
MOTOTAKA NAKAMURA AND YI CHAO
Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California

(Manuscript received 7 January 1999, in final form 21 August 2001)

ABSTRACT

Output of an eddy-resolving model of the North Atlantic is diagnosed in the vicinity of the Gulf Stream (GS), using quasigeostrophic potential vorticity (QGPV), quasigeostrophic potential enstrophy (QGPE), and modified divergent eddy potential vorticity flux, \( (\nabla \times \mathbf{q})_{\lambda} \). A tongue or an elongated island of large mean QGPV along the model GS in the top 1000 m is associated with predominantly downgradient \( (\nabla \times \mathbf{q})_{\lambda} \), suggesting that the horizontal eddy fluxes are balancing a sink of eddy QGPE in most of the tongue or island by converting the mean QGPE into eddy QGPE. Some large upgradient \( (\nabla \times \mathbf{q})_{\lambda} \) is observed to the north of the center of the tongue or island, however, suggesting that some of the eddy fluxes in the northern half of the tongue or island of high QGPV are balancing a source of eddy QGPE there by converting eddy QGPE into the mean QGPE. At the intermediate levels of the model, under the GS, eddy QGPE is small, and the role of eddies appears to be mixed; they are forcing the mean to some extent and dissipating the mean to some extent. At the deep levels, the eddies show predominantly a dissipative role, tending to convert the mean QGPE into eddy QGPE. In the region of large time-mean meanders in the model GS, eddies are found to be reinforcing the meanders in a way very similar to that found in diagnoses of atmospheric blocking, which is essentially a large quasi-stationary meander in the subtropical jet. It suggests that the problem of the excessive meander amplitude in the model may be due to an imbalance between eddy forcing in the vicinity of the separation point and zonal acceleration of the GS simulated by the model or due to an unrealistically strong topographic stationary wave forcing in the model.

1. Introduction

In a companion paper, Nakamura and Chao (2001, hereafter Part I), quasigeostrophic potential vorticity \( (\mathbf{q}) \) and its time mean \( \langle \mathbf{q} \rangle \), calculated from output of an eddy-resolving model of the North Atlantic, are shown and discussed for the vicinity of the Gulf Stream (GS). The horizontal structures of \( \langle \mathbf{q} \rangle \) shown in Part I have qualitative similarity to those of the climatological large-scale potential vorticity (PV) field calculated by Kefler (1985) and Lozier (1997), suggesting that the model output may be useful in studying the circulations in this region as a pseudo-dataset. A conspicuous feature in \( \langle \mathbf{q} \rangle \) in the top 1000 m of the model ocean is a tongue or an island of large \( \langle \mathbf{q} \rangle \), extending from the vicinity of the GS separation point east-northeastward along the path of the GS. Trajectory calculations, as well as visual inspections of movies of \( \mathbf{q} \) evolution, strongly suggest that the tongue of large \( \langle \mathbf{q} \rangle \) is a product of strong \( \mathbf{q} \) input by the GS, vigorous mixing by eddies along the path of the GS, and dissipation of \( \mathbf{q} \). Between 1000 and 2500 m of the model ocean, a large and deep pool of horizontally weakly varying \( \langle \mathbf{q} \rangle \), which corresponds to a pool of nearly homogenized mean PV on a surface of \( \langle \mathbf{q} \rangle_{3125m} = 37.5 \) in the model, is found to the west of the Mid-Atlantic Ridge (MAR), while large \( \langle \mathbf{q} \rangle \) is seen to be created by the Mediterranean overflow to the east of the MAR. In Part I we speculated that the pool of weakly varying \( \langle \mathbf{q} \rangle \) under the GS may be a consequence of the tendency of baroclinic oceans toward baroclinic neutrality driven by eddies, as suggested for the midlatitude atmosphere by Lindzen (1993). Below 2500 m, \( \langle \mathbf{q} \rangle \) shows substantial meridional gradient to the west of the MAR, while it becomes nearly homogenized to the east of the MAR.

Although examination of \( \mathbf{q} \) and \( \langle \mathbf{q} \rangle \) does provide us with some insight into the mechanisms through which \( \langle \mathbf{q} \rangle \) and the mean circulation are generated and maintained, it does not yield any clear assessment of the role of eddies in the maintenance of \( \langle \mathbf{q} \rangle \) and the mean flow. Given the observed and simulated vigorous eddy activity in the vicinity of the GS and their critical role in the formation and maintenance of the tongue of large \( \langle \mathbf{q} \rangle \), as suggested in Part I, examination of the model output in a framework of eddy–time mean relationship is essential to obtaining the complete picture of the important pro-
cesses underlying the mean state. Indeed, such an ex-
amination of the role of eddies is just what an eddy-
resolving model output may be most useful for, since
there is a climatological dataset but very limited data
to calculate various eddy statistics on spatially and tem-
porally extensive scales. While diagnostic studies on
the relationship between the mean or low-frequency state
eddies in the atmosphere have been common, due
to the relative abundance of data and model output to
perform such calculations with, similar calculations for
oceanic circulations have not been done commonly be-
cause of the limited availability of the necessary data.
Reflecting the scarcity of such data, studies of eddy-
time mean relationships in large-scale ocean circulations
are rare, although some attempts have been made with
limited data on regional scales (e.g., Chester et al. 1994;
Cronin and Watts 1996; Cronin 1996). Although the
model simulation is far from perfect, some of the key
features of the GS and its surroundings have been re-
produced reasonably well (Chao et al. 1996; Part I).
Therefore, diagnoses of the model output in a dynamical
framework may be useful in studying important pro-
cesses and their roles in the real oceanic flows. Such
diagnoses may also provide us with clues to the causes
of unrealistic features in the model simulation. We may
also learn information useful for making our future ob-
servational efforts more effective.

Here we attempt to study circulations in the vicinity
of the GS in a framework based on $q$ and potential
enstrophy, using a 5-yr output of an eddy-resolving gen-
eral circulation model (GCM) of the North Atlantic. We
focus our attention on the relationship between the mean
flow and eddies in this paper, diagnosing the output with
equations derived from the equation of potential en-
strophy, while Part I describes and discusses time-de-
pendent $q$ fields and the time-mean state. The main
purpose of this work is to examine the structure of eddy
potential enstrophy and its sources and sinks on hori-
zontal surfaces and how the horizontal eddy fluxes con-
tribute to these structures in the vicinity of the GS, using
the model output as pseudodata. This may be considered
as an extension of earlier studies that used idealized
ocean models (e.g., Holland and Rhines 1980) to a more
realistic model ocean. Another objective of the study is
to examine the model output in a dynamical framework
to deduce potential mechanisms that may be responsible
for interesting model-generated features that are also
observed in the real ocean. Section 2 describes the di-
agnostic framework used for the current paper; section
3 describes and discusses the results of the calculations;
and finally, section 4 summarizes the results.

2. Calculation of eddy potential enstrophy and
eddy potential vorticity flux

The model output used for this work was generated
by an eddy-resolving primitive-equation ocean GCM of
the North Atlantic, based on the Parallel Ocean Program
(POP) developed at Los Alamos National Laboratory
(Dukowicz and Smith 1994). The model has a free sur-
face and a realistic topography. The horizontal grid
spacing of the model is approximately $\frac{1}{6}^\circ$ (0.1843° lati-
tude and 0.1875° longitude). There are 37 unevenly spaced vertical levels in the model. The portion of in-
tegration from which we extract the output is driven
with the surface salinity restored toward the Levitus
climatology (Levitus et al. 1994), the surface heat flux
derived from the European Centre for Medium-Range
Weather Forecasts (ECMWF) analysis, and the surface
wind stress derived from the ECMWF analysis. The
output provides potential temperature, salinity, and the
horizontal flow velocity every three days from year 25
to year 29 of the model integration at all grid points in
the model. We focus our attention on a domain that
covers 30°–60°N, 75°–15°W for the present work. The
choice is made to cover most of the GS after separation
from the western boundary and the North Atlantic Cur-
rent (NAC) and to make calculations of the diagnostics
reasonably efficient on our computational resources.
The reader is referred to Part I for further details of the
model output.

While Ertel’s PV on isopycnal surfaces would be de-
sirable for diagnosing large-scale oceanic circulations,
there are some problems associated with calculation of
the quantity from descretized model output, as described
in Part I. In particular, vertical interpolation of PV and
velocities onto isopycnal surfaces that are not naturally
defined in the model can introduce large spurious PV
fluxes and their divergence. Large-scale oceanic flows
are quasigeostrophic to first order except at a sharp front,
such as the GS. However, the front at the GS simulated
by the model is not as sharp as that at the real GS and,
thus, the model flow is even closer to quasigeostrophy
than the real flows. Indeed, with the horizontal grid
spacing of the model, which is approximately 20 km,
the finest barely resolved structure has a horizontal scale
of roughly 60 km. With this horizontal scale and flow
speeds of roughly 2 m s$^{-1}$ at the maximum, the Rossby
number of the flows simulated by the model is much less
than 1 everywhere all the time. We have, therefore, chosen
quasigeostrophic PV (QGPV), $q$, as the central quantity
of the diagnoses. The form of $q$ we use is given by

$$q = f + \zeta + \int \frac{\partial \sigma^*}{\partial \zeta} \frac{d \sigma^*}{d \zeta} = f + \zeta - \frac{g}{\sigma_o N^2} \frac{\partial \sigma^*}{\partial \zeta}, \quad (1)$$

where $N^2$ is the Brunt–Väisälä frequency, $\sigma_o$ is the ref-
ence potential density at a constant depth, and $\sigma^*$ is
the deviation in $\sigma$ from $\sigma_o$. Here, $f$ is the planetary
vorticity, $g$ the gravitational constant, and $\zeta$ the relative
vorticity. For details of $q$ formulation and computation,
the reader is referred to Part I.

To study the relationship between the time mean and
eddies, an approach based on eddy potential enstrophy
is used. It has been used by many investigators in the
past, providing useful information on the role of eddies
in the time-mean circulation (e.g., Rhines and Holland 1979; Holland and Rhines 1980; Marshall and Shutts 1981). Here, we follow the framework used by Nakamura (1998).

Let us begin with an equation for \( \frac{\partial}{\partial t} \left( \hat{q}^2 \right) + \mathbf{V} \cdot \mathbf{\nabla} \left( \frac{\hat{q}^2}{2} \right) \) averaged over time,
\[
\frac{\partial}{\partial t} \left( \frac{\hat{q}^2}{2} \right) + \mathbf{V} \cdot \mathbf{\nabla} \left( \frac{\hat{q}^2}{2} \right) = \mathbf{s},
\]
where \( \mathbf{V} \) is the flow velocity and \( \mathbf{s} \) is the source/sink of \( q \). An overbar denotes a time mean. When this equation is separated into two parts, one for the time mean and the other for perturbations from the mean, it yields the following equations:
\[
\begin{align*}
\frac{\partial}{\partial t} \left( \frac{\langle q^2 \rangle}{2} \right) + \mathbf{V} \cdot \mathbf{\nabla} \left( \frac{\langle q^2 \rangle}{2} \right) &= \mathbf{\pi} \langle q \rangle \\
\frac{\partial}{\partial t} \left( \frac{\langle q^2 \rangle}{2} \right) + \mathbf{V} \cdot \mathbf{\nabla} \left( \frac{\langle q^2 \rangle}{2} \right) &= \mathbf{s}^\prime \langle q \rangle,
\end{align*}
\]
where a prime denotes a deviation from the time mean. The fourth term on the left-hand side of Eqs. (3) and (4) is conversion terms between \( \langle \hat{q}^2 \rangle \) and \( \langle q^2 \rangle \). This framework, consisting of Eqs. (3) and (4), can be useful for diagnosing any scalar that is conserved following a low-frequency-extracting operation, changing the mean of a simple form, \( \mathbf{D} q/\mathbf{D} t = \mathbf{s} \). These equations may be applied to frequency-filtered data as well. For example, one may replace the time-averaging operation with a low-frequency-extracting operation, changing the meaning of primed quantities from deviations from the mean to the high-frequency component, in which case \( \partial/\partial t )|_{|t^2/2|} \) and \( \partial/\partial t |_{t^2/2} \) in Eqs. (3) and (4) are nonzero.

We now focus our attention on Eq. (4). Marshall and Shutts (1981) pointed out that when \( \mathbf{V} \cdot \mathbf{\nabla} = 0 \) and \( \mathbf{\nabla} \cdot \mathbf{\nabla} = 0 \), a part of \( \langle q^2 \rangle \) simply balances the advection of \( \hat{q}^2 \) by \( \mathbf{V} \) and that part of the eddy flux is rotational. (Here a subscript \text{h} denotes a horizontal vector.) Upon this recognition, they defined a rotational eddy flux of \( \mathbf{\nabla} q^ \prime \langle q^2 \rangle \rangle^\prime \), that balances the mean flow advection of \( \langle q^2 \rangle \) by
\[
\langle \mathbf{V} \cdot q^ \prime \rangle^\prime - \mathbf{\nabla} \left( \frac{\langle q^2 \rangle}{2} \right).
\]
They further defined the “divergent” eddy flux (which is not necessarily purely divergent) or residual eddy flux
\[
\langle \mathbf{V} \cdot q^ \prime \rangle^\prime - \langle \mathbf{V} \cdot q^ \prime \rangle^\prime \text{MS}.
\]
The superscript MS is used to distinguish the Marshall–Shutts residual flux from Nakamura’s \( \langle \mathbf{V} \cdot q^ \prime \rangle^\prime \), \langle V \cdot q^2 \rangle^\prime. With this definition and a further assumption that \( \mathbf{V} \cdot q^ \prime = 0 \), they rewrote Eq. (4) as
\[
\frac{\partial}{\partial t} \left( \frac{\langle q^2 \rangle}{2} \right) + \langle \mathbf{V} \cdot q^ \prime \rangle^\prime - \mathbf{\nabla} \left( \frac{\langle q^2 \rangle}{2} \right) = \mathbf{s}^\prime \left( \langle q \rangle \right).
\]
When this is the case, the “divergent” or residual component of \( \langle \mathbf{V} \cdot q^ \prime \rangle \rangle^\prime \), the portion of \( \langle \mathbf{V} \cdot q^ \prime \rangle \rangle^\prime \) that balances the net source/sink of \( q^2 \) on a horizontal plane and contributes to the net transport of \( q \) and a rotational component of \( \langle \mathbf{V} \cdot q^2 \rangle \rangle^\prime \), \langle \mathbf{V} \cdot q^2 \rangle \rangle^\prime \) simply balances the mean flow advection of \( q^2 \) in a steady state, that is,
\[
\langle \mathbf{V} \cdot q^ \prime \rangle^\prime + \langle \mathbf{V} \cdot q^2 \rangle^\prime = \mathbf{s}^\prime \left( \langle q \rangle \right). \tag{6}
\]
However, these three conditions are not met readily in general when eddy contribution to \( \mathbf{s} \) forcing is not negligible. These three conditions are likely to be met when the role of eddies in \( q^2 \) advection and mixing is negligible. For the mean flow to be parallel to the contour of \( q \), that is, \( \mathbf{V} \cdot q^2 = 0 \), it must be true that \( \mathbf{V} \cdot q^ \prime = 0 \) in an adiabatic-inviscid flow. This is not likely to be met in general in realistic ocean flows with geostrophic turbulence. Also, Nakamura (1998) found that \( \mathbf{V} \cdot q^ \prime \cdot q^ \prime \) has the same order of magnitude as \( \langle \mathbf{V} \cdot q^ \prime \rangle \rangle^\prime \) (\( \langle \mathbf{V} \cdot q^2 \rangle \rangle^\prime \) will be defined below) in the upper troposphere where the relative strength of the mean flow advection with respect to eddies is likely to be substantially greater than it is in the ocean. Thus, we believe that the three conditions are not met in real oceans and in flows generated by eddy-resolving models.

More generally, when the three conditions are not met, Nakamura (1998) showed that
\[
\frac{\partial}{\partial t} \left( \frac{\langle q^2 \rangle}{2} \right) + \langle \mathbf{V} \cdot q^ \prime \rangle^\prime \cdot \mathbf{\nabla} = \mathbf{S}_E, \quad \tag{8}
\]
\[
\langle \mathbf{V} \cdot q^ \prime \rangle^\prime \cdot \mathbf{\nabla} = - \mathbf{\nabla} \cdot \mathbf{\nabla} \left( \frac{\langle q^2 \rangle}{2} \right), \quad \tag{9}
\]
where the modified “divergent” eddy flux or the residual flux, \( \langle \mathbf{V} \cdot q^2 \rangle \rangle^\prime \), and the modified “rotational” flux of the mean balancing flux, \( \langle \mathbf{V} \cdot q^2 \rangle \rangle^\prime \), are defined by
\[
\langle \mathbf{V} \cdot q^2 \rangle \rangle^\prime = \langle \mathbf{V} \cdot q^2 \rangle^\prime - \mathbf{F}_h = \langle \mathbf{V} \cdot q^2 \rangle^\prime, \quad \tag{10}
\]
\[
\langle \mathbf{V} \cdot q^2 \rangle \rangle^\prime = \gamma k \times \mathbf{\nabla} \left( \frac{\langle q^2 \rangle}{2} \right) \left( \frac{\mathbf{\nabla} \cdot \mathbf{\nabla} \left( \langle q^2 \rangle / 2 \right)}{\mathbf{\nabla} \cdot \mathbf{\nabla} \left( \langle q^2 \rangle / 2 \right)} \right). \quad \tag{11}
\]
Here, \( \mathbf{F}_h \) is an effective eddy flux that represents the effect of horizontal eddy self advection, defined by
\[
\mathbf{F}_h = \langle \mathbf{V} \cdot q^2 \rangle^\prime \mathbf{v}^\prime \mathbf{q}^\prime \cdot \mathbf{\nabla} = \mathbf{V} \cdot q^2 \cdot \mathbf{V} \cdot q^2 = \frac{\mathbf{V} \cdot q^2 \cdot \mathbf{V} \cdot q^2}{2|\mathbf{\nabla} \cdot \mathbf{\nabla} \left( \langle q^2 \rangle / 2 \right)|} \quad \tag{12}
\]
and \( S_E, D, \) and \( \gamma \) are defined by...
where \( q \) is a small number, which vanishes in the limit of time average. Then, from Eqs. (8) and (16), we have

\[
\frac{\partial}{\partial t} \left( \frac{(q')^2}{2} \right) = \epsilon + (V q') \cdot \nabla \bar{q} - S_e. \tag{17}
\]

Therefore, \((V q') \cdot \nabla \bar{q}\) represents the flux of \( q' \) that tends to convert \( \bar{q} \) into \( \frac{(q')^2}{2} \) where it is down the gradient of \( \bar{q} \), and vice versa. It is important to remember that interpretation of \((V q') \cdot \nabla \bar{q}\) is meaningful, in general, only in combination with \( \nabla \bar{q} \). As mentioned earlier, the equations derived by Nakamura (1998) reduce to those derived by Marshall and Shutts (1981) when \( \nabla \bar{q} = 0 \), \( \nabla \times \nabla \bar{q} = 0 \), and \( \nabla q' \cdot \nabla q' = 0 \). Since these conditions are not satisfied readily in reality, Nakamura’s formulas are more appropriate for general use in analyzing eddy enstrophy. An advantage of examining \((V q') \cdot \nabla \bar{q}\) along with \( \bar{q} \) and \( \frac{(q')^2}{2} \) over examining \((V q') \cdot \nabla q\) is that \((V q') \cdot \nabla \bar{q}\) represents all of the horizontal flux of \( q' \) that has a net contribution to exchanges between \( \bar{q} \) and \( \frac{(q')^2}{2} \) and, thus, provides a concise picture of the relationship between \( \bar{q} \) and \( \frac{(q')^2}{2} \) through horizontal fluxes of \( q' \). When the overbar denotes a time averaging operation, thus making Eq. (8) \((V q') \cdot \nabla \bar{q} = S_e\), an examination of \((V q') \cdot \nabla \bar{q}\) along with \( \bar{q} \) provides information on the net source/sink of eddy enstrophy on a horizontal plane due to diabatic effects and vertical fluxes, and on how horizontal fluxes of \( q' \) are compensating for the source/sink. One does not have to calculate individual terms on the right-hand side of Eq. (13) in such a case, unless one wishes to isolate contributions from individual terms in Eq. (13).

In the analyses to be presented here, we use time averaging as the operation denoted by an overbar and, therefore, use a prime for a deviation from the time mean. (As noted earlier, one may use other time-filtering operations with this diagnostic framework. In particular, separating time series into low-frequency and high-frequency components may prove more useful in certain cases.) We compute \( \bar{q} \) and \( \nabla \bar{q} \) by averaging \( q \) and \( \nabla q \) over the entire five years and then taking an equal-weight 9-point spatial average to smooth the fields. Then, \( \frac{(q')^2}{2} \), \( (V q') \cdot \nabla q' \), \( \nabla \cdot (V q') \cdot \nabla q' \), \( \nabla \cdot (V q') \cdot \nabla q' \cdot \nabla \bar{q} \), and \( \nabla \cdot (V q') \cdot \nabla q' \cdot \nabla \bar{q} \cdot \nabla \bar{q} \) are calculated by time averaging \( q'^2 \) and \( \nabla q' \cdot \nabla q' \) and using Eqs. (10)–(15). We do not evaluate individual terms in Eq. (13) since our primary objective here is to determine the net source/sink of \( \frac{(q')^2}{2} \) on horizontal planes and how horizontal eddy fluxes contribute to the balance over the 5-yr period. Also, the extreme noisiness and very small magnitudes of \( w' \) in the model also make meaningful evaluation of the individual terms on the right-hand side of Eq. (13) very difficult.

3. Results and discussion

a. Mean potential vorticity

The time-mean \( q \) fields are described and discussed in detail in Part I. Only the major features of interest are summarized here. Figure 1 shows \( \bar{q} \) and \( \nabla \bar{q} \) at the levels shown for \( \bar{q} \) in Part I. For a reference, \( \bar{q} \)
structures of /2 reflect those of Mediterranean overflow. Below 2500 m, shows a structure baroclinic instability, which may well account for the meridional gradient in this region at deep levels is collocated with a region of significant negative meridional gradient in at shallower levels and are seen to the north of the approximate center of the tongue shifts southward with depth until it disappears at around 1000 m. Similar structures are seen in the 5-yr mean PV calculated from the model potential temperature and salinity and also in the climatological PV fields calculated by Keffer (1985) from the Levitus (1982) dataset. In Part I, it was demonstrated that the tongue of high /2 is a product of strong q input by the GS, mixing by eddies, and dissipation of q along the path of water injected by the GS. Between 1000 and 2500 m, /2 shows large values with strong gradients to the east of the MAR due to the Mediterranean overflow (Figs. 1e,f). To the west of the MAR, /2 is weakly varying, and the mean Ertel’s PV in the model shows a pool of nearly homogenized PV that appears very similar to that found in a hydrographic dataset by Lozier (1997, 1999). In Part I we reported strong eddy mixing within the pool of weakly varying /2 and a mixing barrier at the edge of high- /2 water mass influenced by the Mediterranean overflow. Below 2500 m, /2 shows a structure with little zonal variation and substantial meridional gradient to the west of the MAR, while it is weakly varying to the east of the MAR (Fig. 1g). In general, /2 is not parallel to the contours of /2 and, thus, /2 is not negligible. Where /2 is very large, for example, along the GS in the upper layers, contours of /2 are nearly parallel to /2. However, even in such a circumstance, /2 is fairly large due to the largeness of /2 and /2. The significance of nonnegligible /2 terms individually have more positive values than does | /2 | in general. However, the triple correlation /2 with each other out and make the /2 noisy than either /2 or /2 alone. Fig 6 shows /2 field less noisy than either /2 or /2 alone. Figure 6 shows /2 field less noisy than either /2 or /2 alone.

c. Eddy potential vorticity fluxes and eddy enstrophy sources and sinks

Examination of ( /2 ) with respect to /2 reveals interesting relationships between /2 and /2 at various depths (Fig. 1). The corresponding eddy enstrophy source/sink field, ( /2 ) /2, is shown in Fig. 5. As expected from values of /2 shown here, although there are some major differences between the two. The diagnoses of the wave activity and its fluxes are described in Nakamura and Chao (2000).

b. Eddy enstrophy

The time-mean eddy enstrophy, /2, at depths corresponding to Fig. 1 is shown in Fig. 3. In general, the structures of /2 reflect those of /2; large values of /2 are seen where /2 is large, and vice versa (see the tightness of /2 contour spacing in Fig. 1). The largest values are seen near the surface in the vicinity of the GS and the values decrease rapidly with depth. Large values of /2 at 577 m (Fig. 3d) and 721 m (not shown) are seen to the north of the approximate center of the tongue of large /2 (Fig. 1d). The narrow band of large /2 at 577 and 721 m lies under a region of large positive meridional gradient in /2 at shallower levels and is collocated with a region of significant negative meridional gradient in /2 between 463- and 721-m levels (dQ/dy at 577 m is shown in Fig. 4 as an example). The meridional /2 gradient in this region at deep levels is positive in general, creating a favorable condition for baroclinic instability, which may well account for the large values in /2 observed to the north of the center of the tongue of large /2 at 577 and 721 m. One should note, also, that a reversal in the sign of the meridional gradient of /2 in the meridional direction tends to support barotropic instability. Large meanders and cutoff eddies in the model, as well as the observed ones, have a vertically coherent structure, suggesting that they may well be triggered by barotropic instability.

At the intermediate depths, from 1000 to 2500 m or so, /2 shows large values in the vicinity of the boundary between low-q water mass to the west of the MAR and high-q water mass that lies to the east of it (Figs. 3e,f; see Fig. 3g for the location of the MAR). Stability analysis and experiments with idealized models suggest that these large values may be generated by baroclinic instability in the open ocean (Spall 1994). There is large /2 near the separation point around 37°N, 72°W generated by the direct influence of the GS down to 1500 m also (Fig. 3e). Below 2500 m, large values in /2 are observed below the model GS (Fig. 3g). Particularly large values at deep levels are seen near 45°N, 42°W (Fig. 3g). Movies of q show vigorous motions that are apparently wave-induced in this area, in agreement with the large values. Finally, we note that transient wave activity, M, defined by cos 2q/2 2 | /2 |, shows structures that are similar to /2 shown here, although there are some major differences between the two. The diagnoses of the wave activity and its fluxes are described in Nakamura and Chao (2000).
FIG. 1. Vector plots of $(\overline{Vq^*})_u$, superimposed on contours of $\overline{q}$ at (a) 17 m, (b) 140 m, (c) 370 m, (d) 577 m, (e) 1375 m, (f) 1875 m, and (g) 2875 m. Units of $\overline{q}$ and $(\overline{Vq^*})_u$ are, respectively, $1 \times 10^{-4}$ s$^{-1}$ and $1 \times 10^{-4}$ m s$^{-2}$. Contour intervals for $\overline{q}$ are (a) 0.2; (b) 0.5; (c,d) 0.2; and (e,f,g) 0.1.
(V_q^{\text{grad}})_d \cdot \nabla \tau \) (cf. Fig. 6 with Fig. 5c). This canceling tendency is found between \((V_q^{\text{grad}})_d \cdot \nabla \rho^{\text{op}}\) and \(V_q^{\text{grad}} \cdot \nabla \rho^{\text{op}}\) as well (not shown). Cronin and Watts (1996) also found that the triple correlation term is not negligible compared to the double correlation term in their analyses of observed temperature data in the vicinity of the GS. This raises a serious question on the validity of an assumption that the eddy self-advection term is negligible when real oceans are examined with an equation of a form similar to Eq. (4). We also note that the mean flow advection of eddy enstrophy, \(V_b \cdot \nabla (q^{\text{grad}}/2)\), which is balanced by \((V_q^{\text{grad}})_d \cdot \nabla \tau\), is comparable in magnitude with other terms, although \((V_q^{\text{grad}})_d\), itself is negligible compared to \((V_q^{\text{grad}})_d\) and \(V_q^{\text{grad}}\). Thus, removal of the \((V_q^{\text{grad}})_d\) term in the present calculation is crucial.

Near the surface, \((V_q^{\text{grad}})_d\) can be either up or down the gradient of \(\tau\) in an unorganized fashion (Fig. 1a). This can be better seen in plots of \((V_q^{\text{grad}})_d \cdot \nabla \tau\) (Fig. 5a). As seen in Fig. 5a, \((V_q^{\text{grad}})_d \cdot \nabla \tau\) is noisy in the entire domain at this level and in the rest of the mixed layer. As noted in Part I, a strong seasonal signal due to heating dominates both \(\tau\) and \(q^{\text{grad}}/2\) in the mixed layer, making interpretation of \((V_q^{\text{grad}})_d \cdot \nabla \tau\) difficult in the mixed layer. Below the top 50 m, \((V_q^{\text{grad}})_d\) begins to show more organized structures and tends to be downgradient of \(\tau\) in the vicinity of the GS. (For reference, \(\tau\) and \(V_b\) at 17 m are shown in Fig. 2. The location of the GS does not change noticeably with depth.) Approximately from 100 to 300 m, \((V_q^{\text{grad}})_d\) is predominantly down the gradient of \(\tau\) in the vicinity of the GS except for isolated regions that show considerable meanders in the contours of \(\tau\) (Figs. 1b, 5b), suggesting that eddy enstrophy is dissipating there and eddies tend to balance the loss of eddy enstrophy. Reflecting the strong time-varying input of \(q\) by the GS at the separation point around 37°N, 72°W, very large fluxes are emanating away from the separation point toward the interior along the meander in the contours of \(\tau\). Very small \((V_q^{\text{grad}})_d\) away from the GS are still unorganized, and \((V_q^{\text{grad}})_d \cdot \nabla \tau\) shows noisy structures with both positive and negative values scattered rather randomly.

At 370 m, \((V_q^{\text{grad}})_d\) begins to exhibit some up-\(\tau\)-gradient character in the vicinity of the GS and a strong tendency to orient itself along the contours of \(\tau\) (Fig. 1c). The scattered areas of large up-\(\tau\)-gradient \((V_q^{\text{grad}})_d\), that is, positive \((V_q^{\text{grad}})_d \cdot \nabla \tau\), lie within the tongue of large \(\tau\) (Fig. 5c). The region of predominantly negative \((V_q^{\text{grad}})_d \cdot \nabla \tau\) has spread southward and is significantly greater than that at upper levels. The southward spread of the area of predominantly negative \((V_q^{\text{grad}})_d \cdot \nabla \tau\) appears to be associated with the southward shift in depth of the tongue of large \(\tau\) discussed in Part I (Figs. 1b,c).

The area of the high-\(\tau\) tongue at these levels still appears mainly as an area of eddy enstrophy dissipation in the model. At 577 m, \((V_q^{\text{grad}})_d\) begins to exhibit a stronger up-\(\tau\)-gradient character in the vicinity of the GS, particularly to the north of the center of the large-\(\tau\) tongue (Figs. 1d, 5d). Although the areas of up-\(\tau\)-gradient \((V_q^{\text{grad}})_d\) are still far from coherent and occupy only roughly half of the northern half of the tongue of high-\(\tau\), they appear consistently to the north of the center of the tongue of high-\(\tau\) between 300 and 1000 m. In these areas of large positive \((V_q^{\text{grad}})_d \cdot \nabla \tau\), eddies tend to supply large values of \(q\) to areas of large \(\tau\) and balance the source of eddy enstrophy. In other words, these areas of up-\(\tau\)-gradient \((V_q^{\text{grad}})_d\), eddy potential enstrophy is the source of the mean potential enstrophy. Note that it is the region in which the meridional gradient of \(\tau\) is negative where \((V_q^{\text{grad}})_d\) shows a strong up-\(\tau\)-gradient character. It is also the region in which eddy potential energy and eddy kinetic energy are large. However, we note that the up-\(\tau\)-gradient character of \((V_q^{\text{grad}})_d\) in the northern half of the tongue of high-\(\tau\) may be simply a noise that arises from the lack of sufficiently long time series of obtain smooth mean and eddy fields. To the south of this narrow band of scattered large positive \((V_q^{\text{grad}})_d \cdot \nabla \tau\), \((V_q^{\text{grad}})_d\), is again almost entirely down the gradient of \(\tau\), balancing a sink of eddy enstrophy. This pattern of \((V_q^{\text{grad}})_d \cdot \nabla \tau\) along the tongue of large \(\tau\) shifts southward with depth until the tongue disappears at around 1000 m.

Also noteworthy in the thermocline is large \((V_q^{\text{grad}})_d\) emanating away from the vicinity of the separation point, around 37°N, 72°W. It apparently represents the transient component of the strong \(q\) input by the GS. The large \((V_q^{\text{grad}})_d\) in this region appears to play an important role in the large-amplitude meandering of \(\tau\). Here, \((V_q^{\text{grad}})_d\), is strongly up the gradient of \(\tau\) in an organized manner in the region of large meanders in contours of \(\tau\) and also in \(V_b\). The relationship between the mean and eddies in the region is very similar to that observed in atmospheric blocking (Nakamura 1998 and references therein), showing clear signs that eddies tend to reinforce the mean structure. It will be discussed later in detail as an important part of the mechanism responsible for the large meanders in the GS downstream of the separation point.

Below the top 1000 m, as the characteristics of the \(\tau\) field change substantially, patterns of \((V_q^{\text{grad}})_d\) and \((V_q^{\text{grad}})_d \cdot \nabla \tau\) also change noticeably (Figs. 1e,f and 5e,f). The structures in the vicinity of the GS at 1125 and 1375 m suggest large-amplitude eddies forcing the mean field, with large values of positive \((V_q^{\text{grad}})_d \cdot \nabla \tau\) and \((V_q^{\text{grad}})_d \cdot \nabla \tau\), and are reminiscent of the upper levels (Figs. 1e, 5e; the fields at 1125 and 1375 m are very similar to each other except for an island of \(\tau\) present at 1375 m). Indeed, an island of \(\tau\) is present at 1375 m, indicating the presence of strong mean-eddy interactions observed at upper levels (Fig. 1e). Large values of \(\tau\) extending from the northeastern corner of the domain due to the water mass influenced by the Mediterranean overflow and strong gradient of \(\tau\) associated with it dramatically...
change the picture of \( (\nabla_z q^*)_z \cdot \nabla \bar{q} \) at these levels. (Note that the model Mediterranean overflow is displaced to the north of the observed.) Fairly large values are now seen to the east of the MAR and in the areas that are affected by the Mediterranean overflow. There are some scattered areas of positive \( (\nabla_z q^*)_z \cdot \nabla \bar{q} \) to the east of the MAR, but \( (\nabla_z q^*)_z \) is predominately down the gradient of \( \bar{q} \) there. Large \( (\nabla_z q^*)_z \) occurs near the apparent boundary between two water masses: the water mass that appears to be influenced by the Mediterranean overflow and the mass that lies to the west of it. This is presumably due to baroclinic instability of weak jets in the open ocean (Spall 1994). Fairly large \( (\nabla_z q^*)_z \) is also seen where \( \bar{q} \) is large to the east of the MAR, almost uniformly pointing away from the area of large \( \bar{q} \). To the west of the MAR, where \( \bar{q} \) is dominated by \( f \cdot (\nabla_z q^*)_z \) is very small almost everywhere. At these levels, more up-\( q \)-gradient \( (\nabla_z q^*)_z \) is observed where the gradient of \( \bar{q} \) is small. This is illustrated in Fig. 7, which shows a scatterplot of \( |\nabla q| \) and \( (\nabla_z q^*)_z \). \( \nabla \bar{q} \), normalized by the absolute values of their respective averages, at 1375 m as a representative example for layers below 1000 m. Note how the points are spread on both positive and negative sides of \( (\nabla_z q^*)_z \cdot \nabla \bar{q} \) where the normalized QGPV gradient is 1.5 or less, whereas the points are scattered substantially more on the negative side where the normalized QGPV gradient is 2 or more. (The scatter diagram for the top 1000 m shows some hint of this characteristic, but is not convincing.) This suggests that eddy fluxes play an important role in creating \( \bar{q} \) structures with weak \( \bar{q} \) gradient, such as a pool of weakly varying \( \bar{q} \) and an island of \( \bar{q} \) at these levels.

Below 1500 m, the impact of \( q \) input by the GS is absent, and nearly the entire region to the west of the MAR shows very small \( (\nabla_z q^*)_z \). Between levels 1875 and 2875 m, there is large \( (\nabla_z q^*)_z \), radiating away from the area of large \( \bar{q} \) near the northeastern corner of the domain, suggesting that the role of \( (\nabla_z q^*)_z \) to the east of the MAR at these levels is dissipative, spreading large \( q \) from the source region (Figs. 1f, 4f). It is important to note that negligible \( (\nabla_z q^*)_z \) does not mean negligible wave activity flux. In the large domain at these levels to the west of the MAR where \( (\nabla_z q^*)_z \) is very small, transient wave activity fluxes have fairly large magnitudes (Nakamura and Chao 2000). Likewise, large \( (\nabla_z q^*)_z \) in the vicinity of the source region to the east of the MAR does not necessarily suggest large wave activity fluxes. In fact, wave activity fluxes in the region to the east and west of the MAR have the same order of magnitude (Nakamura and Chao 2000).

At 2875 m, \( (\nabla_z q^*)_z \) is large again to the west of the MAR and is very small to the east of the MAR, except for the northern end of the eastern side (Fig. 1g). Perhaps reflecting the predominately positive and strong meridional gradient of \( \bar{q} \), \( (\nabla_z q^*)_z \) is almost all down the
Fig. 3. Contours of $\overline{q^2}$ at (a) 17 m, (b) 140 m, (c) 370 m, (d) 577 m, (e) 1375 m, (f) 1875 m, and (g) 2875 m. Units of $\overline{q^2}$ are $1 \times 10^{-10}$ s$^{-2}$. Contour values shown are (a,b) 20, 40, 60, 80, 100, 120, 140, 160, 180, 200, 250, and 300; (c) 1, 5, 10, 15, 20, 25, 30, and 35; (d) 0.5, 1, 2, 3, 4, and 5; and (e,f,g) 0.2, 0.5, 1, 1.5, 2, 2.5, 3, 3.5, and 4.
gradient of $\nabla q$ to the west of the MAR. Therefore, eddy fluxes are mostly dissipative and tend to convert the mean enstrophy into eddy enstrophy here. Large $(\bar{V}_q q^*)_d$ is seen mostly near the western boundary, radiating away from the boundary. Particularly large $(\bar{V}_q q^*)_d$ is seen just off the western boundary near 48°N. Visual examination of $q$ evolution shows vigorous wave motions in this region. This is also the region of maximum eddy potential enstrophy at this level (Fig. 3g). On the other hand, to the east of the MAR, $(\bar{V}_q q^*)_d$ is very small and does not show organized characteristics with respect to the mean gradient. Wave motions are very weak in this region also. The only exception is the northern edge of the eastern side, where eddy fluxes are fairly large and dissipative. Farther below this level, the basic characteristics of $(\bar{V}_q q^*)_d$ and $(\bar{V}_q q^*)_d \cdot \nabla q$ are the same as those at 2875 m, except that $(\bar{V}_q q^*)_d$ at lower levels shows the down-$\nabla q$-gradient nature more clearly to the west of the MAR and is extremely small everywhere to the east of the MAR. These characteristics are observed down to the bottom of the domain.

Due to the expected long evolution timescales of the oceanic flows at various depths, one must note that the results presented here suffer from somewhat noisy mean fields and very noisy eddy covariance fields that arise from the relatively short period considered here. The annual mean flow of the five years considered here shows some sign of low-frequency variability at all depths. Also, deep levels of the model are almost certainly not in equilibrium after 30 years of integration, which probably means that the upper-level flows are not in equilibrium either since the deep flow presumably exerts substantial influence on the upper flow through its impact on the deep PV structure. To assess the sensitivity of the present results to the time period considered, we repeated selected diagnostic calculations using only the first three years of the 5-yr output. We find that the overall features remain the same at all depths, both qualitatively and quantitatively. However, close inspections reveal many differences in the detail of the noisy structures. Also, the noisiness of the results of the 3-yr diagnoses is indistinguishable from that of the 5-yr diagnoses, suggesting that much longer time series are needed just upstream of the background blocking flow, which is characterized by a large poleward meander of the background flow, reinforce the blocking pattern. The

well show some differences in the mean fields. It is important to note that the 5-yr mean flow and eddy statistics about the 5-yr mean should be interpreted as a low-frequency state and high-frequency eddy statistics for the given low-frequency state. We believe, however, that the gross features shown here would be present in diagnoses of a much longer time series or another 5-yr period also. In particular, the large areas of $q^{2/2}$ sink in the southern half of the tongue of high $\nabla q$ in the thermocline and east of the MAR in the intermediate layer appear to be robust features in the model ocean.

d. Eddy forcing of large meanders in the model GS

Large meanders in the GS just downstream of the separation point are an intriguing feature observed in the real ocean (e.g., Halliwell and Mooers 1983; Watts et al. 1995) and are also produced by the model, although the amplitude of the meanders in the model is overestimated. The meanders in the model are quasi-stationary, while their amplitudes do fluctuate in time. The meanders in the model produce excessive numbers of cold-core rings at the troughs, perhaps due to the underestimated amplitude. These cold-core rings are seen to propagate westward and eventually merge with the GS near the separation point, interacting with the GS in complex ways. The unrealistically large amplitude of the meanders is likely to contribute to producing other unrealistic features in the model. For example, it is very likely to reduce the downstream transport of heat and salt by the GS since momentum loss in the vicinity of the meanders is probably overestimated. If the momentum loss near the meanders is overestimated, it tends to weaken the model GS farther downstream as well.

The mechanism responsible for the meanders is not obvious from the current diagnoses. However, the diagnoses suggest that eddies may be reinforcing the meanders in a way similar to that observed in atmospheric blocking phenomena (e.g., Shutts 1983; Illari and Marshall 1983, Haines and Marshall 1987; Nakamura et al. 1997). In idealized models of atmospheric blocking, eddies are deformed by diffusional large-scale flow and deposit potential vorticity in such a way that they reinforce the large-scale diffusive flow pattern as the eddies break and dissipate in the region (e.g., Shutts 1983; Haines and Marshall 1987). This pattern of eddy forcing is observed in diagnoses of atmospheric blocking events also (e.g., Nakamura and Wallace 1990; Nakamura et al. 1997). A common feature in the structure of $(\bar{V}_q q^*)_d \cdot \nabla q$ or equivalent in the vicinity of blocking shown by Shutts (1983), Illari and Marshall (1983), and Nakamura (1998) is a region of up-$\nabla q$-gradient eddy fluxes just upstream of the region of the blocking ridge or just downstream of the peak of the trough that sits immediately upstream of the blocking ridge. Eddies generated just upstream of the background blocking flow, which is characterized by a large poleward meander of the background flow, reinforce the blocking pattern. The
Fig. 4. Contours of $\varpi$ at 577-m level. Negative values are indicated by dashed contours. Units are $10^{-4}$ s$^{-1}$ (100 km)$^{-1}$. The contour interval is 0.04.

up-$\varpi$-gradient eddy fluxes are presumably balancing the source of eddy potential enstrophy that tends to generate eddies there. Downstream of the region of up-$\varpi$-gradient eddy fluxes, in the region of weak zonal flow, down-$\varpi$-gradient eddy fluxes are observed in their diagnoses, suggesting eddy enstrophy dissipation or absorption by the mean.

In the current diagnoses of the model GS, we observe very similar structures. Shown in Figs. 8 and 9 are, respectively, $\varpi$ with $\nabla \cdot$ and $(\nabla \cdot)^r$, near the meanders at 140 and 577 m. Figure 10 shows the corresponding fields of $(\nabla \cdot)^r$. These two levels are shown because the former and the latter represent, respectively, situations with primarily positive and negative meridional gradient of $\varpi$. At both levels, large $(\nabla \cdot)^r$ emanates away from the vicinity of the separation point with some sense of anticyclonic rotation along the GS and contours of $\varpi$. These fluxes are almost all down the gradient of $\varpi$ in the region of northward turn of the model GS, in a manner very similar to that reported by Shutts (1983), Illari and Marshall (1983), and Nakamura (1998).

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Although weaker than the first pair of meanders (a trough and a ridge) after the separation, the same structures are seen in the region of the second pair of meanders also. These structures suggest that the relationship between the mean and eddies in this vicinity may be qualitatively the same as that found in the atmospheric blocking; that is, eddies tend to reinforce the meandering mean flow. The structures of $(\nabla \cdot)^r$ suggest that eddy potential enstrophy is generated in regions just upstream of the time-mean ridges (see large positive values centered near 69° and 64°W) created by the meandering GS and dissipated or absorbed by the mean in regions of the time-mean ridge and the upstream half of the trough (see large negative values centered near 71° and 66°W). Divergent heat fluxes calculated by Cronin (1996) in the vicinity of a trough in the observed GS show some hint of the upgradient character just downstream of the trough also [see Figs. 3 and 4 of Cronin (1996)].

The same structures are observed in $(\nabla \cdot)^r$, suggesting that eddy potential energy sources and sinks in the vicinity of the model GS have qualitatively the same spatial distributions as those of eddy potential enstrophy. In fact, the spatial relationship between the regions of source and sink of eddy potential energy and the phase of the meanders appears more robust than that between eddy potential enstrophy and the meanders. We also observe a clear sign of transient wave activity flux convergence in the meandering ridges in a manner very similar to that reported by Nakamura et al. (1997) and Nakamura and Chao (2000).

In studies of the atmospheric blocking, it has been emphasized that the background flow tends to organize the eddy forcing on the background flow in such a way that the eddy forcing is a positive feedback to the block-
Fig. 5. Contours of eddy enstrophy source/sink, $S_E$, calculated by $(\nabla E \cdot \nabla)$, at (a) 17 m, (b) 140 m, (c) 577 m, (d) 1375 m, (e) 1875 m, and (g) 2875 m. Units of $(\nabla E \cdot \nabla)$ are $1 \times 10^{-12}$ s$^{-2}$ day$^{-1}$. Contour values shown are (a) -200, -100, -50, -40, -30, -20, -10, -5, -2, -1, -0.5, 0, 0.5, 1, 2, 5, 10, 20, 30, 40, 50, 100, and 200; (b,c,d) -200, -100, -50, -40, -30, -20, -10, -5, -2, -1, 0, 1, 2, 5, 10, 20, 30, 40, 50, 100, 200, and 400; and (e,f,g) -100, -50, -20, -10, -5, -2, -1, -0.5, -0.2, -0.1, -0.05, -0.02, -0.01, 0, 0.01, 0.02, 0.05, 0.1, 0.2, 0.5, 1, 2, 5, 10, 20, 50, and 100. Positive values (source) and negative values (sink) are shown by, respectively, red solid contours and blue dashed contours. Zero contours are solid black.
Fig. 6. Contours of (a) \( \nabla (v'q') \cdot \nabla q' \) and (b) \( \nabla (v'q') \cdot \nabla q' \) at 370 m. Units of \( \nabla (v'q') \cdot \nabla q' \) and \( \nabla (v'q') \cdot \nabla q' \) and contour values shown are the same as those in Fig. 5c. Positive and negative values are shown by, respectively, red solid contours and blue dashed contours. Zero contours are solid black. Note that the sum of the fields shown in (a) and (b) is the field shown in Fig. 5c.
ing flow (e.g., Shutts 1983; Illari and Marshall 1983; Nakamura et al. 1997). The results presented here suggest that the same relationship may exist in the vicinity of the large time-mean meanders near the separation point. If this is the case, the results suggest some plausible causes of the overestimated amplitude of the meanders in the model GS. One possibility is that insufficient model resolution for baroclinic waves of deformation radius scales results in underestimated zonal acceleration of the GS after its separation from the boundary. It may create an imbalance between the downstream advection of eddy enstrophy and input of eddy potential enstrophy immediately downstream of the separation point and trigger growth of the quasi-stationary meanders through the positive feedback action of eddies. Once the meanders are created, eddies will keep reinforcing the meanders, in which dissipation of eddy potential enstrophy is enhanced to compensate for the underestimated downstream advection of eddy potential enstrophy by the GS. Since such meanders maintain a favorable condition for local accumulation of eddy potential enstrophy or wave activity (e.g., Nakamura et al. 1997), the meanders will remain in place while repeating growth and decay in amplitude in the absence of sufficient forcing to generate a strong zonal flow.

Model studies (Schmitz and Thompson 1993; Smith et al. 2000) demonstrate that a horizontal resolution of 0.1° may be required to obtain realistic eddy kinetic energy release by baroclinic eddies in this region, suggesting that the lack of resolution may indeed play an important factor in the large meanders. We must note, however, that wave breaking and generation of filaments of fluid required for the eddy forcing of a large-amplitude background meander is resolved better by higher resolution. Therefore, higher resolution itself may not resolve the problem of the overestimated amplitude. Another possibility is that the topographic forcing of large-scale waves near the separation point is overestimated in the model, resulting in further growth of the waves or meanders through their capability to organize baroclinic waves of deformation radius scales in such a way to reinforce themselves. The bottom flow in this region is fairly strong, and the topographic forcing of waves associated with this flow may be sufficiently strong to overcome the downstream advection of wave activity by the model GS.

4. Summary and concluding remarks

We have examined the output of an eddy-resolving GCM of the North Atlantic in a framework based on quasigeostrophic potential vorticity, $q$, and quasigeostrophic potential enstrophy, $q^2/2$, to study large-scale circulations in the vicinity of the GS with an emphasis on the relationship between the mean and eddies. In general, the time-mean fields of quasigeostrophic eddy potential enstrophy, $q^2/2$, have structures that are very similar to those of $\nabla q\overline{q}$; that is, $q^2/2$ is large where $\nabla q\overline{q}$ is large, and vice versa. Values of $q^2/2$ are very large near the surface and rapidly decrease with depth.

The modified “divergent” eddy flux defined by Nakamura (1998), $(\nabla q^2)^*$, whose dot product with $\nabla q\overline{q}$ balances source/sink of eddy potential enstrophy on a
Fig. 8. Contours of $\nabla$ and vectors of $\nabla_s$ in the vicinity of the large-amplitude meanders in the model GS just downstream of the separation point at (a) 140 and (b) 577 m. Units of $\nabla$ and $\nabla_s$ are, respectively, $1 \times 10^{-4}$ s$^{-1}$ and m s$^{-1}$. The contour intervals for $\nabla$ are (a) 0.5 and (b) 0.1.
Fig. 9. Contours of $\overline{\gamma}$ and vectors of $(\nabla \cdot \mathbf{q})_d$ in the vicinity of the large-amplitude meanders in the model GS just downstream of the separation point at (a) 140 and (b) 577 m. Units of $\overline{\gamma}$ and $(\nabla \cdot \mathbf{q})_d$ are, respectively, $1 \times 10^{-5}$ s$^{-1}$ and $1 \times 10^{-4}$ m s$^{-1}$. The contour intervals for $\overline{\gamma}$ are (a) 0.5 and (b) 0.1.
Fig. 10. Contours of $S_e$ in the vicinity of the large-amplitude meanders in the model GS just downstream of the separation point at (a) 140 and (b) 577 m. Units of $S_e$ are $1 \times 10^{-12} \text{ s}^{-2} \text{day}^{-1}$. Contour values shown are (a) $-2000, -1500, -1000, -500, -200, -100, -50, -20, -10, 0, 10, 20, 50, 100, 200, 500, 1000, 1500, \text{and} 2000$; and (b) $-50, -20, -10, -5, -2, -1, -0.5, -0.2, -0.1, 0, 0.1, 0.2, 0.5, 1, 2, 5, 10, 20, \text{and} 50$. Positive and negative values are shown by, respectively, red solid contours and blue dashed contours. Zero contours are solid black.
horizontal plane, was calculated and examined along with \( \left( \nabla \cdot \mathbf{q} \right)_0 \cdot \nabla \theta \). In general, \( \left( \nabla \cdot \mathbf{q} \right)_0 \), and \( \left( \nabla \cdot \mathbf{q} \right)_0 \cdot \nabla \theta \) are large near the surface and decrease rapidly with depth. Near the surface, \( \left( \nabla \cdot \mathbf{q} \right)_0 \) is large and has noisy structures, being up and down the gradient of \( \theta \) in an unorganized fashion. From approximately 100 to 300 m, \( \left( \nabla \cdot \mathbf{q} \right)_0 \) is predominantly down the gradient of \( \theta \) in the vicinity of the GS except for small areas in the vicinity of large meanders in the model GS, suggesting that eddy fluxes tend to balance the sink of \( \mbox{q}^2 / 2 \) by converting from the mean enstrophy. Away from the GS, \( \left( \nabla \cdot \mathbf{q} \right)_0 \) is unorganized at these levels. Between 370 and 901 m, \( \left( \nabla \cdot \mathbf{q} \right)_0 \) shows some up-\( \theta \)-gradient character to the north of the center of the tongue of high \( \theta \) along the GS. Thus, there may be some systematic conversion of eddy enstrophy into the mean enstrophy in the northern half of the tongue or island of high \( \theta \) in the thermocline. Except for some areas of up-\( \theta \)-gradient \( \left( \nabla \cdot \mathbf{q} \right)_0 \), \( \left( \nabla \cdot \mathbf{q} \right)_0 \) is almost entirely down the gradient of \( \theta \) in the tongue of high \( \theta \), indicating \( \mbox{q}^2 / 2 \) dissipation in most of the tongue. Below 1000 m, characteristics of \( \left( \nabla \cdot \mathbf{q} \right)_0 \) change again as those of \( \theta \) and \( \mbox{q}^2 / 2 \) change dramatically. From approximately 1000 to 2500 m, \( \left( \nabla \cdot \mathbf{q} \right)_0 \) is very small and unorganized with respect to \( \theta \) in most of the region to the west of the MAR, whereas \( \left( \nabla \cdot \mathbf{q} \right)_0 \) is relatively large and predominantly down the gradient of \( \theta \) to the east of the MAR. The predominantly down-\( \theta \)-gradient characteristic of \( \left( \nabla \cdot \mathbf{q} \right)_0 \) to the east of the MAR reflects a mean-dissipating role of eddies and sink of \( \mbox{q}^2 / 2 \) in the region. Finally, below 2500 m, as substantial gradient in \( \theta \) is reestablished to the west of the MAR, \( \left( \nabla \cdot \mathbf{q} \right)_0 \) is fairly large and mostly down the gradient of \( \theta \) to the west of the MAR, indicating a mostly mean-dissipating role of eddies at these deep levels. On the contrary, to the east of the MAR where \( \theta \) is small and dominated by the planetary vorticity, \( \left( \nabla \cdot \mathbf{q} \right)_0 \) is very small and unorganized with respect to the \( \theta \) field.

Careful inspections of \( \left( \nabla \cdot \mathbf{q} \right)_0 \cdot \nabla \theta \) reveal that the role of eddies in the region of large meanders in the GS immediately downstream of the separation point is qualitatively the same as that of atmospheric eddies in a phenomenon called blocking that manifests itself as a large quasi-stationary poleward meander of the subtropical jet. Spatial relationships between \( \left( \nabla \cdot \mathbf{q} \right)_0 \cdot \nabla \theta \) and the meander phases and those between \( \left( \nabla \cdot \mathbf{q} \right)_0 \cdot \nabla \rho \) and the meander phases suggest that eddies are organized by the meanders in such a way that the eddies tend to reinforce the meanders. These results suggest that the overestimated amplitude of the time-mean meanders in the model may result from a slight deficiency in the model, amplified by the positive feedback of the transient eddies.

While the manuscript was in review, we extended the model integration by four years and repeated the same diagnoses on the 4-yr model output. We found the gross patterns discussed here in the diagnoses of the four additional years of the simulated ocean. The 4-yr mean circulation is very similar to the 5-yr mean presented here. Also, the qualitative aspect of the eddy statistics presented here is found in the new calculation. Thus, the qualitative picture of the relationship between the large-scale low-frequency circulation and eddies presented in this paper is likely to be robust so long as the large-scale low-frequency circulation does not change substantially. Of course, to what extent the current results represent reality is a different issue. The overestimated meander amplitude of the GS and underestimated deep GS in the model are likely to be accompanied by unrealistic features in the eddy statistics. The reader needs to be cautious in interpreting the results presented here for the real ocean.

A potentially questionable aspect of the results is the noisiness in the 5-yr mean of various eddy-related quantities, for example, \( \left( \nabla \cdot \mathbf{q} \right)_0 \cdot \nabla \theta \) shown in Fig. 5. The noisiness mostly arises from various combinations of several factors that depend on the location. Five factors that are likely to be most important are the following: 1) Eddy motions are not sufficiently rapid to complete many cycles of growth and decay during the analyzed period; 2) the raw “data” sampling frequency is not sufficiently high with respect to eddy motion frequencies; 3) eddies are small; 4) eddies have preferred locations of growth and decay; and 5) eddies actively interact with the low-frequency flow. We believe that factor 2 is likely to be negligible in our case since the timescale of eddy motions is tens of days and our model output is available every three days for five years. Also, since the “data” used here are model output, the noisiness is not due to errors in the “data.” In other words, the 5-yr mean fields shown here have no uncertainties in a mathematical or statistical sense so long as factor 2 is negligible. It is likely that combinations of the other four factors, the importance of which depends on the location, produce the noisiness in the eddy-related statistics. Instantaneous values of these eddy-related quantities are much larger than their 5-yr mean values and fluctuate rapidly in time. Except for squared deviation quantities, such as \( \mbox{q}^2 \), they fluctuate between positive maximum and negative maximum. Eddy covariances typically show this fluctuating character. Naturally, the time mean of an eddy covariance or its derivative tends to be much smaller than its instantaneous values, because of the strong tendency for positive and negative values to cancel out over a long period. (The time mean of an eddy quantity is, of course, zero by definition). This nature of eddy covariances makes the time mean fields of these quantities very sensitive to any of the five factors mentioned above. To address the issue of noisy 5-yr mean eddy-related fields, one would need to repeat the diagnoses on many 5-yr periods and examine the variability of the low-frequency state and associated eddy statistics. It would also be very useful for learning the dynamic interactions between the low-frequency and high-frequency flows. Such a study is, however, beyond
the capacity of our computational resources and beyond the scope of the present work.

In the present work, we have attempted to demonstrate the potential usefulness of pseudodata in studying oceanic circulations and also to show how diagnoses of the pseudodata may be useful for studying a potential mechanism responsible for a feature observed in the real ocean or an unrealistic feature generated in the model. Extensive measurements of oceanic properties at intervals sufficient to resolve the important eddies are not likely to materialize in the foreseeable future. In the light of this limitation in the data availability, utilizing eddy-resolving model output with caution may be of substantial merit in studying large-scale oceanic circulations.

Acknowledgments. The research described in this paper was carried out by the Jet Propulsion Laboratory (JPL), California Institute of Technology, under contract with the National Aeronautics and Space Administration. Computations were carried out on the 256-processor Cray T3D through the JPL Supercomputing Project. M. Nakamura is indebted to John Marshall for the stimulating conversations that led to this series of diagnoses. We wish to thank Mike Spall for helpful comments on the draft of this paper. Finally, we wish to thank Meghan Cronin and an anonymous reviewer for helpful comments and suggestions on the manuscript.

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