Seasonal Transport Variations in the Florida Straits: A Model Study

DAVID L. T. ANDERSON AND ROBERT A. CORRY

Department of Atmospheric Physics, Oxford University, England

(Manuscript received 16 July 1984, in final form 2 February 1985)

ABSTRACT

In a previous study Anderson and Corry used a wind-driven two-layer model to study the effects of topography and islands on the seasonal variation of western boundary currents. The work is continued here with topography, geography and winds appropriate to the North Atlantic to examine the seasonal cycle of the Florida Straits transport. A summer maximum of transport is predicted consistent with observations. The areas of importance and processes giving rise to the seasonal cycle are considered.

1. Introduction

The pioneering work of Niiler and Richardson (1973) showed that the Florida Current had a mean transport of 30×10^6 m^3 s^{-1} (30 Sv) and suggested an annual cycle of 4 Sv. These results were based on a series of experiments each lasting a few weeks but spanning several years. The fact that the measurements were not distributed uniformly over the year, combined with the existence of energetic high-frequency fluctuations (Wunsch and Wimbush, 1977; Brooks, 1979), made it difficult to define precisely the annual cycle and to distinguish seasonal from interannual variability. Figure 1 is a replot of the Niiler and Richardson data aimed at visually maximizing the seasonal effect; this includes a maximum in summer and a minimum in fall. There is a gap in the data during August, but this is filled on the figure by later data from Brooks (1979). Recent work by Molinari *et al.* (1985) and Schott and Zantopp (1985) further supports such a seasonal cycle.

The mean transport of 30 Sv is consistent with expectations based on integrating the annual mean wind-stress curl across the basin at latitude 31°N (Leetmaa *et al.*, 1977; Leetmaa and Bunker, 1978). However, integration of the seasonal wind-stress curl across latitude 31°N suggests a seasonal variation of the Florida Straits transport of ±15 Sv with a maximum in winter. The fact that neither strength nor phase of the observed seasonal transport is consistent with the Sverdrup relationship suggests that dynamics governing the mean flow must be different from those governing the seasonal variation.

Anderson and Corry (1985) have investigated the dynamics of the seasonal variation of western boundary currents. They showed that the Sverdrup balance is unlikely to hold at annual period for the North Atlantic. For periods much less than the time taken for the wind-generated baroclinic Rossby waves to pass over bottom topography, the ocean response is primarily that for a homogeneous ocean and thus strongly modified by topography. For periods much longer than this time, the Rossby waves compensate for the effect of topography and the nontopographic Sverdrup balance holds. For the Atlantic at 25°N it can take from years to decades for compensation by baroclinic Rossby waves to occur, depending on the distance of the wind stress variability from the western boundary. This means that at annual period the Sverdrup balance is not likely to be applicable. At such periods (annual or less), variations in transport can be due to barotropic Rossby waves generated by $\mathrm{curl}_z(\tau/H)$ where τ is the wind stress and H is the ocean depth. This can be split into two parts, one resulting from the curl of the wind stress and the other from gradients in topography. It is important to note that transport variations can be forced even in the absence of wind stress curl. Since the streamlines follow roughly f/H contours, the Florida Straits transport variations could result from wind forcing to the northeast. In addition to the barotropic mechanism, Anderson and Corry (1985) showed that a baroclinic Kelvin wave generated farther north could induce transport variations as it passes over topography.

In this paper we wish to model the seasonal cycle of transport through the Florida Straits, to find the areas of wind stress which induce the seasonal variation, and to compare the model results with observations. While emphasis is given primarily to the Florida Straits (Section 3), some consideration is also given to other regions such as the Antilles Current region, the Caribbean, the Gulf Stream and an area to the east of the Amazon (Section 4).

2. The ocean models and the wind stress forcing

Two models are used. The first is linear and has two layers in the vertical. The model equations and

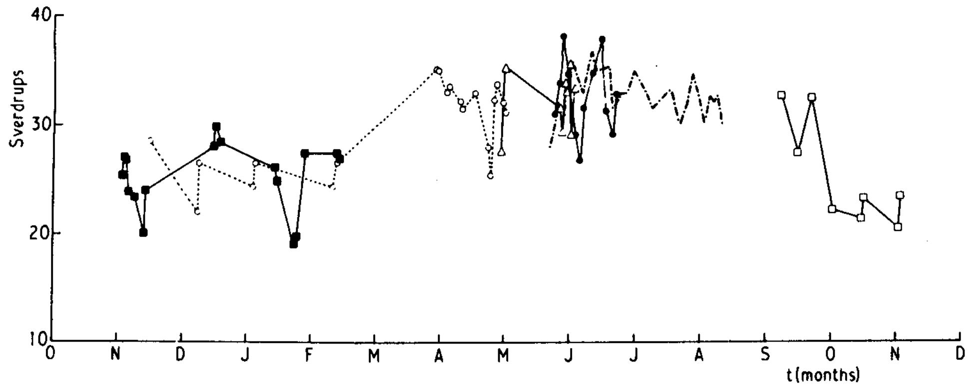


FIG. 1. A replot of the Niiler and Richardson (1973) data aimed at visually maximizing the seasonal effect. Later data from Brooks (1979) are also included on the dot-dash line. The other symbols correspond to data sequences.

method of solution are the same as described in Appendix A of Anderson and Corry (1985) except that the equations are for spherical geometry rather than the β-plane. Two types of friction are included: Laplacian friction with a mixing value of 10^4 m^2 s^{-1} and bottom friction with a damping time of 100 days. Continuous bottom topography can be readily incorporated provided it is confined to the lower layer. In Appendix A the bottom topography actually used and the choice of islands is discussed. Realistic geometry is used as far as possible at a resolution of one degree. The domain extends from 10°S to 50°N. Free slip boundary conditions are used throughout. In addition, a single layer barotropic model, with or without bottom topography, is used to assist in the interpretation of results. The baroclinic model is integrated for five years, but the barotropic model only for shorter times.

The wind stress data were prepared on a one-degree grid by Hellerman from the data compiled by Bunker and Goldsmith (1979), and consist of averages for each month over several years. The annual mean wind stress is shown in Fig. 2. Since the model to be used is linear, we can separate the seasonal variations of wind stress from the mean wind stress and consider the effect of each separately. The seasonal wind stresses are the monthly means after the annual mean has been subtracted. The transport values and stream-function patterns which will be presented are those at the end of each month's integration. There was very little difference between the fourth and fifth years of integration at all latitudes, and virtually no

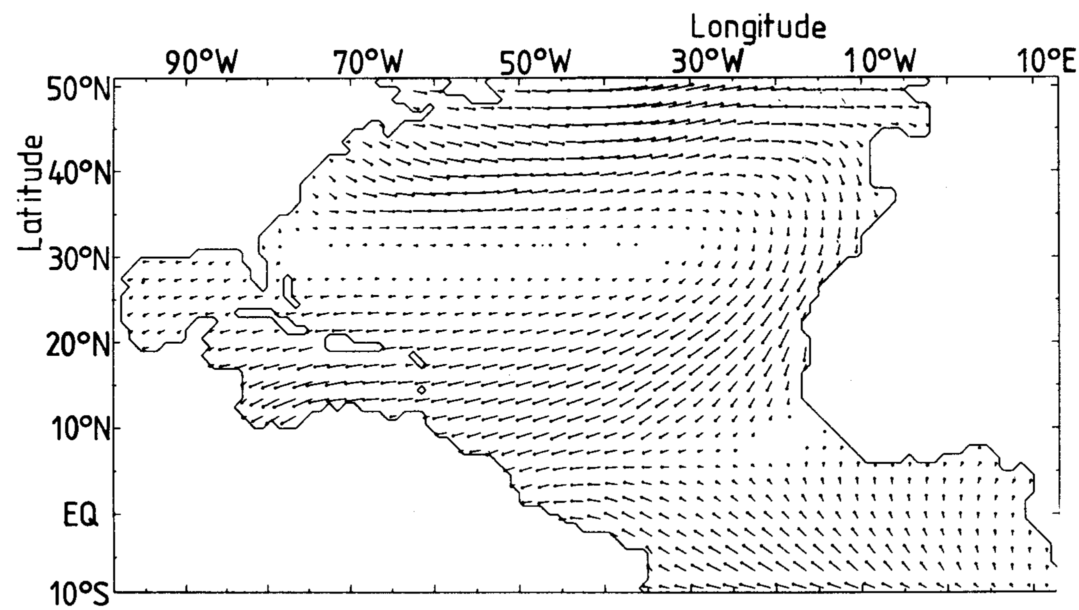


FIG. 2. The annual mean wind stress. The space between arrowtails is equivalent to 0.1 N m^{-2}.

difference in the transport of the Florida Straits between the first and fifth years of integration, confirming a result of Anderson and Corry (1985), that wind-forced baroclinic Rossby waves have very little effect on the seasonal variations of transport at these or higher latitudes.

Figure 3 shows the streamfunction throughout the basin during the fifth year of integration of the baroclinic model. For comparison we show in Fig. 4 the equilibrium streamfunctions for January, April, July and October for the case of a flat-bottomed ocean. In the latter case the western boundary transports are close to the Sverdrup values, and thus reflect the curl of the wind stress to the east.

The models used in this paper are more appropriate for studying variability than the mean state, but for completeness, the results for the mean wind stress will be considered in Section 5, after discussing the seasonal variation in Sections 3 and 4.

3. The seasonal variation of transport through the Florida Straits

a. Transport variations in a baroclinic ocean

Figure 5a shows the predicted variation of transport through the Florida Straits. The variation in transport is small (a peak to peak variation of 3.5 Sv), with a maximum at the end of July and a pronounced decrease to a minimum at the end of October. These predicted variations appear to agree in phase with the annual cycle of Fig. 1, though the magnitude is roughly a factor of 2 too small. For comparison Fig. 5b shows the predicted transports through the model straits in the absence of bottom topography. We see here that the peak-to-peak variation in transport is greater than 30 Sv, and that the maximum in transport is now in winter. The great difference between Figs. 5a and b shows the importance of bottom topography for the Florida Straits transport at annual period. Comparison of Figs. 3 and 4 further emphasizes this fact.

We can determine the transport induced by the passage of the baroclinic signal over topography if we subtract the transport variations for a homogeneous ocean from those for the baroclinic ocean. Figure 5c shows the Florida Straits transport variations for a homogeneous ocean with bottom topography. Since we have shown in Section 2 that baroclinic Rossby waves are not important in modulating the seasonal transport at the latitude of the Florida Straits, most of the difference between Figs. 5a and c, plotted as Fig. 5d, is probably a result of the passage of the baroclinic Kelvin wave as suggested by Anderson (1979). Though Fig. 5d does not exhibit an outright summer maximum, baroclinic effects are shown to increase the peak-to-peak variation between July and October.

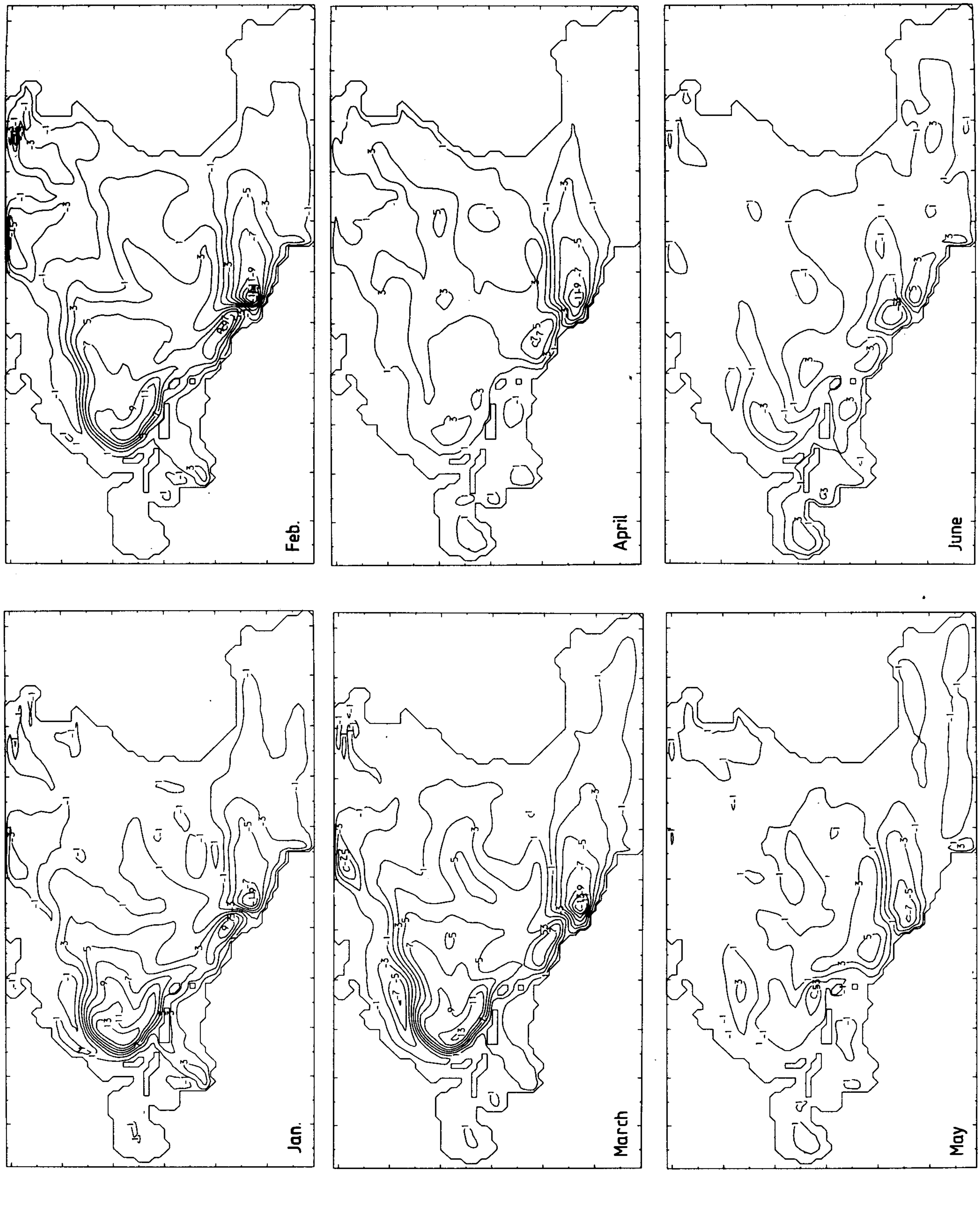
In order to assess the importance of varying geometry on the seasonal cycle of the Florida Straits transport, two further integrations were performed, each for one year. In the first, the passage between Cuba and the island representing the Bahamas was blocked, and in the second the Windward Passage (see Fig. 8) was blocked. The Florida Straits transport for each of the baroclinic calculations is shown in Fig. 6. There is little difference to Fig. 5a, suggesting that neither of these passages is important for the seasonal variation of Florida Straits transport.

b. Transport variations in a homogeneous ocean

We consider here the separate effects of zonal and meridional components of wind stress for the case of a homogeneous ocean with bottom topography. Figure 7a shows the transport variation predicted using the zonal wind stress only, while Fig. 7b shows the corresponding plot for the meridional wind stress. The meridional wind stress reduces the size of the December maximum expected from the zonal wind stress, and creates the overall summer maximum of Fig. 5c. The pronounced annual cycle of Fig. 7b is in contrast to the large semiannual signal of Fig. 5c, which has maxima at the end of July and December, and minima at the end of October and March.

The results of Anderson and Corry (1985) suggest that because of the strong topographic effect of the Mid-Atlantic Ridge on transport variations, only the wind stress in the western half basin is likely to be important in forcing seasonal variation in Florida Straits transport. The importance of winds to the west is further enhanced by the fact that most of the seasonal wind change occurs there. To determine the relative importance of different regions for the seasonal variation of the Florida Current we consider two forcing regions marked A and B on Fig. 8. This is not done in the case of the baroclinic model because the wind stress itself is needed to force the baroclinic mode, and restructuring the region of wind stress introduces spurious effects at the edges of the regions. The barotropic model, however, is driven by $\mathrm{curl}_z(\tau/H)$, and it is this which is split up into regions A and B rather than the wind stress.

The contributions to Florida Straits transport forced by the two areas are shown in Figs. 9a, b. The contribution from areas other than A and B (not shown) was negligible during summer, and less than 0.2 Sv at other times of the year. Region A to the north of the Bahamas and to the west of the mid-Atlantic ridge, accounts for a large part of the variation in transport of the Florida Current. While the forcing of region B results in a smaller Florida Straits transport than that of region A, it is none the less in good phase agreement with the observed transport (Fig. 1). It is possible to speculate that processes not contained in this model could act to enhance the Florida Straits


Feb.
April
June
Jan.
March
May

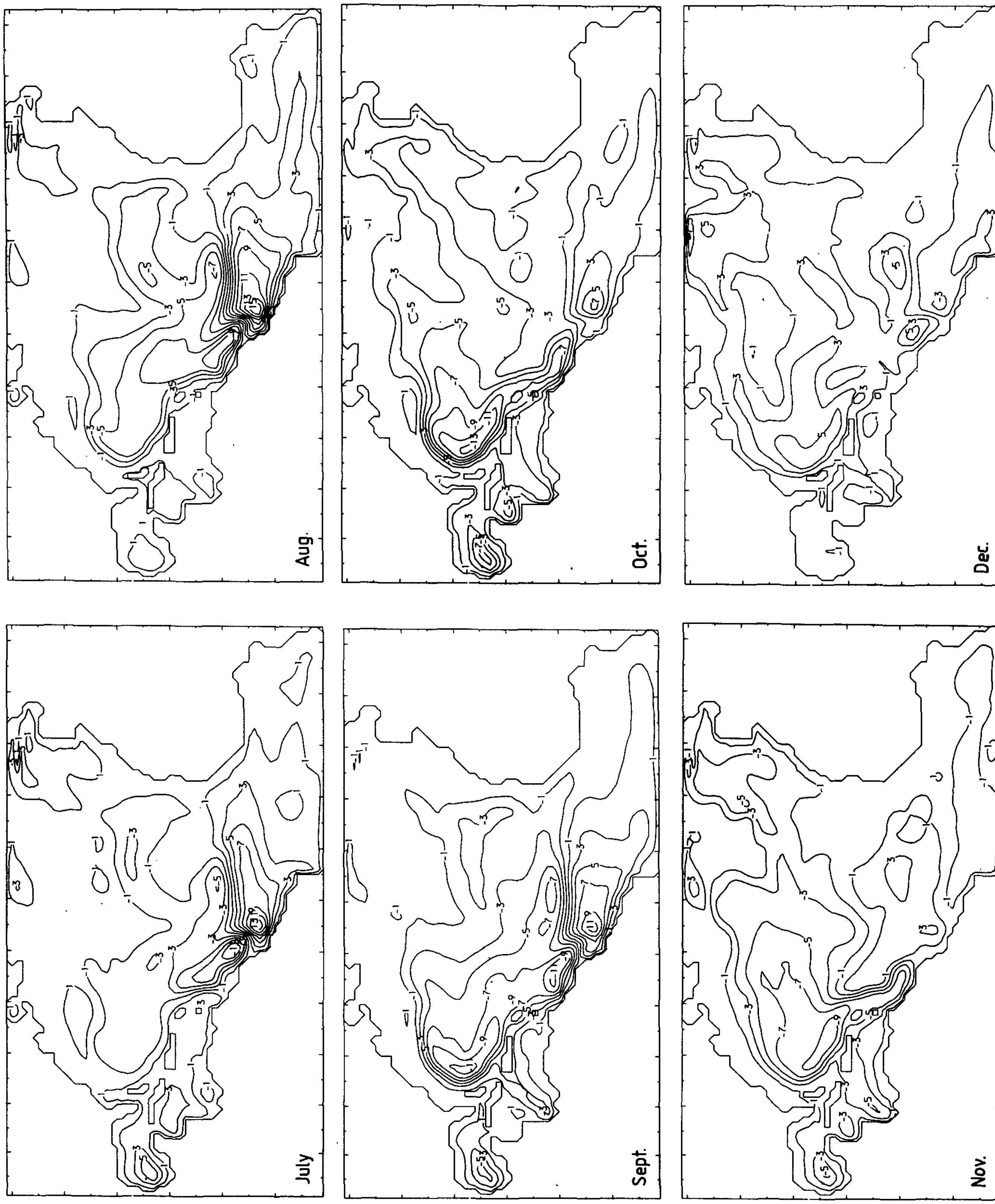


FIG. 3. The streamfunctions during the fifth year of integration when the seasonal wind stresses for January to December are used. Bottom topography is smoothed as in Appendix A.

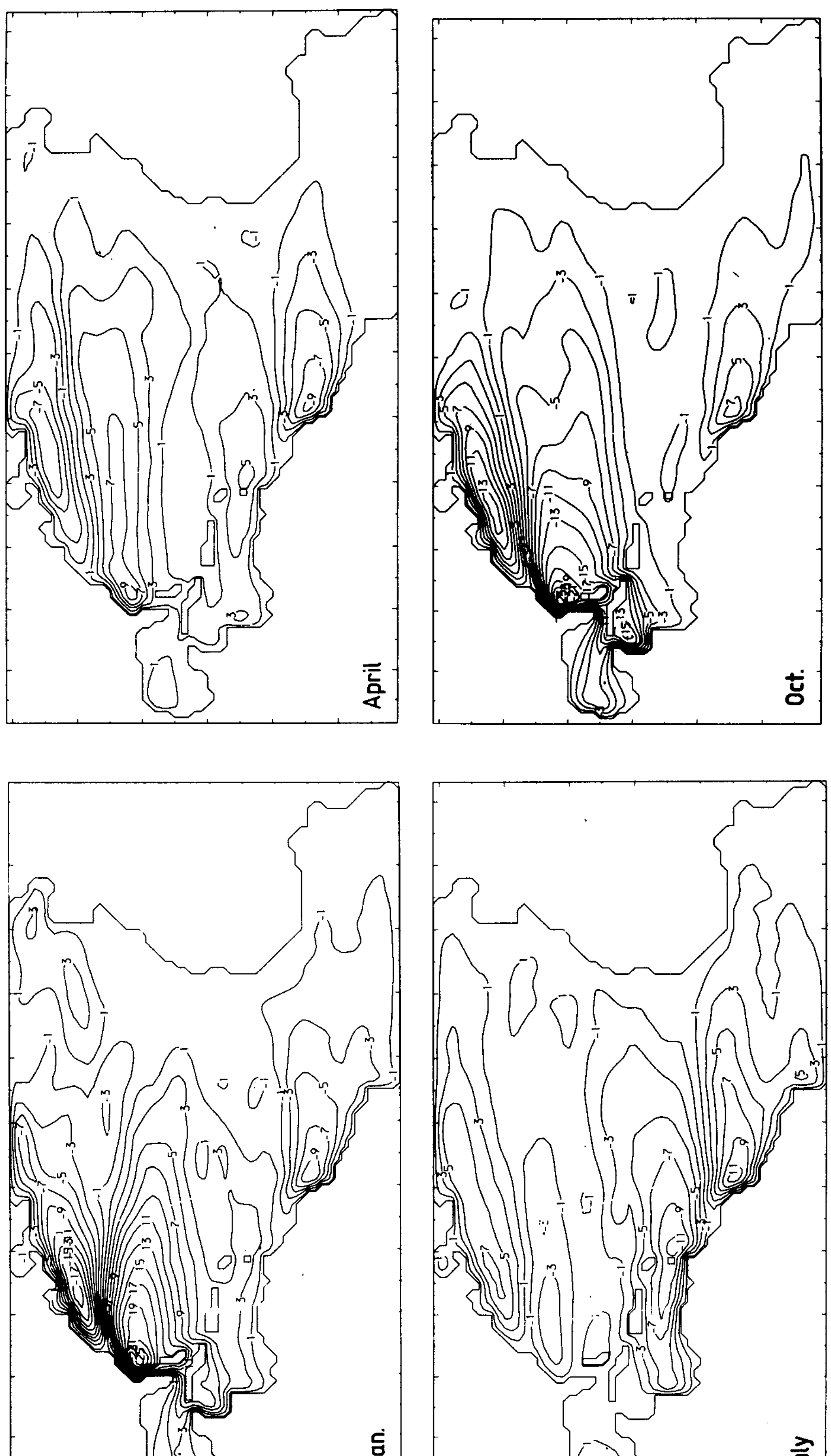


FIG. 4. The resulting streamfunctions when a flat-bottomed model is integrated with the seasonal wind stresses for January, April, July and October. North of Cape Hatteras we see that there is a gyre which has opposite sign to the gyre east of Florida. This is true of every month except December, which lacks the two gyre structure.

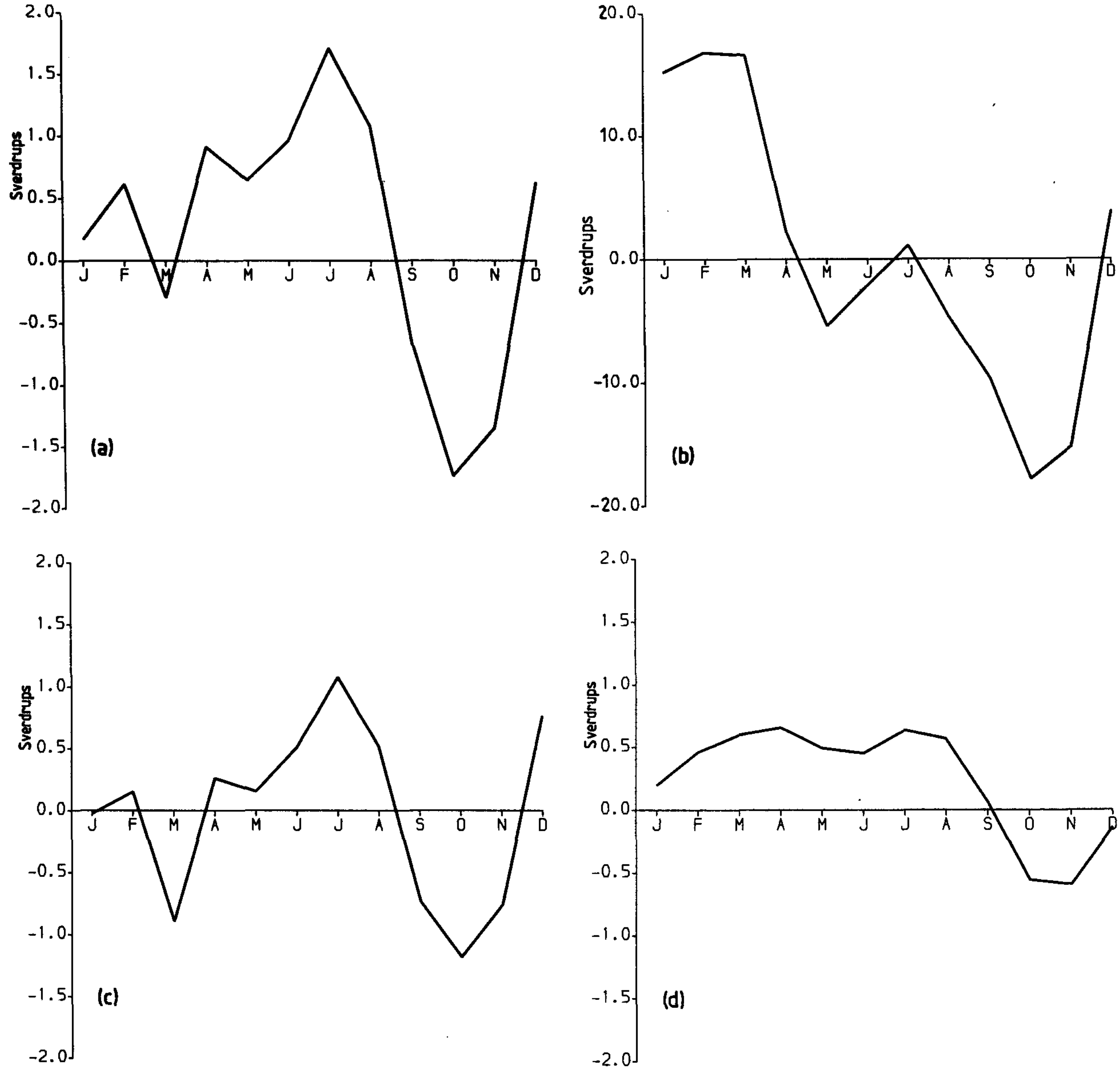


FIG. 5. Transport (Sv) through the Florida Straits when (a) the baroclinic model with bottom topography is forced by the seasonal wind stresses, (b) a flat-bottomed ocean is forced by the seasonal wind stresses [compare with (a)], and (c) for a homogeneous ocean when forced by the seasonal wind stresses. (d) The contribution to Fig. 5a of baroclinic effects over topography, i.e., the difference between (a) and (c).

response to Caribbean forcing. A possible contender is the enhanced bottom torques that could result from downstream advection of the density field over topography. This is a process which needs more careful study, in a model more comprehensive than the adiabatic one used here.

The total transport variation is forced by $\mathrm{curl}_z(\tau/H)$. This can be split into a part depending on the curl of the wind stress and another depending on gradients of topography. Figure 10a shows the contribution due to curl of the wind stress for area A, while Fig. 10b shows the contribution associated with gradients in topography. From these figures we see that the winter contributions roughly cancel, whereas the subsidiary summer maximum of Fig. 10b is augmented by Fig. 10a. Figures 5b and 10a are both curl-generated transports but show very different behavior. The reason for this is that due to topography they are sampling different areas of ocean. In the case of Fig. 5b, the transport is a measure of the curl to the east, whereas in the case of Fig. 10a the transport variations result from curl changes to the northeast. These curl-forced variations in the northwestern Atlantic are deflected southwestward by topography to

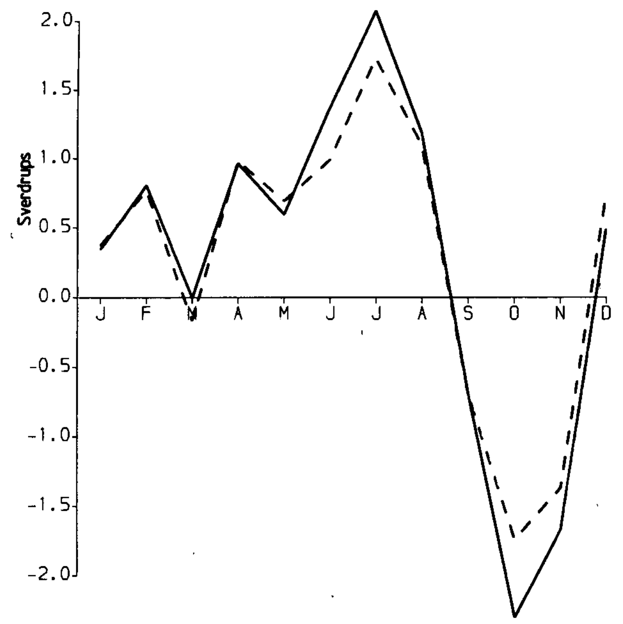


FIG. 6. The transport through the Florida Straits during the first year of integration of the baroclinic model, with the passage between Cuba and the Bahamas blocked (solid line), and with the Windward Passage blocked (dashed line). Compare with Fig. 5a.

affect transport variations in the Florida Current. Examination of Fig. 4 shows that north of Cape Hatteras the curl drives a cyclonic circulation in winter with anticyclonic circulation during the fall. The deflection southward of the streamlines by topography gives rise to the winter minimum and fall maximum of Fig. 10a.

4. Seasonal variation of transport in other parts of the North Atlantic

In this section we consider transport variation in areas of the North Atlantic other than the Florida Current. Regrettably there are as yet no measurements of total transport in these areas with which to compare the model results. Instead, transport measurements are usually based on hydrographic data with an assumed level of no motion. Since the seasonal variation of transport in our model is mainly barotropic, this will not, in general, be well represented in geostrophic transport calculations, so no comparison with observations is attempted in this section.

The interested reader may, however, wish to consult the studies of Gunn and Watts (1982), Olson *et al.* (1984) and Worthington (1977), where some estimates of seasonal variations in the Antilles Current and Gulf Stream are given.

One question of particular interest is whether or not there is an Antilles Current flowing to the east of the Antilles and Bahamas island arc (see Fig. 8). The model seasonal variations of Fig. 3 almost represent total transport in this region since there is little mean transport (see Fig. 14). Between December and March there is a northwestward transport of the order of 10 Sv, from April to July there is little transport and during fall the transport is predicted to be southeastward. The seasonal variability of the Antilles Current is a strong feature of this model. Schott (private communication, 1984), however, is of the opinion that measurements reported by Olson *et al.* (1984) are not consistent with model results. These measurements show a southeastward transport to the east of the Bahamas during April, but not during the fall.

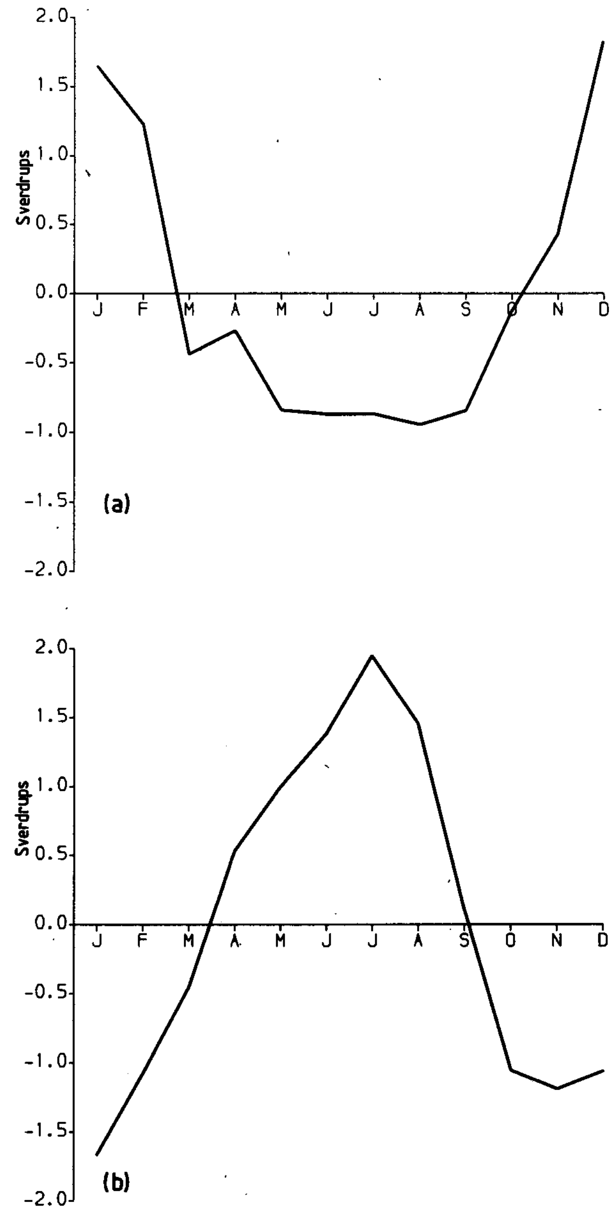


FIG. 7. The transport variation through the Florida Straits when a homogeneous ocean with topography is forced by (a) the zonal component of the seasonal wind stresses and (b) the meridional wind stress.

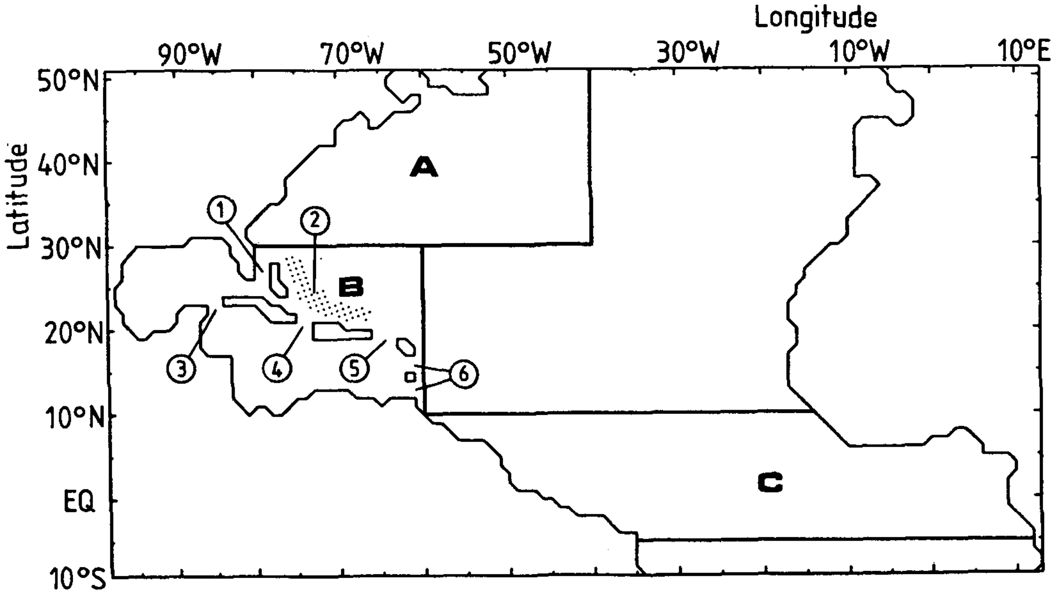


FIG. 8. Areas A, B and C referred to in the text are shown, together with ① The Florida Straits; ② The Antilles Current area (shaded); ③ The Yucatan Straits; ④ The Windward Passage; ⑤ The Anegada passage; and ⑥ The passages of the Lesser Antilles.

It is therefore of considerable interest to know whether or not such variability really exists.

The model variation of transport through the Windward Passage is shown in Fig. 11. During late summer and early fall, water flows into the Caribbean whereas during late fall and winter the flow is outward.

Figure 12 shows the model seasonal transport variation of the Gulf Stream between Bermuda and the North American coast. There is a winter maximum with a minimum in October–November. This is in marked contrast to the phase of the Florida Current variation with its summer maximum.

In Fig. 13 we show the variation of transport through passages of the Caribbean (shown in Fig. 8). The solid line shows the seasonal variation of transport eastward through the Lesser Antilles. During winter there is an increased westward flow, balanced by a flow out of the Caribbean through the other three passages. During late summer and fall, the situation is reversed with a large reduction in the westward flow through the Lesser Antilles balanced again by flow inward through the other three passages. During spring and summer, the situation is more complex, and the flow out through the Yucatan straits has a maximum. It is worth noting that it is only the flow out through the Yucatan Straits (almost equal to the flow through the Florida Straits) that exhibits the summer maximum expected from the data of Niiler and Richardson (1973).

From Fig. A1 we see that the Venezuela and Cayman basins are separated by the Jamaica ridge. This explains the deflection of streamlines toward the Windward passage seen in Fig. 3, whereby streamlines from the east are deflected northeastward when they encounter the South American coast at 12°N, 80°W. This topographic deflection of the seasonal variations is expected since they are primarily barotropic.

In Fig. 3 we can see an intense feature off the mouth of the Amazon. It has a predicted seasonal variation greater than that of the Antilles Current farther north. To its northwest is a similar feature which varies in antiphase, and to its northeast is a zonal current which changes direction during the year. If we force with the seasonal winds over area C of Fig. 8, which includes the tropical belt from 5°S to 10°N, then the southern feature is well reproduced each month, in most cases without the northern feature. It was found that both northern and southern features were forced by curl of the wind stress, rather than by the effect of wind stress over varying topography. The northern feature is forced by winds to the north of area C, with the resulting transport variations deflected southward by topography to enter this area.

5. The mean flow

a. Introduction

Anderson and Killworth (1977) show that for a "switched-on" wind stress the initial effect of bottom topography in reducing the boundary current transport can be compensated for by the passage over the topography of baroclinic Rossby waves. For the baroclinic model used in this study and for time independent forcing, the transport with bottom topography becomes the same as that for a homogeneous ocean with a flat bottom. Since it is quicker to integrate the

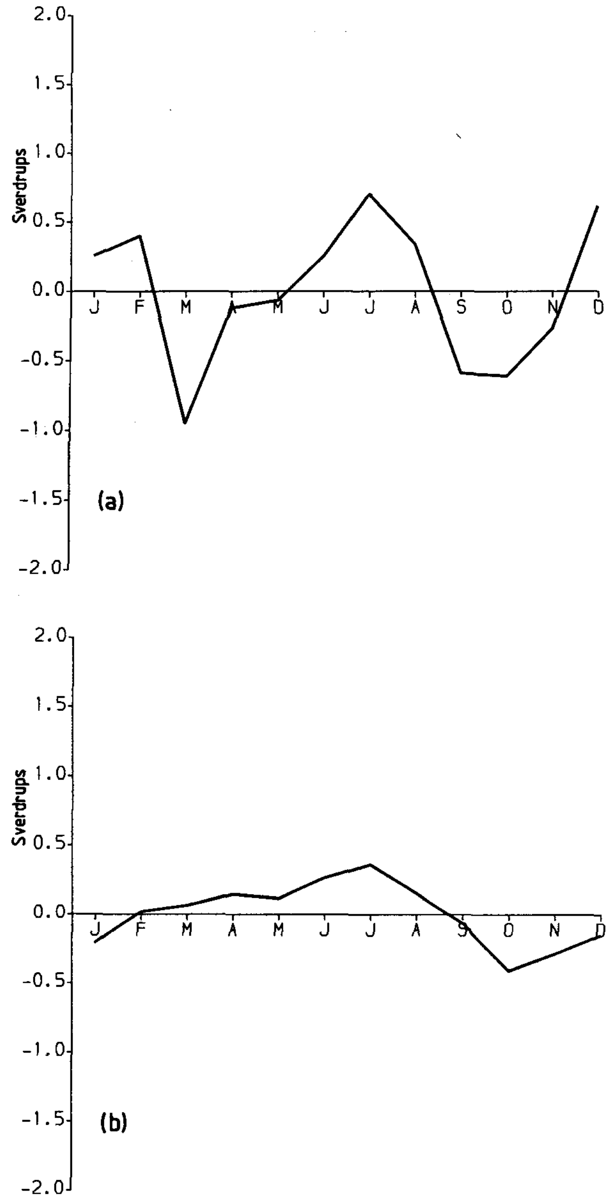


FIG. 9. (a) The contribution of the wind stress over area A to the seasonal variation of the Florida Straits transport in a homogeneous ocean. All the features of Fig. 5c are reproduced, with most of the amplitude. (b) The contribution of the wind stress over area B to the seasonal variation of the Florida Straits transport in a homogeneous ocean. The variation is less than that of Fig. 9a, though still significant.

barotropic equations, the mean streamfunction was obtained by this means.

b. Comparison of the model with data

The model transport through the Florida Straits is 30.5 Sv (Fig. 14), which is in close agreement with the value measured by Niiler and Richardson (1973). Leetmaa *et al.* (1977) show that this value is consistent with the integral of the wind stress curl across the basin at the latitude of the Florida Straits. A consequence of this result is that the Antilles current, to the east of the Bahamas, has no mean northward transport. This conclusion is supported by Gunn and Watts (1982). Figure 14 also indicates little mean transport by the Antilles Current.

To the north, the transport of the model Gulf Stream increases downstream to about 40 Sv. This maximum transport is much less than the observed, which can reach 120 Sv. The disparity between observed and predicted transport indicates the importance to the mean circulation of processes not

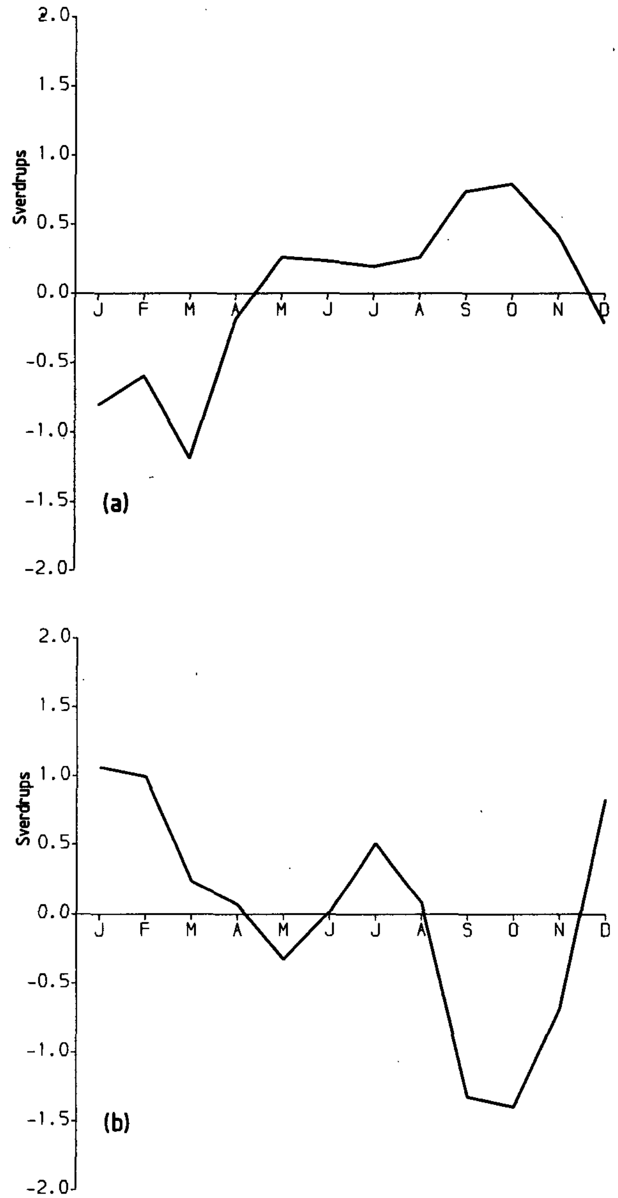


FIG. 10. The contribution to Fig. 9a due to (a) the curl of the wind stress and (b) the gradients in topography associated with the wind stress.

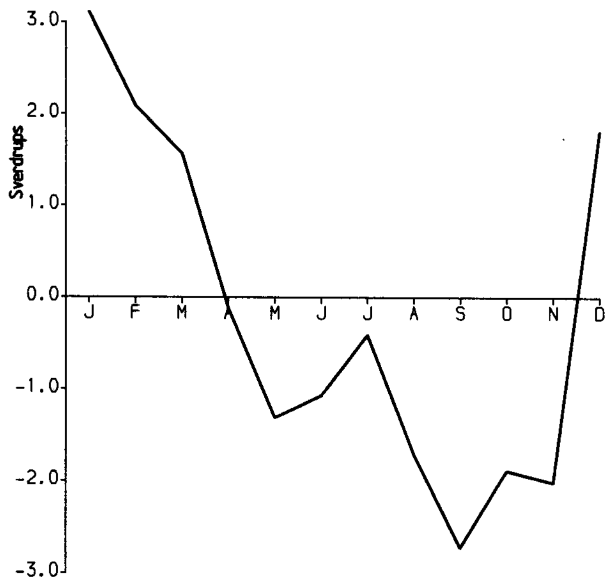


FIG. 11. The northward transport through the Windward Passage for the baroclinic model forced by the seasonal wind stresses.

included in this model, such as eddy driven mean flows (Holland and Rhines, 1980).

To the south, the model shows substantial transport through the passages of the northern Caribbean (see Fig. 8). In particular most of the transport through the Yucatan Straits (22.5 Sv) is contributed by the flow through the Windward passage (12.5 Sv). This is at variance with several authors. For instance, Morrison and Nowlin (1982) suggest that almost all the transport through the Yucatan Straits flows from the eastern Caribbean rather than through the Windward passage. On the other hand, Roemmich (1981) suggests that while the bulk of the transport through the Yucatan Straits comes from the eastern Caribbean, a substantial part (7 Sv) enters via the Windward

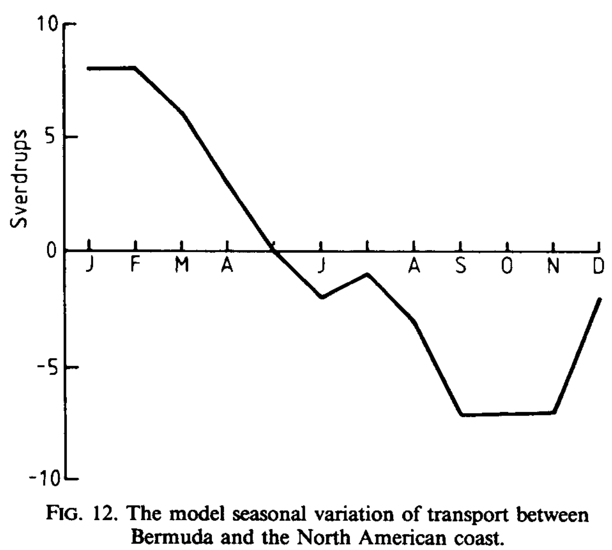


FIG. 12. The model seasonal variation of transport between Bermuda and the North American coast.

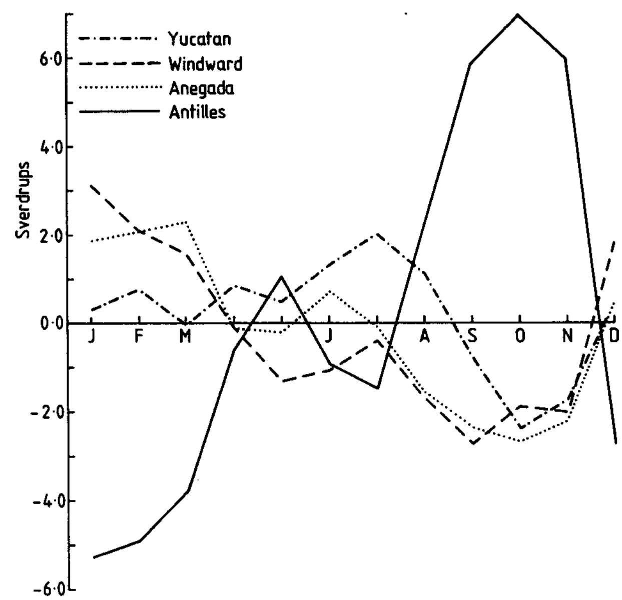


FIG. 13. The flow out of the Caribbean through the Lesser Antilles, together with the flow out of the Caribbean through the Yucatan Straits, the Windward Passage and the Anegada Passage, as calculated from the baroclinic model. See also Fig. 8.

passage. Gunn and Watts (1982) have also measured a substantial geostrophic transport through the Windward passage of 15 ± 5 Sv during the summer of 1972, though it is not clear that this is a reliable measure of the mean transport.

The model shows a mean transport through the Yucatan Straits that is less than the mean transport through the Florida Straits. The highest figure for transport through the Yucatan Straits in the literature appears to be 27 Sv during October 1970 (Hansen and Molinari, 1979). This is 3 Sv lower than the mean flow through the Florida Straits, so it is possible that several Sverdrups do flow westward through passages in the Bahamas to the south of Miami, though probably not as much as the 8 Sv shown in Fig. 14.

In the model Gulf of Mexico a western boundary current of about 7 Sv is evident. This compares well with the 8 Sv estimate of Blaha and Sturges (1981).

6. Summary and conclusion

The seasonal variation of the Florida Current, measured by Niiler and Richardson (1973), has not been understood because application of the Sverdrup relationship predicted seasonal variations of far greater magnitude and almost opposite phase to those measured. The same relationship has, however, been successfully used to predict the mean flow of the Florida Current (Leetmaa *et al.*, 1977). This suggests that there is a difference between dynamics of the mean flow and that of the annual cycle.

Anderson and Corry (1985) investigated the dynamics of the seasonal variation of western boundary currents. They showed that the nontopographic Sverdrup balance is unlikely to hold at annual period for the North Atlantic. For periods much less than the time taken for the wind generated baroclinic Rossby waves to pass over bottom topography, the ocean response is primarily that for a homogeneous ocean and thus strongly modified by topography. For periods much longer than this time, the baroclinic Rossby waves compensate for the effect of topography and the nontopographic Sverdrup balance holds. For the Atlantic at 25°N it can take years to decades for compensation by baroclinic Rossby waves to occur, depending on the distance of the wind stress variability from the western boundary, so at annual period the nontopographic Sverdrup balance is not likely to be applicable.

In Section 3 we investigate the seasonal variation of the transport through the Florida Straits. The total transport variation in the baroclinic calculation was found to be small (±2 Sv) and to agree in phase with observations, although the data show a larger amplitude of ±4 Sv. Part of the seasonal variation results from baroclinic activity over topography. A Kelvin wave, generated by winds to the north of the Florida Straits, generates transport variations as it passes south over varying bottom topography. As expected, the effect of baroclinic Rossby waves at the latitude of the Florida Straits was found to be small relative to the effect of the coastal Kelvin wave. The part of the transport variability which would be forced in a homogeneous ocean can be split into two independent contributions due to the zonal and meridional wind stress. It is the meridional component of wind stress which forces the summer maximum. In contrast, the mean transport is forced primarily by the zonal wind stress curl.

To determine the relative importance of different regions for the seasonal variation of the Florida Straits transport we considered two forcing regions marked A and B (Fig. 8) in the homogeneous model. Region A, to the north of the Bahamas and to the west of the Mid-Atlantic Ridge, accounts for a large part of the variation in transport of the Florida Current, since transport streamlines from this region are deflected southwestward to affect the Florida Current. The variability due to winds over region A can further be split into a part depending on the curl of the wind stress and a part depending on the wind stress over gradients in topography. It was found that in winter the curl-forced variations contributed a minimum in northward transport which roughly cancelled a maximum in the variations forced by wind stress over topographic gradients, and that during the summer both processes added to give the summer maximum. The October minimum is due to the variations forced by topographic gradients.

Although the primary motivation of this paper is to explain the seasonal cycle of transport variations in the Florida Straits, other parts of the North Atlantic are considered in Section 4. Observations of transport in these regions are scarce, and it is not possible to adequately compare model results with observations. In the Antilles region, the model results suggest that if the Antilles Current exists, it is seasonal in nature, with a northwestward transport of about 10 Sv during winter, and a southeastward transport of about 10 Sv during fall. However, more recently Olson *et al.*

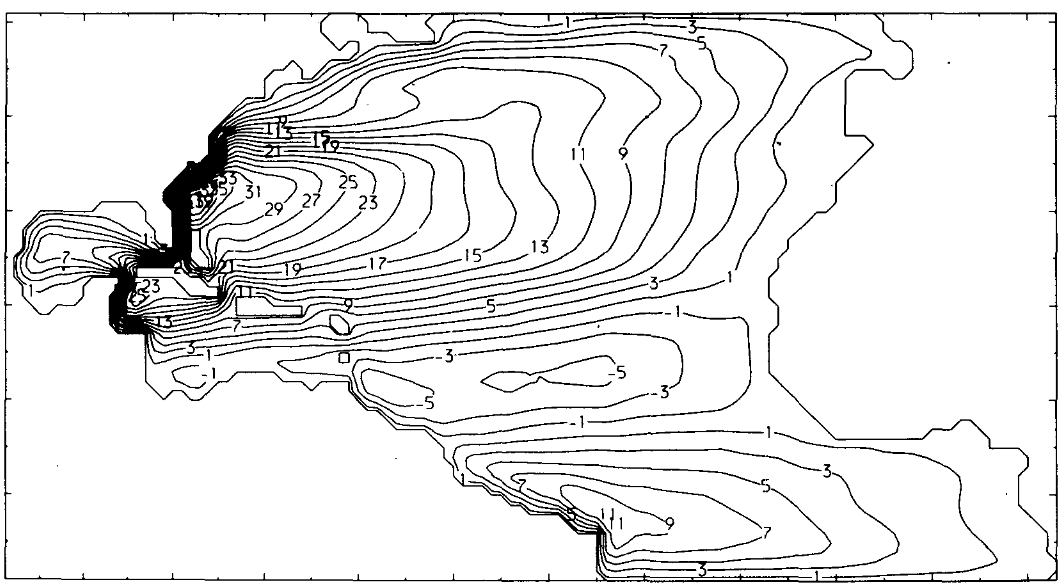


FIG. 14. A plot of the streamfunction forced by the wind stress of Fig. 2 in a flat-bottomed ocean. All the islands of Appendix A are used in this calculation.

(1984) have questioned the existence of a seasonal Antilles Current. The seasonal variation of the Antilles Current is a robust feature of our model. At annual period, only a small part of the seasonal transport passes through the Florida Straits. Instead, most of the transport is to the east of the Bahamas as a result of topographic blocking in the Caribbean and Florida Straits region. It is therefore of particular interest to determine if the seasonal variation in transport suggested by Fig. 3 in the Antilles region really occurs.

APPENDIX A

The Smoothing of the Bottom Topography

The bottom topography was obtained from a file containing world surface topography on a 1° grid. The coastline used was derived from this, and the Mediterranean Sea, the Pacific Ocean and any lakes were removed. The basin (defined with respect to the grid for the streamfunction) extends from 10°S to 50°N, and from 99°W to 13°E. The five islands were to a certain extent manufactured. In particular the Bahamas were represented by a single island, Puerto Rico was joined to the Dominican Republic, thus eliminating the Mona Passage, and the two small eastern islands are intended to represent the Lesser Antilles.

In order to obtain reasonable results from the numerical model, it was necessary to smooth the bottom topography. The way in which this is done is important, because it is the topography which substantially reduces the amplitude of the variation of the Florida Current, and the forcing of the streamfunction (and thus the phase of the transport variations) depends on the gradients of both topography and wind stress.

To start with, it was decided to fix the topography of the Florida Straits, and allow the rest of the ocean bottom to relax into a smoother field. A very simple Laplacian smoother was used such that the increment in depth ΔH was:

$$\Delta H = \delta(HE + HW + HN + HS - 4 \times H)$$

where HE, HW, HN and HS are respectively the most up-to-date depths to the east, west, north and south of the current grid point. Where one of these corresponded to coastline or an island, a value of 400 m was used. At the northern and southern boundaries HN and HS took the value H. The extent of the smoothing, determined by the size of δ, was made dependent on the depth. We wished to preserve, as far as possible, the shallowness of the shallow topography, since it is this shallowness which blocks the transport variations near the western boundary, but at the same time smooth the data. The degree of smoothing was therefore increased with depth. Thus, close to a topographic ridge, the deeper points decreased their depth more than the shallow points increased their depth. For depths greater than 1500 m, the factor δ was given by

$$\delta = 0.1 \times (H/4000 \text{ m})^2.$$

On the other hand, there are areas, e.g., the Florida Shelf, where the topography is too shallow to be handled by the model. For depths less than 1000 m, δ simply took the value of 0.1. Between 1000 and 1500 m δ was $9/640$, the value given by the above formula at 1500 m. The above smoothing operation

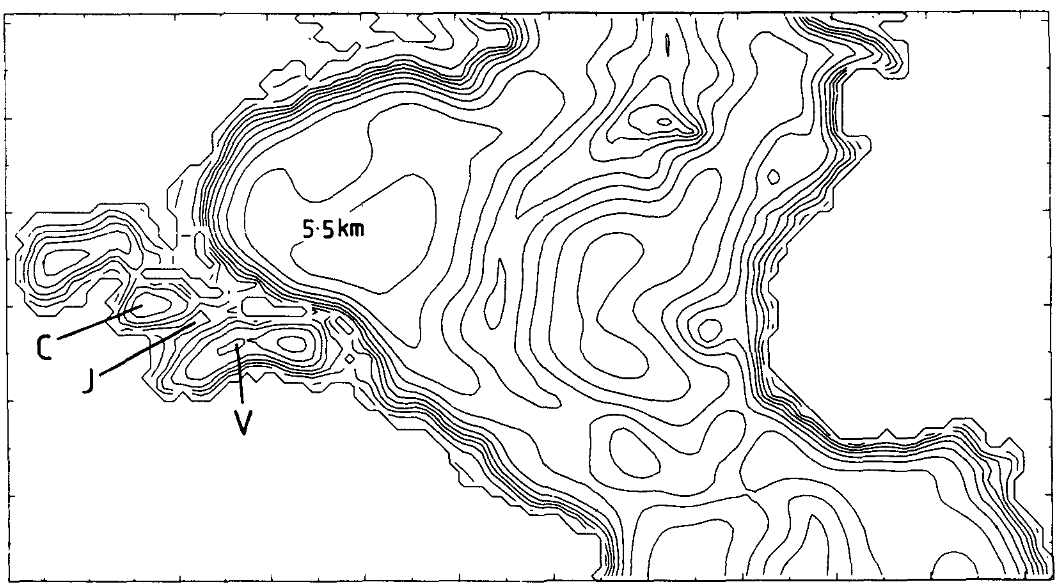


FIG. A1. A plot of the North Atlantic bottom topography used. The contour interval is 500 m. The Venezuala basin (V), the Cayman basin (C) and the Jamaica ridge (J) are also shown.

was applied ten times. Any depths that were then still less than 400 m were set equal to 400 m. The depth of the North Atlantic after this smoothing operation is shown in Fig. A1. The contour interval is 500 m.

REFERENCES

Anderson, D. L. T., 1979: Basin models: The general circulation of regions of the world ocean. *Dyn. Atmos. Oceans.*, **3**, 345–371.

——, and P. D. Killworth, 1977: Spin-up of a stratified ocean, with topography. *Deep-Sea Res.*, **24**, 709–732.

——, and R. A. Corry, 1985: Ocean response to low frequency wind forcing with application to the seasonal variation in the Florida Straits-Gulf Stream transport. *Progress in Oceanography*, Vol. 14, Pergamon, 7–40.

Blaha, J., and W. Sturges, 1981: Evidence for wind-forced circulation in the Gulf of Mexico. *J. Mar. Res.*, **9**, 711–734.

Brooks, I. H., 1979: Fluctuations in the transport of the Florida Current at periods between tidal and two weeks. *J. Phys. Oceanogr.*, **9**, 1048–1053.

Bunker, A. F., and R. A. Goldsmith, 1979: Archived time-series of Atlantic Ocean meteorological variables and surface fluxes. Woods Hole Oceanographic Institution Tech. Rep. WHOI 79-3, 27 pp.

Gunn, J. T., and D. R. Watts, 1982: On the currents and water masses north of the Antilles/Bahamas arc. *J. Mar. Res.*, **40**, 1–18.

Hansen, D. V., and R. L. Molinari, 1979: Deep currents in the Yucatan Strait. *J. Geophys. Res.*, **84**, 359–362.

Holland, W. R., and P. B. Rhines, 1980: An example of eddy-induced ocean circulation. *J. Phys. Oceanogr.*, **10**, 1010–1031.

Leetmaa, A., and A. F. Bunker, 1978: Updated charts of the mean annual wind stress, convergences in the Ekman layers and Sverdrup transports in the North Atlantic. *J. Mar. Res.*, **36**, 311–322.

——, P. P. Niiler and H. Stommel, 1977: Does the Sverdrup relation account for the mid-Atlantic circulation. *J. Mar. Res.*, **35**, 1–10.

Molinari, R. L., W. D. Wilson and K. D. Leaman, 1985: Volume and heat transports of the Florida Current: April 1982 to August 1983. *Science*, **227**, 295–297.

Morrison, J. M., and W. D. Nowlin Jr., 1982: General distribution of water masses within the eastern Caribbean sea during the winter of 1972 and the fall of 1973. *J. Geophys. Res.*, **87**, 4207–4229.

Niiler, P. P., and W. S. Richardson, 1973: Seasonal variability of the Florida Current. *J. Mar. Res.*, **31**, 144–167.

Olson, D. B., F. A. Schott, R. J. Zantopp and K. D. Leaman, 1984: The mean circulation east of the Bahamas as determined from a recent measurement program and historical XBT data. *J. Phys. Oceanogr.*, **14**, 1470–1487.

Roemmich, D., 1981: Circulation in the Caribbean Sea: A well resolved inverse problem. *J. Geophys. Res.*, **86**, 7993–8005.

Schott, F. A., and R. J. Zantopp, 1985: On the seasonal and interannual variability of the Florida Current: Seasonal and interannual variability. *Science*, **227**, 308–311.

Worthington, L. V., 1977: Intensification of the Gulf Stream after the winter of 1976–77. *Nature*, **270**, 415–417.

Wunsch, C., and M. Wimbush, 1977: Simultaneous pressure, velocity and temperature measurements in the Florida Straits. *J. Mar. Res.*, **35**, 75–104.