The Influence of Deep Mesoscale Eddies on Sea Surface Temperature in the North Atlantic Subtropical Convergence

ARTHUR D. VOORHIS AND ELIZABETH H. SCHROEDER
Woods Hole Oceanographic Institution, Woods Hole, Mass. 02543

ANTS LEETMAA
Atlantic Oceanographic and Meteorological Laboratory, NOAA, Miami, Fla. 33149
(Manuscript received 24 February 1976, in revised form 8 July 1976)

ABSTRACT

Maps of sea surface temperature in the North Atlantic subtropical convergence during the 1973 MODE field experiment (and recent satellite imagery) show large meridional and zonal features on a scale of 40-400 km which are superimposed on the seasonal meridional temperature gradient. After comparing these maps with dynamic topography relative to 1500 d\(\), it is argued that these features are mainly due to advective distortion by surface currents associated with the deep baroclinic mesoscale eddy field. Wind-induced surface currents appear to have a lesser effect in generating such structure. Surface frontogenesis observed during MODE and by earlier workers in the area suggests that jet-like shallow surface density currents may be also significant in advecting and distorting the surface temperature field on scales of 10 km and less. Finally, rough calculations indicate that these advective processes of the sea surface may supply annually an amount of heat to the surface water mass of the northern Sargasso Sea which is significant compared with that lost to the atmosphere.

1. Introduction

An important goal of contemporary oceanography is to understand the horizontal distributions of properties at the sea surface and the mechanisms that produce them. Major programs such as the North Pacific Experiment (NORPAX) in the United States and the Joint Air-Sea Interaction Experiment (JASIN) in the United Kingdom are actively working in this area. Although progress has been made in modeling the vertical structure of the upper ocean, much remains to be done in modeling its horizontal structure, particularly over intermediate oceanic scales of 100 to 500 km. Presented here are observations in the subtropical convergence of the western North Atlantic which provide a reasonably detailed look at one aspect of this problem.

The subtropical convergence is one of the classical transition zones (Wüst, 1928) separating two meteorological regimes. In the western North Atlantic it lies roughly between 22\(^\circ\)N and 32\(^\circ\)N latitude and separates the prevailing westerlies to the north from the easterly trades to the south. Maps of monthly mean sea surface temperature in this zone are relatively simple eastward of the Gulf Stream's influence. This can be seen in Fig. 1. In general, the temperature decreases northward at all times of the year by an amount which varies seasonally. The maximum meridional gradient occurs in late winter (approximately 0.5\(^\circ\)C per degree of latitude) and the minimum in late summer (approximately 0.1\(^\circ\)C per degree of latitude). The large-scale zonal temperature variation is small.

How does the synoptic temperature distribution, that is, the actual temperatures at any one time, differ from the above average picture? Surveys of thermal fronts in the area (Voorhis, 1969) suggested that major differences occurred on surprisingly large scales of hundreds of kilometers. More recent evidence comes from satellite infrared imagery of the sea surface, such as shown in Fig. 2. In this image the synoptic temperature is dominated by large meridional and zonal variations on the same scale. What is the reason for this large-scale structure?

Except for the Aries data (Crease, 1962) little was known about the sub-surface currents in this region until the Mid-Ocean Dynamics Experiment (MODE), which was conducted over a period of four months during the spring of 1973 in a 400 km square area centered at 28\(^\circ\)99'N, 69\(^\circ\)40'W. The principle finding from MODE was that subsurface currents were dominated by eddy-like motions, having spatial scales of several hundred kilometers and time scales (residence time) of two to three months, which were

---

\(^1\) Contribution No. 3709 from the Woods Hole Oceanographic Institution.
nearly in geostrophic balance with vertical deformations of the main thermocline. We have collected all useful sea surface temperature measurements made during MODE and constructed maps which, in the following, are presented and compared with the surface geostrophic circulation. From this we argue that most of the large-scale synoptic surface temperature structure in the subtropical convergence is due directly to surface advection by the geostrophic eddy field.

2. The surface temperature field

MODE was designed to study deep currents and water structures and very little effort was expended in surface measurements. In addition, no adequate satellite thermal images of the surface were obtained during the experiment. Nevertheless, surface temperatures were recorded continuously from three ships as they maneuvered about the area during the four months of the experiment. We have used these data plus that from CTD and STD casts to describe the large-scale evolution of the surface temperature field. In order to retain adequate spatial coverage, we chose to group the data in successive time periods of about 15 days. The spatial coverage during the period 31 March to 14 April, which was typical, is shown in the upper half of Fig. 5.

For each period the records of surface temperature were sub-sampled every 10 min to the nearest 0.2°C and plotted along the ship’s tracks. These were combined with temperatures from the STD and CTD casts, which were also used to calibrate the surface temperature records. The total data set was then subjectively contoured by one of us (Schroeder) in half-degree intervals without any prior knowledge of the field of surface currents discussed in the next section. For the most part spurious spatial effects introduced by diurnal heating and cooling along the tracks of the moving ships could be recognized and eliminated in the contouring. Data were rejected, however, on about ten days when afternoon diurnal heating during calm weather exceeded 0.5°C. The resulting maps from 9 March through 13 July are shown in the upper part of each picture in Figs. 3 and 4.

The average surface temperature and its average meridional gradient over the MODE area were computed from each map and these are shown as a function of time in the lower half of Fig. 5. The spatially averaged temperature was close to its late winter minimum (approximately 22°C) at the start of the experiment. Thereafter it increased, due to surface heating, to almost 27°C at the end of the experiment. The average meridional gradient is, perhaps, not too meaningful because of the fluctuations introduced by eddy distortion in the MODE area. Nevertheless, it shows an overall decrease during the experiment from approximately 0.007 to 0.002°C km⁻¹, with cooler water always to the north.

All of the temperature maps in Figs. 3 and 4 show a large, changing zonal and meridional structure superimposed on the mean meridional gradient. Usually, this structure is dominated by long intrusive features or tongues of alternate warm and cool water, 40–50 km wide, which can extend for distances of several hundred kilometers. The resemblance between the temperature pattern in most of these maps and that shown in the satellite image in Fig. 2 is remarkable.

3. Mesoscale eddy surface currents

The mesoscale eddy field found during MODE has been discussed by Robinson (1975), McWilliams (1976), and by other participants in the report of the MODE-I Dynamics Group (1975) cited in the references. The purpose here is to describe the motion of the sea surface in a way which allows one to see its effect on the distribution of surface temperature. Over 800 CTD, STD, and hydrographic lowerings were made during the experiment from the sea surface to depths greater than 2000 m. In addition, cur-
rents were measured extensively at depths below 400 m, primarily by moored current meters and by drifting neutral buoyant floats. These currents, when averaged over a period of several days, have been shown by Bryden (1974) to be in geostrophic equilibrium with the density field within the limits imposed by measurement noise. Furthermore, the vertical structure of these time-averaged currents is highly baroclinic, with most of the eddy energy confined to the main thermocline and above. One concludes, in fact, from the analysis of Schmitz et al. (1976) that about 80% of the geostrophic surface current is, on average, due to density structure above 1500 m. To describe the eddy surface motion, therefore, we computed the dynamic height of the sea surface relative to 1500 db from the CTD and STD data and constructed the maps of dynamic topography (in meters) shown by the solid contours in the lower portion of each picture in Figs. 3 and 4. The mapping time interval was chosen to be the same as that for the accompanying surface temperature maps. For each map the direction of geostrophic current is indicated by the arrows and the magnitude can be estimated from the geostrophic speed scale shown. The positions of the CTD and STD stations are given by small dots. The maps were contoured by computer in intervals of 4 dyn cm using an objective analysis program which smoothed the dynamic heights over a space scale of 60 km and a time scale of 30 days. The ability of the contours to resolve the spatial structure of the field depends on the density of the station data. Near the map centers about 80% of the total dynamic height variance has been resolved. On the map periphery the resolution is much poorer, only 20 to 30% of the variance being resolved.

The sequence in Figs. 3 and 4 shows a slowly moving, close-packed array of cyclonic and anticyclonic surface pressure disturbances having a spatial periodicity of about 400 km (eddy diameter of 200 km) and an amplitude of about 0.1 dyn m. The field is clearly irregular and unsteady. Features on scales of several hundred kilometers tend to persist throughout the mapping sequence (4 months) while on the smallest scale (60 km) they cannot be traced from one map to another (15 days). In March the center of the MODE area appears to be in a saddle between two high pressure cells (anticyclonic eddies) to the east and west and two low pressure cells (cyclonic eddies) to the north and south. The eastern anticyclonic eddy moves into the center during the first half of April and then enlarges and dominates the MODE area until the end of June. During this period it slowly drifts westward, moving out of the area in the first half of July at the end of MODE.
Fig. 3, Sea surface temperature maps (upper) and surface dynamic topography (lower), relative to 1500 db, of the MODE area for four successive periods in the first half of the experiment. The cross-hatched area on the maps of dynamic height show all surface water cooler than the mean for that period.
Fig. 4. As in Fig. 3 except for the second half of the experiment.
between four apparent eddies advects southward a 200–300 km tongue of cool northern water along longitude 70°30' W, and advects northward a similar tongue of warm southern water along 69°00' W. Similar patterns occur in the other maps. At times, isolated pools of warm water (30 April–14 May) or cool water (14 June–28 June) are formed as a result of the eddy currents.

Important changes can occur within the mapping periods of Figs. 3 and 4. It is possible to examine some of this in more detail on a one-week time scale by examining selected STD data. These fields are shown for the weeks of 14 May and 21 May in Fig. 6. The temperature pattern at a depth of 50 m looks very similar to the surface pattern for the same time in Fig. 4. Two tongues of water are evident, a warm tongue extending to the north, and a developing cold tongue extending to the south. Even on a weekly time scale considerable changes occur (in both the depicted fields). It is interesting to note that the temperature field is asymmetric relative to the dynamic topography. The warm tongue sits over the eddy defined by closed contours of dynamic height, suggesting a recirculation of the warm waters within the tongue. The cold tongue, however, is situated over dynamic height contours that do not close.

In general, surface isotherms do not coincide on the large scale with contours of dynamic height. The pattern of the first is intrusive or finger-like while the second is circular or eddy-like. This is in direct contrast with the deep horizontal temperature structure in the main thermocline where the patterns were remarkably similar, with cyclonic eddies having cool centers and anticyclonic eddies warm centers.

The anomalous surface temperature structure in our maps extends at least over a decade of decreasing scales, from 400 km down to the order of 40 km. The former corresponds to the average wavelength between eddies and the latter is typical of the width of the long intrusions of warm and cool water. Features on a smaller scale, although occasionally resolved, were likely to be advected and distorted beyond recognition in the mapping intervals.) Over this range it is reasonable to suppose that temperature variance is extracted from the mean meridional temperature gradient at the large scale and cascades to the small scales. The Lagrangian time for this cascade would depend on how quickly the surface currents can distort the temperature field. From the spatial structure on the maps of dynamic height one estimates that large temperature features are stretched and thinned by a factor of 2 in a period of 5 days to a week. Hence, one estimates that the long intrusions are formed in 15 to 30 days, that is, in the time required for surface water to move one-half to once around the periphery of a typical eddy.

The above cascade extends to much smaller scales than those resolved in our maps. The most obvious

---

4. Surface temperature advection

By comparing maps in Figs. 3 and 4 it becomes very clear that large-scale features in the surface temperature field are due primarily to eddy surface currents which advectively distort the mean meridional gradient. To facilitate this comparison we have cross-hatched on the maps of dynamic topography all those areas of the sea surface which are cooler than the mean temperature of the corresponding temperature map. Note, in particular, the situation during the period 9–30 March in Fig. 3. Here, the interplay between four apparent eddies advects southward a 200–300 km tongue of cool northern water along longitude 70°30' W, and advects northward a similar tongue of warm southern water along 69°00' W. Similar patterns occur in the other maps. At times, isolated pools of warm water (30 April–14 May) or cool water (14 June–28 June) are formed as a result of the eddy currents.

Important changes can occur within the mapping periods of Figs. 3 and 4. It is possible to examine some of this in more detail on a one-week time scale by examining selected STD data. These fields are shown for the weeks of 14 May and 21 May in Fig. 6. The temperature pattern at a depth of 50 m looks very similar to the surface pattern for the same time in Fig. 4. Two tongues of water are evident, a warm tongue extending to the north, and a developing cold tongue extending to the south. Even on a weekly time scale considerable changes occur (in both the depicted fields). It is interesting to note that the temperature field is asymmetric relative to the dynamic topography. The warm tongue sits over the eddy defined by closed contours of dynamic height, suggesting a recirculation of the warm waters within the tongue. The cold tongue, however, is situated over dynamic height contours that do not close.

In general, surface isotherms do not coincide on the large scale with contours of dynamic height. The pattern of the first is intrusive or finger-like while the second is circular or eddy-like. This is in direct contrast with the deep horizontal temperature structure in the main thermocline where the patterns were remarkably similar, with cyclonic eddies having cool centers and anticyclonic eddies warm centers.

The anomalous surface temperature structure in our maps extends at least over a decade of decreasing scales, from 400 km down to the order of 40 km. The former corresponds to the average wavelength between eddies and the latter is typical of the width of the long intrusions of warm and cool water. Features on a smaller scale, although occasionally resolved, were likely to be advected and distorted beyond recognition in the mapping intervals.) Over this range it is reasonable to suppose that temperature variance is extracted from the mean meridional temperature gradient at the large scale and cascades to the small scales. The Lagrangian time for this cascade would depend on how quickly the surface currents can distort the temperature field. From the spatial structure on the maps of dynamic height one estimates that large temperature features are stretched and thinned by a factor of 2 in a period of 5 days to a week. Hence, one estimates that the long intrusions are formed in 15 to 30 days, that is, in the time required for surface water to move one-half to once around the periphery of a typical eddy.

The above cascade extends to much smaller scales than those resolved in our maps. The most obvious

---

2 Ideally one would like to quantify this statement by numerical correlation between temperature and eddy fields. The density of our data set is not high enough, unfortunately, to make such a correlation statistically significant.
of such features observed during MODE were those associated with surface frontogenesis. Voorhis and Hersey (1964), Voorhis (1969) and Katz (1969) had shown prior to MODE that these surface fronts are very narrow transition zones separating adjacent surface water masses of different temperatures, salinities and densities which can meander along the sea surface for distances of several hundred kilometers. Surface temperature changes of 1–2°C are frequently observed across a front in a distance of only 100 to 200 m. Beneath the surface a frontal pycnocline slopes downward beneath the lighter water and becomes level in a horizontal distance of the order of 10 km and at depths usually of the order of 100 to 200 m, although occasionally it is much deeper. Associated with the sloping pycnocline is a geostrophic current jet flowing along the front with surface speeds as large as 50 to 100 cm s⁻¹.

The continuous shipboard records of surface temperature from MODE showed numerous frontal crossings along the boundaries of the long intrusive warm and cool tongues in Figs. 3 and 4, and fronts were frequently seen in these areas by the high concentrations of surface debris. The nature of the program and the haphazard sampling, however, made it impossible to map particular frontal features. It is significant, however, that a synoptic image of the surface temperature field from a satellite (Fig. 2) usually shows the boundaries between the intrusive tongues to be much sharper (greater thermal gradients) than in our 7–15 day maps. We suggest that frontogenesis is common along these boundaries and that frontal currents may contribute an important near surface circulation around the boundaries of the long intrusive features which is superimposed on broader scale eddy surface currents along their axis.

5. Surface wind drift

We have so far neglected the advection and distortion of surface temperature structure by surface currents other than those due to mesoscale eddies or possible near surface geostrophic currents associated with frontogenesis. The most important of the former are the shallow surface currents driven by wind stress.

During MODE all ships routinely reported wind speed and direction once daily. The weather from March through mid-May was dominated by a succession of moderate high and low pressure disturbance every 5 to 10 days with mainly veering winds which varied in speed from less than 1 m s⁻¹ to no more than 15 m s⁻¹. Conditions were somewhat steadier from mid-May onward with disturbance every 10 to 15 days. The wind backed and veered with maximum speeds less than 10 m s⁻¹.

Surface drift current was computed using the model of Gonella (1971), which assumes an Ekman current
Table 1. Mean surface stress magnitude ($\tau$) and direction ($\phi$) computed from observed wind speed and direction, assuming a drag coefficient of $1.2 \times 10^{-3}$; mean mixed layer depth ($h$) from CTD observations; and mean surface drift speed ($V$) and direction ($\theta$) computed from Gonella (1971), assuming an eddy viscosity of $10^3$ cm$^2$ s$^{-1}$.

<table>
<thead>
<tr>
<th>Period</th>
<th>$\tau$ (dyn cm$^{-2}$)</th>
<th>$\phi$ (°T)</th>
<th>$h$ (m)</th>
<th>$V$ (m day$^{-1}$)</th>
<th>$\theta$ (°T)</th>
</tr>
</thead>
<tbody>
<tr>
<td>9 Mar–30 Mar</td>
<td>0.16</td>
<td>68</td>
<td>38</td>
<td>1.7</td>
<td>115</td>
</tr>
<tr>
<td>31 Mar–14 Apr</td>
<td>0.30</td>
<td>37</td>
<td>34</td>
<td>3.1</td>
<td>81</td>
</tr>
<tr>
<td>15 Apr–29 Apr</td>
<td>0.82</td>
<td>260</td>
<td>40</td>
<td>8.5</td>
<td>305</td>
</tr>
<tr>
<td>30 Apr–14 May</td>
<td>0.40</td>
<td>260</td>
<td>28</td>
<td>4.1</td>
<td>305</td>
</tr>
<tr>
<td>15 May–29 May</td>
<td>0.26</td>
<td>333</td>
<td>11</td>
<td>3.2</td>
<td>68</td>
</tr>
<tr>
<td>30 May–13 June</td>
<td>0.20</td>
<td>321</td>
<td>13</td>
<td>6.1</td>
<td>335</td>
</tr>
<tr>
<td>14 Jun–28 Jun</td>
<td>0.21</td>
<td>329</td>
<td>12</td>
<td>3.1</td>
<td>36</td>
</tr>
<tr>
<td>29 Jun–13 Jul</td>
<td>0.20</td>
<td>323</td>
<td>11</td>
<td>3.0</td>
<td>33</td>
</tr>
</tbody>
</table>

Mode mean        | 0.20                   | 293         | 25     | 2.2                | 356           |

* 1 dyn cm$^{-2}$ = 0.1 N m$^{-2}$.

6. Conclusions and discussion

In the MODE area we conclude that the surface temperature field, on time scales less than about one month and over space scales from 400 to about 40 km, tags primarily the surface currents associated with the baroclinic mesoscale eddy field of the main thermocline. Surface currents induced by wind stress appear to be of secondary importance in generating spatial structure in this scale range. Relatively little can be said about scales less than 40 km. However, there is some evidence from MODE but mostly from previous measurements in the same area that much of the spatial structure on scales less than about 10 km$^4$ tags not only the mesoscale current field but also a relatively shallow field of currents which are in geostrophic equilibrium with horizontal density gradients in the near surface layers. This new field of currents is often jet-like and is associated with surface frontogenesis.

Mesoscale eddies appear to be an effective mechanism for stirring the large-scale thermal (and haline) field imposed on the near surface layers by the atmosphere. One can speculate that this surface process on an eddy time scale may generate a net meridional heat transport in the surface layer on a longer time scale. For example, if a single anticyclonic eddy develops in the convergence zone one would expect warm water to move initially northward on its western side and cool water southward on its eastern side. (The flows will change sides if the eddy is cyclonic.) In time both flows will simply circulate in a complex manner around the eddy with a great deal of stirring but no net heat transport if there is no heat exchange between the warm and cool water. However, if there

---

It is significant that this scale is of the order of the internal radius of deformation of the near-surface pycnocline.
are many eddies, which are evolving, moving, and decaying; it is highly likely that surface water is exchanged\(^4\) from eddy to eddy and one might expect to observe at times long tongues of warm and cool surface water running north and south. This is very similar to what one sees in Fig. 2. The result would be a mean meridional heat transport northward in the MODE area of the order of \(VH\) per eddy, where \(V\) is the geostrophic advecting surface velocity, and \(H\) is the anomalous heat carried by each tongue. The latter can be approximated by \(\rho_c C_p DL \Delta T\), where \(\Delta T\) is the temperature difference between north and south flowing tongues, \(L\) is the zonal width of the tongue, and \(D\) is the depth of the heat anomaly. Representative values for these parameters are \(V = 20\ \text{cm s}^{-1}\), \(\rho_c = 1\ \text{g cm}^{-3}\), \(C_p = 4.18\ \text{J g}^{-1}\ \text{K}^{-1}\), \(D = 50\ \text{m}\), \(L = 100\ \text{km}\), \(\Delta T = 2^\circ\text{C}\). Using these values one computes a transport of \(8.2 \times 10^{12}\ \text{W per eddy}\). Taking 200 km as a mean zonal spacing between eddies one finds a northward eddy heat transport of \(4.2 \times 10^{19}\ \text{W across each kilometer in an east-west direction. Assuming the northern Sargasso Sea to be bounded on the north and west by the Gulf Stream, on the east by 50\(^\circ\text{W}\) longitude, and on the south by 30\(^\circ\text{N}\) latitude, one computes an annual heat input of \(32 \times 10^{10}\ \text{J across its southern boundary (length 2400 km)}\) by the eddy mechanism. This is of the same order as the annual heat loss to the atmosphere across its surface area (\(2.2 \times 10^6\ \text{km}^2\)) computed from Bunker and Worthington (1976), using an average net heat surface flux of 66 W m\(^{-2}\) (50 kcal cm\(^{-2}\) year\(^{-1}\)). Speculating on a still larger scale and assuming that the observed mesoscale eddy activity extends across both the northern Atlantic and Pacific Oceans at mid-latitudes, a total distance of the order of \(1.6 \times 10^4\ \text{km}\), one finds an annual poleward heat transport by the eddies at these latitudes of the order of \(6.8 \times 10^{14}\ \text{W}\). This can be compared with the annual oceanic poleward energy transport of about \(22.6 \times 10^{14}\ \text{W (1.7} \times 10^{22}\ \text{cal year}^{-1}\)) estimated by Vonder Haar and Oort (1973). Considering the uncertainties in all of these estimates one concludes that the mesoscale eddy heat transport may not be inconsequential.

Finally, our results can have important implications for oceanographers and meteorologists interested in annual or longer term changes in sea surface temperature and their effect on world climate. The fluctuating mesoscale temperature field is unwanted noise from their point of view and introduces an uncertainty to estimates of mean temperatures. For data collected from a fixed point (or within an eddy radius of this point) this uncertainty is of the order of \(\Delta T_n/\sqrt{n}\), where \(\Delta T_n\) is the rms temperature change due to a typical eddy, and \(n\) is the number of eddy events in the averaging time. Assuming \(\langle \Delta T_n \rangle \approx 0.5^\circ\text{C}\) and no other sources of noise, one would have to average over 25 eddy events in the MODE area in order to resolve a climatic 0.1\(^\circ\text{C}\) change in mean surface temperature. If the eddy residence time is of the order of 2 months this would take 4 to 5 years.

Acknowledgments. This work was supported by the Office of Naval Research under Contract N00014-74-C-0262, NR 083-004 and by the Office of the International Decade for Ocean Exploration of the National Science Foundation under Funding Agreement AG-385.

The data used in this paper were collected and processed by many people in the MODE program and the authors wish to acknowledge all of this work and to express their gratitude. We would also like to thank N. Fofonoff of the Woods Hole Oceanographic Institution, Woods Hole, Mass., who programmed and computed the objective maps of dynamic height in Figs. 3 and 4.

REFERENCES


\(^4\) This may be greatly enhanced by the unusually strong surface currents associated with surface frontogenesis.