equal to 0.025, which is a more commonly accepted value. Thus the 10-day frictional time scale is probably a minimum value, and it could be several times longer, justifying the neglect of bottom friction in the above model.

4. Comparisons with data

Time series measurements of volume transport through the Straits of Florida at 27°N were recently initiated as part of the Subtropical Atlantic Climate Studies (STACS) program (Molinari et al. 1985b). The measuring strategies developed to monitor transports of mass and heat through this channel are shown in Fig. 2, together with the first two years of volume transports from the three primary measurement systems: moored current meters, PEGASUS sections and submarine cable. Detailed discussions of these datasets can be found in Molinari et al. (1985a), Molinari et al. (1985b), Lee et al. (1985), Larsen and Sanford (1985), Maul et al. (1985), Schott and Zantopp (1985) and Leaman et al. (1987). The three measurement techniques were found to agree within 1 to 2 Sv. Transports were also highly correlated with bottom pressure from the Florida shelf edge, indicating a positive relationship with cross-channel sea level slopes.

Transport time series (Fig. 2) show large fluctuations of 5 to 10 Sv, occurring on time scales of 2 to 20 days. Lee et al. (1985) found significant coherence between transport and along-channel winds at periods of 5 to 6 days and 10 days for the winter period of November 1982 to April 1983. The 1982 seasonal signal is embedded within the background of intense low-frequency fluctuations, but there is a large transport burst in August reaching nearly 40 Sv and several minima in the fall where the transport drops to near 20 Sv (Fig. 2). These two extremes appear to make up the seasonal signal for 1982, in general agreement with that first reported by Niiler and Richardson (1973). In 1983 the seasonal signal is made up of two bursts of increased

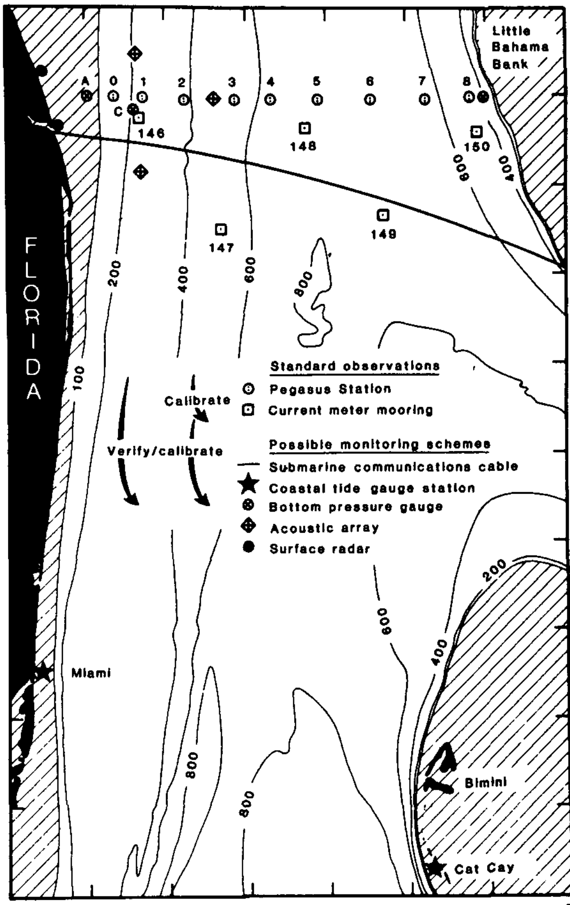

FIG. 2a. Subtropical Atlantic Climate Studies (STACS) volume and heat transport monitoring strategies.

transports in May and July and several minima in the fall.

Transports derived from the STACS moored array (T_m) for the period 21 December 1983 to 30 January 1984 are shown in Fig. 3, together with along-channel wind and wind stress from Palm Beach and cross-channel sea level difference between Cat Cay (Bahamas) and Miami. The Palm Beach coastal wind speeds were multiplied by 2 to be more representative of oceanic winds as determined by Kourafalou et al. (1984). The wind stress vectors were computed from $\tau = \rho C_d |\mathbf{W}| \mathbf{W}$, where $\mathbf{W}$ is the wind vector and C_d is a variable drag coefficient; $C_d = 1.6 \times 10^{-3}$ for $|\mathbf{W}| \leq 7$ m s^{-1} and $C_d = 2.5 \times 10^{-3}$ for $|\mathbf{W}| > 7$ m s^{-1}. The vertical event lines identify northward (solid) and southward (dashed) wind episodes. Increased (decreased) transports lag northward (southward) wind events by less than one day and are near in phase with sea level differences. Variance conserving spectra show well-defined variance peaks centered at a period of 7 days for the along-channel wind stress, moored transport and cross-channel sea level slopes (Fig. 4). These

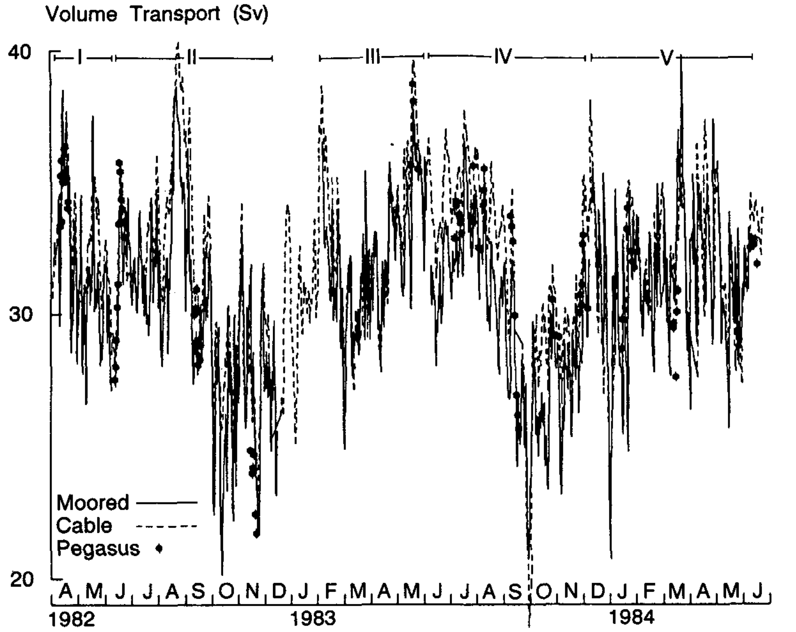

FIG. 2b. Time series of 48-hour low-pass filtered volume transports at 27°N determined from moored current meters (solid), cable voltages (dashed) and Pegasus sections (dots).

variables were mutually coherent above the 95% significance level at the 7 day period with τ_{ys} leading T_m by 45° in phase and T_m leading $\Delta\eta_x$ also by 45°, which amounts to a 21 hour lag for both, in agreement with the wind forced model presented earlier.

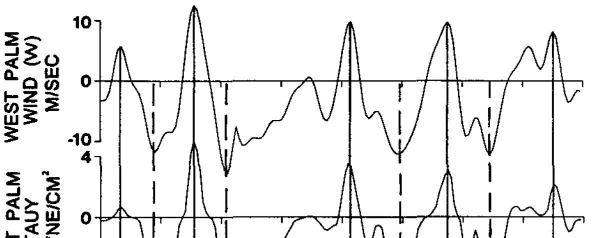

FIG. 3. Time series of 40-hour low-pass filtered Palm Beach along-channel wind speeds and wind stress (τ_{ys}), cross-channel sea level difference for Cat Cay, Bahamas, minus Miami ($\Delta\eta_x$), and volume transports (T_m) from STACS current meter moorings. Vertical event lines are for northward (solid) and southward (dashed) wind events.

Johns and Schott (1987) also found a near constant time lag of about one day between along-channel wind stress and transport at the coherent periods of 3.5, 8 and 20 days for the STACS-V measurements. This indicates that the transport response time to wind forcing is on the order of one day. The coastal response model presented here indicates a similar time scale for the transport response, since Ekman transports will develop as f^{-1} or over several hours and this in turn must accelerate the along-channel transport, which should occur within an inertial period or about one day for this latitude. The geostrophic adjustment time, $t_g = L_x/\sqrt{gH}$ of sea level to the alongshore transport is short, on the order of one hour for the Straits of Florida horizontal and vertical scales.

The applicability of an Ekman coastal response model to account for FC transport fluctuations forced by local along-channel winds is clearly indicated in the STACS-V data, a period of significant along-channel winds. Figure 5 shows the along-channel wind stress plotted, together with the vertically averaged cross-channel current from the westernmost mooring of STACS-V [$\bar{u}$ (171)] and across the FC front [$\bar{u}$ (174) and $\bar{u}$ (175)], FC transport (T_m), cross- and along-channel sea level difference ($\Delta\eta_x$, $\Delta\eta_y$), and the gradient of demeaned vertical-averaged temperature from moorings across the FC cyclonic front ($\partial\overline{T'}/\partial x$). The overbar indicates a vertical average. The cross-channel current and temperatures are averaged over the extent

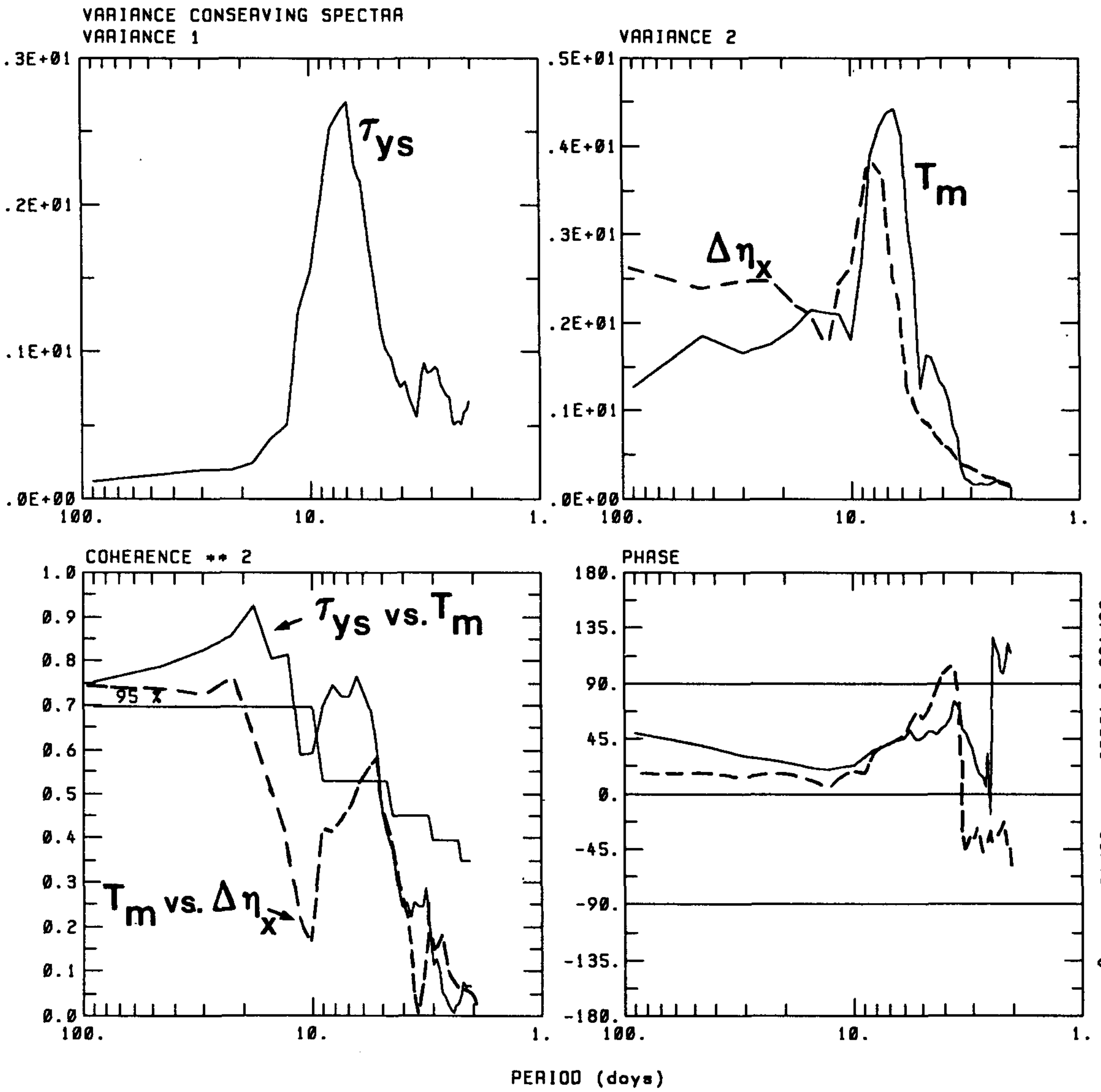

FIG. 4. Variance conserving spectra, coherence squared and phase between Palm Beach along-channel wind stress (τ_{ys}) with moored transports (T_m) and moored transports with cross-channel sea level differences ($\Delta\eta_x$) at Cat Cay minus Miami for the period 1 December 1983 to 1 March 1984.

of the mooring only, which is below the surface Ekman layer in the geostrophic interior. Cross-channel sea level differences are for Cat Cay, Bahamas, and Miami, which is about 130 km upstream from where transports were measured. The short record of along-channel sea level differences was determined from the available Palm Beach and Miami sea level records.

Vertical average cross-channel flow is clearly 180° out of phase with along-channel wind stress with no visible phase lag. Since the current meters were located below the surface Ekman layer this out of phase relationship is consistent with the "adjustment drift" that "coastal constraint" ($U = 0$ in the coastal boundary layer) requires to develop opposite to the surface Ekman transport on a time scale of f^{-1} or several hours. Similar cross-channel flows can be observed at the deeper moorings; however, due to the greater depths these flows are weaker and can be masked by meander motions.. Florida Current transport fluctuations are visibly well-correlated with cross-channel flows with about a one day lag. Negative or westward cross-channel flow accelerates the transport and eastward flow is followed by a reduction in transport. Sea level then appears to adjust geostrophically to the developing transport fluctuations with little phase lag, according to equation (1). Both cross- and along-channel sea surface slopes appear to adjust as part of the response process. Transport increases are balanced by positive cross-channel slopes as anticipated from (1); however negative along-channel slopes also occur, apparently as a response to the westward adjustment drift. Along-channel slopes were omitted from (2), but this should not seriously affect our results since this process could be considered as a combined effect of the adjustment drift so that the solution for along-channel transport (6) remains the same. The cross-channel temperature gradient also appears to be influenced by this barotropic response process, which could affect baroclinic phenomena; this will be discussed later.

Volume transport was hindcast using the Palm Beach winds for the period 21 December 1983 to 30 January 1984 and the simple model presented in Eq. (8) integrated across the channel. The model can be used to predict transports from real winds, for the time scale of the transport response was shown to be short, about one day, compared to the 2- to 20-day time scales of the wind forcing. In particular, during STACS-V the dominant wind forcing occurred at a period of 7 days (Fig. 4). Modeled wind-driven transports (T_w) are compared to moored transports (T_m) and the along-channel wind stress used in the model (Fig. 6). The model was iterated over different time integration intervals until the best fit with measured transports was obtained.

The best fit was found for a 48-hour integration interval, which has a linear regression coefficient of 0.85

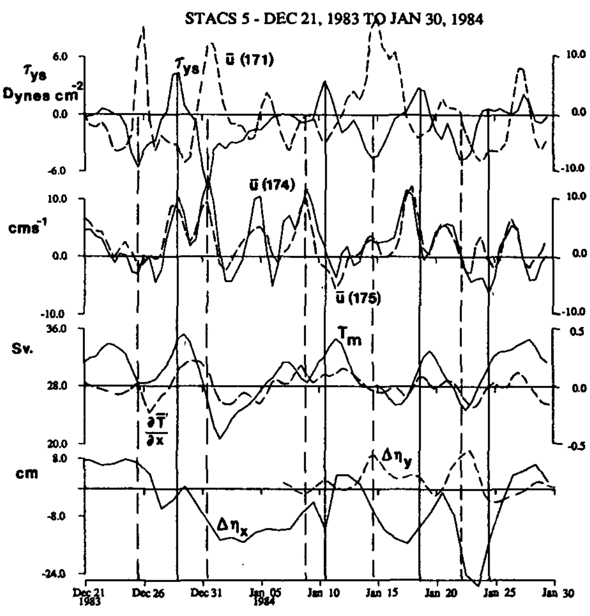

FIG. 5. Time series of 40-hour low-pass filtered Palm Beach wind stress (τ_{ys}), vertical averaged cross-channel currents ($\bar{u}$) at moorings 171 (shelf edge), 174 (cyclonic shear zone), 175 (anticyclonic shear zone), moored transports (T_m), gradient of demeaned vertical averaged temperature from moorings across the cyclonic front and cross- and along-channel sea level differences ($\Delta\eta_x$, $\Delta\eta_y$).

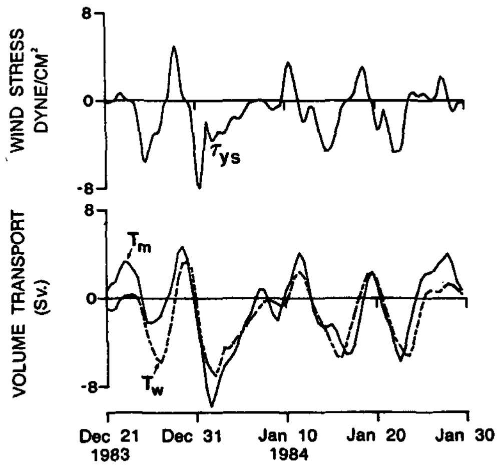

FIG. 6. Volume-transport time series derived from moored current meters (T_m) and hindcast from along-channel winds (T_w) for a 48-hour integration period, together with along-channel wind stress.

with measured transports and a rms difference of 1.9 Sv. Figure 6 shows the amplitude and phase of the modeled transport to agree quite well with measured values. Spectra comparison between modeled transports using the 48-hour integration and measured transports for the 3-month winter period from 1 December 1983 to 1 March 1984 (Fig. 7) shows that the modeled transports clearly resolve the 7-day period peak contained in the moored transports. Coherence squared is also high between the measured and modeled time series in the 5- to 10-day period band and at periods greater than 15 days, with small phase differences.

An interesting question is why a 48-hour integration should produce the best results. The effect of introducing an integration interval is similar to including a mechanism to balance the momentum input by the wind. We have shown previously that friction cannot provide this on time scales of the wind forcing. Without friction, wind-driven transports will continue to grow until the wind stress changes sign. This is clearly not supported by the measurements (Fig. 6). The good results obtained from the 48-hour integration suggests that the fluid has no memory of the wind forcing that occurred two days prior. This can be understood when one considers that the model applies to along-channel winds over an infinitely long north–south oriented channel, whereas the Straits of Florida has a finite length. The northern extremity of the channel occurs where the Bahama Bank drops off into the Blake Plateau and the Florida shelf begins to widen, just north of the STACS section. The southern end of the north–south oriented channel occurs where the channel turns west off the Florida Keys and aligns with an east–west orientation. At this point meridional winds will no longer be in the along-channel direction, but rather cross-channel, and will have negligible influence on local generation of along-channel transports in this

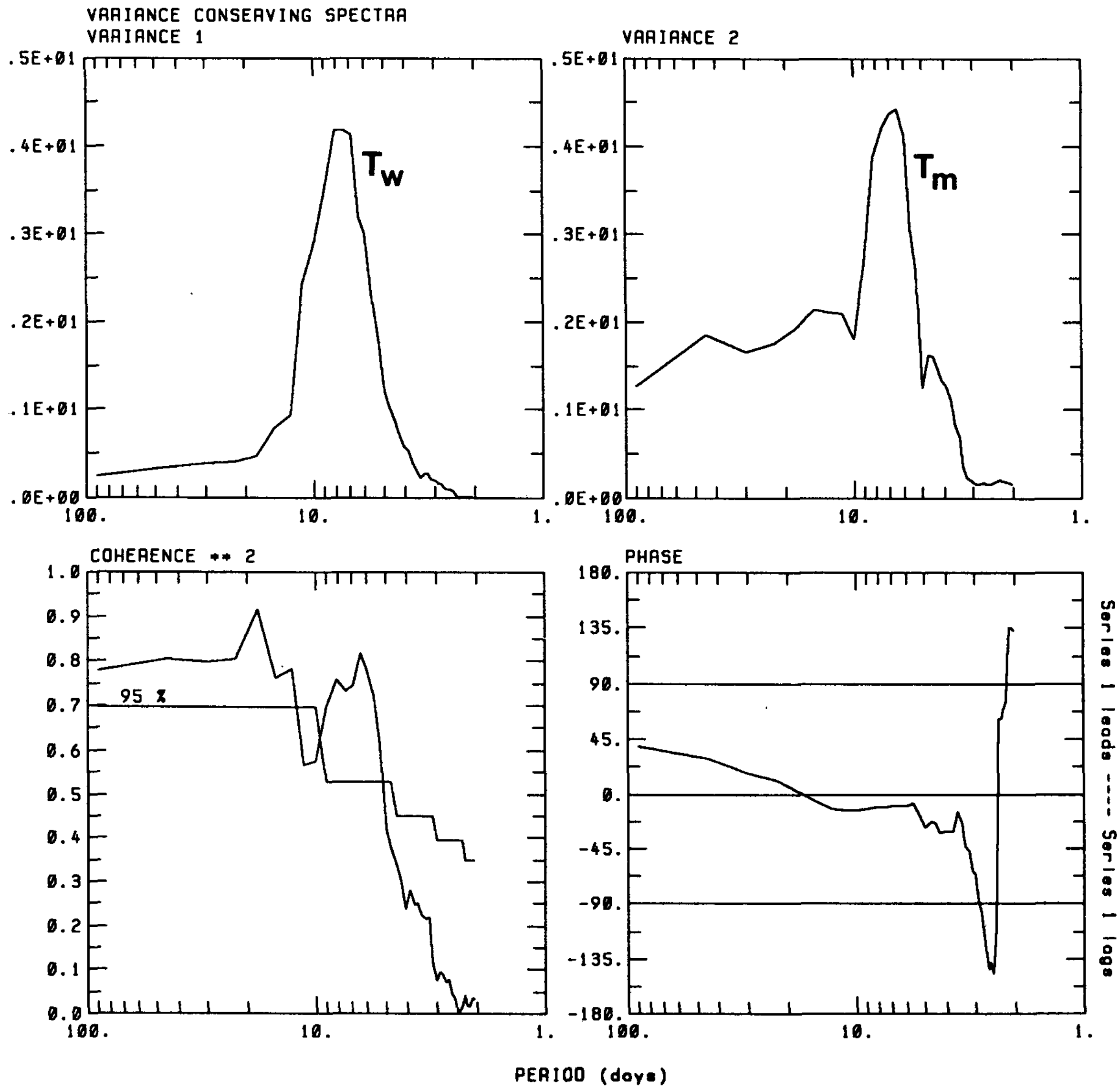

FIG. 7. Variance conserving spectra, coherence squared and phase between transports derived from moored current meters (T_m) and predicted from along-channel wind stress (T_w) for 1 December 1983 to 1 March 1984.

east–west portion of the channel. Thus the along-channel transport response to meridional winds will cease after the response has traveled the effective length of the north–south oriented channel, expressed as $l_y = L_y/2$, where L_y is the actual length of the north-south portion of the channel, 220 km. The effective length scale corresponds to the region inside the channel where the assumption of infinite length is more appropriate.

The wind-forced transport response will be advected through the channel at approximately the rate of the cross-sectional averaged mean downstream current $\langle v \rangle$, which is 75 cm s^{-1} at the STACS section where the mean northward transport is 32 Sv. Therefore, the time required for the wind-forced transport response to be advected through the effective length of the north–south oriented portion of the Straits of Florida is $l_y/\langle v \rangle$, or approximately 40 hours, which is quite close to the 48-hour integration time that was found to produce the best model results. A similar result was arrived at by Boudra et al. (1987) using an isopycnic coordinate numerical model of the Straits of Florida with the Florida Current forced by meridional winds varying sinusoidally in time. They conclude that the one-day transport lag to the wind forcing is too short to be explained by friction, and suggest that the fluid residence time in the Straits of Florida could be responsible.

Transports were modeled for the total STACS period of moored measurements from April 1982 to June 1984 using the Palm Beach winds and the 48-hour in-

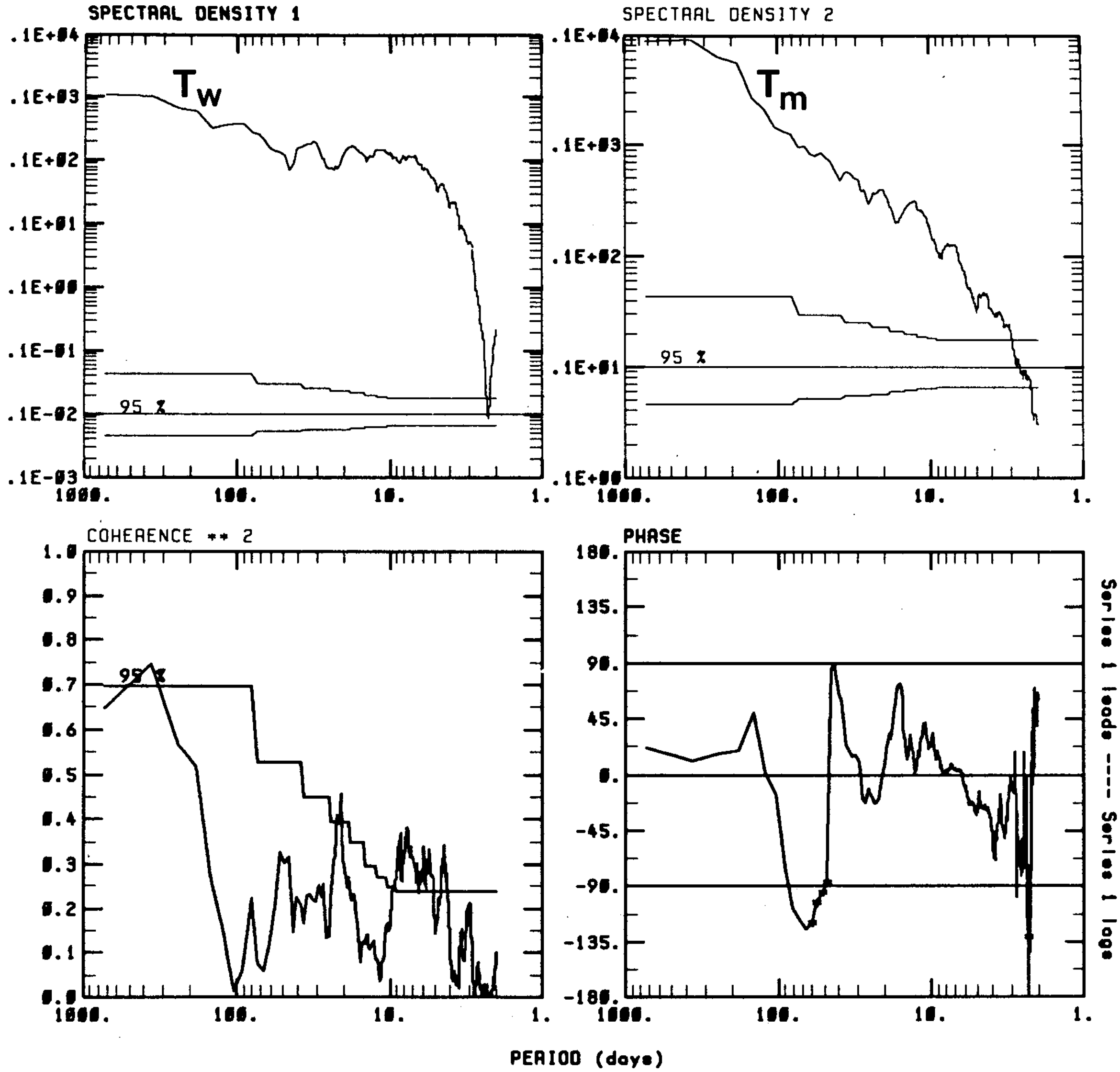

FIG. 8. Spectral density, coherence squared and phase between transports and transports predicted from the along-channel winds (T_w) for the 2-year period 9 April 1982 to 9 April 1984.

tegration time discussed previously. The comparison over the combined 2 years of measurements shows significant coherence between modeled and measured transports in the 4 to 10 day weather band and slight coherence at 20-day period (Fig. 8). These are also the periods where coherence was highest between measured transports and local meridional winds. The spectral shapes of the measured and predicted transports are somewhat different. The measured transports have a near continuous decrease of energy with frequency at a slope of −1.5 and several narrow peaks centered at periods of 4, 7, 12 and 20 days. Spectra of predicted wind forced transports are similar to that of the along-channel wind for the same period with higher energy levels in a broad band of about 3 to 20 days. Energy levels fall off rapidly at periods less than 2.5 days, due to the 2-day integration used. Johns and Schott (1988) used EOF analysis to show that energy peaks at periods of 4, 7 and 20 days in the STACS-V data could be accounted for by along-channel wind forcing, whereas a meandering mode occurred at 5 and 12 day periods.

Best modeling results were obtained during periods of strong along-channel wind forcing such as the winter of 1983/84 (Figs. 6 and 7). The winter of 1982/83 showed significant coherence between modeled and moored transports at periods of 4 to 5 days but at lower significance levels than during 1983/84. During the summer seasons 1982 and 1983, there was no significant coherence found. This may result because summer winds tend to be weaker than winter and have more of a cross-channel component. At times of weak wind forcing other processes causing transport variations, such as meandering and instabilities, can make it difficult to observe the wind forced response.

Figure 8 gives an indication of coherence between modeled and moored transports at the annual period in addition to the 4–10 day weather band. This suggests that the strong southward along-channel wind events that occur in the fall, together with the prevailing moderate northwestward winds in the summer, may combine to contribute locally to the annual transport cycle. To investigate this possibility we demeaned the modeled and measured transports (from both moored and cable estimates) and computed monthly averages (Fig. 9a). The annual transport cycle from monthly averaged model results agrees well in phase with the measured transports but the amplitude is about a factor of 2 too small. If we integrate the model over 5-day periods before the monthly averages then the amplitude increases to match the measured values and the phase remains unchanged (Fig. 9b). This suggests that the contribution to the annual FC transport cycle from wind forcing occurs on longer time scales and over a larger region than the north–south portion of the Straits of Florida where local synoptic scale winds have been shown to be important. The indication is that the wind forcing should be in the same direction for 5 days or longer if this result is to be meaningful. Indeed along-channel wind forcing does appear to occur at longer periods during the summer and fall seasons when the greatest contribution to the annual transport cycle occurs (Schott et al. 1988).

5. Conclusions

A simple wind-forced coastal response model was applied to the Straits of Florida to show that local along-channel winds from synoptic scale weather events can generate significant fluctuations in FC volume transport. A northward wind can cause an increase in transport and a southward wind a decrease. This is a barotropic process where uniform along-channel flow is accelerated due to cross-channel interior flows adjusting to surface Ekman transports in a channel where there is a two-sided coastal constraint ($U = 0$ near the coasts) which causes the simultaneous occurrence of divergence and convergence at the channel boundaries. The sea surface then adjusts geostrophically to the developing along-channel transport. Bottom friction appears to be insignificant so that the transport response will continue until the wind forcing changes. However, the finite length of the north–south domain of the Straits of Florida appears to have an effect similar to that of bottom friction in shallow shelf regions, for the fluid has no memory of the wind forcing that occurred on time scales longer than the mean advection time of the north–south portion of the channel, which is approximately 2 days due to the strong northward mean transport.

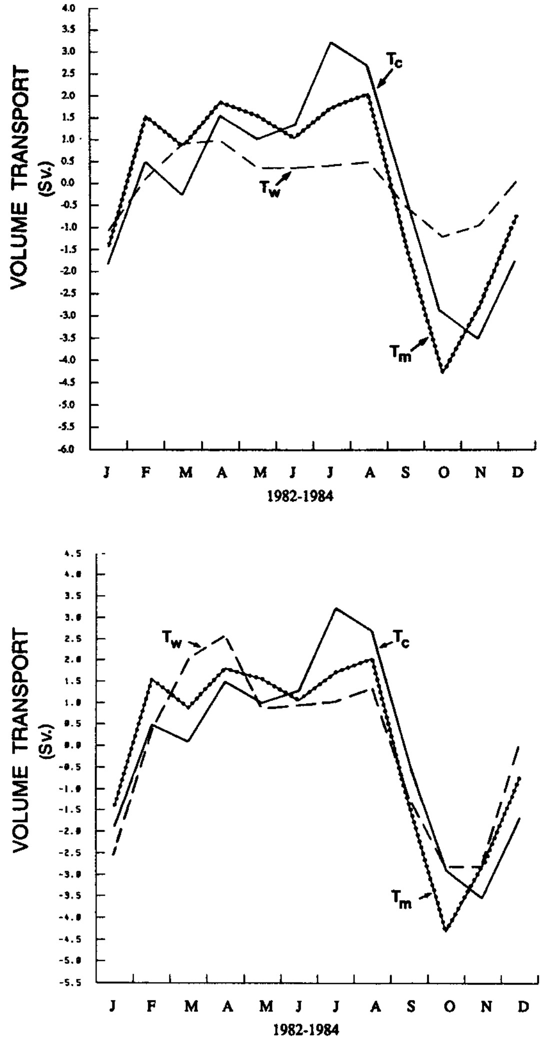

FIG. 9. (a) Monthly averaged volume transport fluctuations during 1982 to 1984 derived from moored current meters (T_m: dotted line), cable voltages (T_c: solid line) and predicted from local along-channel winds (T_w: dashed line) using a 48-hour integration time. (b) As in (a) except for a 5-day integration time.

This barotropic fluctuation, together with the cross-channel interior flows, can cause a steepening of the pycnocline as seen by the positive correlation between transport and horizontal temperature gradients across the front (Fig. 5). The barotropic perturbation and steeper isopycnals within the straits are advected northward by the current where adjustment can take place through baroclinic instability as the disturbed isopycnals convert mean potential energy to perturbation potential energy which converts to eddy kinetic energy and then back to the mean flow. Thus local wind-forced barotropic fluctuations within the Straits of Florida provide a mechanism for meander and eddy growth downstream of the straits. As the flow emerges from the Straits of Florida the effect of the channel and its two-sided constraints are no longer significant.

North–south winds will simply generate surface Ekman transports to the right of the wind with no significant effect on FC volume transport.

The simple local wind-forced model was also used to estimate seasonal changes in transport with surprisingly good results. The results were best when the model was integrated over a 5-day period, which suggests that the wind forcing contribution to the annual FC transport cycle comes from weather events with time scales longer than 5 days and spatial scales larger than the dimensions of the Straits of Florida. If the agreement between modeled and measured transports on annual time scales is not just fortuitous, then it indicates that seasonal differences in along-channel winds over the Straits of Florida from summer to fall provide a significant contribution to the annual transport cycle of the Florida Current. Along-channel winds result from synoptic weather systems that can have spatial scales of several hundred kilometers, so regional wind stress and wind-stress curl patterns over the Caribbean or northwest Atlantic may also contribute to the annual transport cycle. Substantiation of this unexpected finding will require further analysis with longer data records and more comprehensive numerical channel models and ocean circulation models similar to that of Anderson and Corry (1985) where the influence of spatial and temporal scales of the wind forcing fields can be investigated.

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