Impact of the North Atlantic Oscillation on the oceanic eddy flow: dynamical insights from a model-data comparison

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ABSTRACT

Observational studies based on altimeter data have shown that in many regions of the World Ocean the oceanic eddy kinetic energy (EKE) significantly varies on interannual timescales. Comparing altimeter-based EKE maps for 1993 and 1996, Stammer and Wunsch (1998) discuss a significant meridional redistribution of EKE in the North Atlantic and speculated about the possible influence of the NAO cycle. This hypothesis is examined using 7 years of T/P altimeter data and three 1/6°-resolution Atlantic Ocean model simulations from the French CLIPPER numerical experiment performed over the period 1979-2000. The subpolar-subtropical meridional contrast of EKE actually appears to vary on interannual timescales in the real ocean, and the model reproduces it realistically. The NAO cycle forces the meridional contrast of energy input by the wind. Our analysis suggests that after 1993, the large amplitude of the NAO cycle induces an adjustment of the large-scale circulation (Gulf Stream/North Atlantic Current), of its baroclinically unstable character, and, in turn, of the distribution of EKE. Model results suggest that before 1994, most NAO-like fluctuations were not followed by EKE redistributions, probably because NAO oscillations were weaker. Strong NAO index modulations are followed by gyre-scale EKE fluctuations with a 9-10 months lag, suggesting that complex, nonlinear adjustment processes are involved in this oceanic adjustment.

1. Introduction.

Altimeter observations are crucial for the study of the ocean variability at different space and time scales. Historically limited to regional and temporary in-situ surveys, the monitoring of mesoscale motions has become global and quasi real-time. Sea-surface height (SSH) data provide information about the intensity of the surface eddy activity via the eddy kinetic energy (EKE), i.e. the variance of geostrophic surface velocities. Over most of the World Ocean, surface EKE maxima are due to the instability of the main currents. Reciprocally, mesoscale motions affect the path and intensity of the main currents, their interaction with topography (Dewar, 1998; de Miranda et al, 1999-a), the redistribution and dissipation of potential vorticity (Rhines and Young, 1982), convection (Legg et al, 1998), subduction (de Miranda et al, 1999-b), transport (e.g. Agulhas Rings, Tréguier et al, 2002), and mixing of water masses. The latter processes have timescales comparable to (and are involved in) the oceanic variability at interannual and longer timescales. To understand the climatic system thus requires a better knowledge of the interannual variability of mesoscale turbulence in the ocean, i.e. its origins and, ultimately, its effects.

Because most current meter and drifter data are sparse and generally limited to a few years at most, the variability of the eddy flow has been studied more regionally and seasonally than at basin-scale on interannual timescales. The EKE was shown to follow the seasonal fluctuations of the local wind stress (its intensity or curl) with a few months lag: 4 months in the eastern North Atlantic (Richarson, 1983), 3 months within the Gulf Stream/North Atlantic Current (GS/NAC) system (Garnier and Schopp, 1999), 2 months in the Labrador Sea (White and Heywood, 1995), and 1-3 months in the Rockall Trough (Dickson et al, 1986). The seasonal fluctuations of EKE are thus partly controlled by the wind variability in the North Atlantic. Except in some localized regions, such as the East Greenland Current where the EKE fluctuates in phase with the wind forcing (see White and Heywood, 1995), the existence of such a lag led the authors to conclude that the wind drives the seasonal variability of EKE indirectly and/or remotely, i.e. through an adjustment of the large-scale flow and/or via propagative processes. However, these processes are complex and have not been clearly identified so far.

Fewer studies have been dedicated to the interannual variability of EKE. A remarkably long (10 years) current meter record was collected in the vicinity of the Azores Current (Müller and Siedler, 1992), revealing a persistent and strong decrease of EKE during the eighties. This trend could not be explained unequivocally because of the very local character of the dataset. Based on altimeter data, White and Heywood (1995), Garnier and Schopp (1999) and Ducet and Le Traon (2001) have described the year-to-year evolution of EKE over parts the North Atlantic. The interannual variability of the eddy flow was described by Garnier et al (2002) in the upper South Atlantic from a 5-year altimeter data assimilation experiment. An extensive, global-scale description of interannual variability of the surface eddy field was provided by Stammer and Wunsch (1998, noted as SW98 hereafter) from the four years (1993-1996) of TOPEX/Poseidon altimeter data available at that time. These authors highlighted the significant interannual variability of the eddy field in many regions of the World Ocean, and suggested that the EKE decrease detected in the Azores Current area by Müller and Siedler (1992) is due to large-scale adjustment of the ocean circulation to fluctuating forcing fields. Partly because those studies were based on observations only, the origin of this interannual EKE variability and its possible connection with the atmospheric forcing and/or the large-scale circulation are still under debate.

An intriguing feature was outlined by SW98 in the North Atlantic and the North Pacific (their figure 1-c): compared with the 1993 pattern, the 1996 eddy kinetic energy (EKE) is weaker in subpolar regions and stronger in the subtropics by a significant amount (20-30% at gyre scale and up to 70% locally). This dipolar pattern is basin-wide and extends over 10-20° meridionally. As pointed out by SW98 for the Atlantic sector, 1993-1996 is actually the period when the winter NAO index suddenly decreased and became strongly negative. Accordingly, both the mean westerlies and the associated storm track were found 10 to 20° further south in 1996 compared to 1993. The shift of the oceanic turbulence in the North Atlantic might thus be correlated with this atmospheric mode, but SW98 could not demonstrate this due to the short and superficial character of the record. As was shown on seasonal timescales, wind fluctuations seem to have a strong impact also on the interannual fluctuations of oceanic EKE.

The present study is focused on this North Atlantic dipolar EKE pattern. We perform new diagnostics based on (i) longer TOPEX/Poseidon time series and on (ii) the outputs of a numerical model of the Atlantic forced by realistic surface fluxes over the last 2 decades. The aim of the present study is to provide answers to the following questions:

1. Is the model able to simulate (and thus to provide a more detailed description of) the recent EKE evolution diagnosed by SW98?
2. If so, has this event been forced by the concomitant NAO transition (as suggested by SW98) and has this connection been robust over the last two decades?
3. What does the model-data comparison tell about this interannual variability?

The numerical model and altimeter dataset are presented in sections 2 and 3 respectively. Section 3 describes the processing applied to the model and observed datasets to quantify the interannual variability of the eddy field. The horizontal distribution and temporal evolution of the eddy field in the observed and simulated datasets are presented in section 4. The link between the large-scale distribution of the eddy flow and the NAO index is discussed in section 5, and conclusions are given in section 6.

2. Model Configuration and Experiments.

The numerical data used in this study were produced during the CLIPPER experiment (Tréguier et al, 1999). The model used is OPA8.1 (Madec et al, 1998), a geopotential-coordinate primitive equation model with rigid lid, implemented on a 1/6° Mercator grid on the whole Atlantic with forty-two levels in the vertical. The vertical grid spacing increases from 12 m at the surface to 250 m below 1500 m. Horizontal diffusion and viscosity are parameterized as biharmonic operators. The vertical mixing coefficient is given by a second-order closure scheme (Blanke and Delecluse, 1993), and is enhanced in case of static instability. The bottom topography (Fig. 1) is based on the Smith and Sandwell (1997) database. The domain is limited by four open boundaries located at 70°N, in the Gulf of Cadiz (8°W), at the Drake Passage (68°W) and between Africa and Antarctica (30°E). These Orlanski-type open boundaries radiate perturbations outward and relax the model variables to a climatological reference. Details about the implementation and behavior of the open boundaries are given in Tréguier et al (2001). Low-passed model outputs are saved every 5 days, with the output itself being a 5-day average in order to avoid the aliasing of high-frequency processes (Crosnier et al, 2001).

The model is started from rest with temperature and salinity fields taken from the Reynaud et al (1998) climatology, and then spun-up for 8 years with a climatological, low-passed (cutoff period at 10 days), periodic daily forcing computed from ERA15 ECMWF reanalysis. In the experiment labeled HF (high-frequency), the model is restarted from the end of the spinup and integrated between 1979 and the end of 1993 using the daily forcing fields from ECMWF reanalyses interpolated at every timestep. The HF simulation is forced by ECMWF analyses between 1994 and 2000. The continuity of wind stress time series on January 1st, 1994 was verified at each latitude, and the continuity of heat and salt fluxes was insured by replacing the 1994-2000 temporal means by its 1980-1993 counterpart from the reanalysis. A second integration, labeled LF (low-frequency), forced was started from the HF state at March 21, 1992 until the end of year 1999. A low-pass Lanczos filter with a 35-day cutoff period was applied to the HF surface forcing (wind stress, heat and salinity fluxes) to generate LF forcing fields in which temporal variability at scales shorter than one month were filtered out (Fig. 2). A third simulation, labeled PF (periodic forcing) is a continuation of the spinup: it was forced between 1979 and the end of 2000 by the mean ERA15 ECMWF reanalysis to provide some complementary information about the variability generated by a periodic seasonal cycle. Table 1 summarizes the three model integrations.

In the three simulations, the wind stress is applied as a boundary condition in the momentum equations. Heat and virtual salinity fluxes are imposed as described in Barnier (1998): ECMWF fluxes are introduced as source terms in the temperature and salinity equations at the uppermost level and corrected by a retroaction term. This term is proportional to the difference between the tracer value in the model (sea-surface temperature (SST) or salinity (SSS)) and an observed value: weekly SST from Reynolds et al (1995) for temperature and seasonal climatological SSS from Reynaud et al (1998) for salinity. The proportionality coefficient depends on time and space as explained in Barnier (1998).

3. TOPEX/Poseidon data, definition of EKE and Processing

We made use of a set of sea level anomaly (SLA) maps deduced from TOPEX/Poseidon altimeter time series, available every 10 days between 10/22/1992 and 12/29/2000 at a resolution of 0.25° by 0.25°. These fields were built using an improved space/time objective analysis method that takes into account long wavelength errors correlated noise (Le Traon, Nadal and Ducet, 1997). Since SLAs are not directly simulated by our rigid-lid model, we will compare the simulated and observed eddy activities from near-surface geostrophic velocities, which derive from SLA data at time *it*:

$$\begin{bmatrix} u_{it} \\ v_{it} \end{bmatrix} = -\frac{g}{f}\nabla(SLA_{it}).$$

Model velocities are considered at 55 m, i.e. below the ageostrophic Ekman layer. More precisely, we compare the variances of these two velocity fields, i.e. the observed and simulated eddy kinetic energies (EKEs).

EKE^0 time series are often (e.g. SW98; Garnier and Schopp 1999) derived from geostrophic velocity time series $[u_{it}, v_{it}], it \in [1, Nt]$ as follows:

$$EKE_{it}^0 = [(u_{it} - \bar{u})^2 + (v_{it} - \bar{v})^2]/2,$$

where time averages are computed over *Nt* (records) as

$$\begin{bmatrix} \bar{u} \\ \bar{v} \end{bmatrix} = \frac{1}{Nt} \sum_{it=1}^{Nt} \begin{bmatrix} u_{it} \\ v_{it} \end{bmatrix}.$$

The resulting EKE^0 time series quantifies the fluctuations of the eddy kinetic energy. However, all timescales resolved in the dataset contribute to EKE^0, i.e. not only those associated with mesoscale turbulence (in which most authors are interested), but also those with interannual velocity fluctuations. The latter should not be assumed small a-priori. In the North Atlantic for example, the number, location, and transport of the branches of the North Atlantic Current are highly variable (Sy et al., 1992; White and Heywood, 1995). The location and intensity of the Azores Current are also subject to interannual variability (Klein and Siedler, 1989). The contribution of interannual velocity fluctuations can be removed from EKE estimates while keeping the mesoscale contribution by computing successive annual EKEs (velocity variances over one year). This is the choice made by SW98, but this would provide only 7 EKE estimates over our 7-year T/P time series: the high temporal resolution provided by EKE^0 is lost.

We thus computed EKE_{it}^{y} estimates as running variances of the velocity field over overlapping 1-year time intervals at time *it*:

$$EKE_{it}^{y} = [(u_{it} - \bar{u}_{it}^{y})^2 + (v_{it} - \bar{v}_{it}^{y})^2]/2, \text{ where}$$

$$\begin{bmatrix} \bar{u}_{it}^{y} \\ \bar{v}_{it}^{y} \end{bmatrix} = \frac{1}{1year} \sum_{i=it-6months}^{it+6months} \begin{bmatrix} u_i \\ v_i \end{bmatrix}$$

is the running mean of velocities. The above quantities were computed every 5 days from the model outputs (every 10 days from T/P data) throughout our model and observed datasets (excluding the first and last 6 months). The EKE^{y} time series is well designed to monitor the intensity of mesoscale turbulence since timescales longer than one year have no contribution. This quantity also shows the evolution of simulated and observed EKEs with high temporal resolution. However, the seasonal cycle of the velocity field is not monochromatic and has timescales that overlap those of mesoscale turbulence. Both processes are thus impossible to separate properly, even by defining running variances over six-month windows (as we did in a preliminary sensitivity study), or by subtracting a mean seasonal cycle from the velocity time series before variance computations. The seasonal cycle of velocity is thus merged with the mesoscale signal in each EKE_{it}^{y} and does not appear in the resulting time series. This is satisfactory since we are interested in the interannual variability of EKE. EKE will refer to EKE^{y} in the following, unless stated otherwise.

We actually verified that north of 20°N, interannual EKE fluctuations are mainly due to mesoscale variance and not dominated by seasonal velocity fluctuations. Indeed, local EKEs have been computed also from a 1°-resolution CLIPPER simulation initialized and forced exactly as the present simulation. This coarse-resolution EKE quantifies the variance of velocity fluctuations without any turbulent contribution, and turns out to be 15 to 40 times weaker than in the 1/6° run. Moreover, there is no phase link between the EKE fluctuations diagnosed from both runs, showing that mesoscale turbulence is largely responsible for the features described in the present paper.

To summarize, 10-day/0.25°-resolution T/P SLA maps were treated as follows: Computation of anomalous geostrophic velocities u' from SLA maps, bilinear interpolation of u' on the isotropic model grid (1/6° resolution), and computation of EKETP, the running variance of velocities over 1-year windows every 10 days between 04/30/1993 (6 months after the first available SLA map) and 01/04/2000. Model velocities are available every 5 days at 55 m depth. EKEmodel, the running variance derived from these velocity fields was computed exactly like EKETP between 07/07/1980 and 07/03/1999. EKEmodel was also estimated from the same velocity time series sub-sampled at a 10-day period (like T/P data), but the difference with its 5-day version was not noticeable. Space- and time-dependant EKETP and EKEmodel fields were eventually averaged horizontally over the same areas using the same integration tools. The horizontal structure and temporal fluctuations of these running variances are discussed below.

4. Low-frequency variability in the EKE fields

In this section we address the first question raised in the introduction by comparing the spatial and temporal structures of the observed and simulated EKE fields.

** Observed and modeled mean EKE fields (Fig. 3)*

The overall distribution of EKE from the HF simulation is in reasonable agreement with observed data (Fig. 3) in the North Atlantic. On average over the basin, the model simulates more than 80% of the observed EKE level. The most energetic regions are found within the North Equatorial Counter Current retroflection region (quite realistic model EKE field), within the Caribbean Sea (overestimated EKE), within the Azores Current (local maximum of EKE well located but underestimated), within the Gulf Stream (GS) and North Atlantic Current (NAC). Along this latter current system, model EKE reaches a localized maximum just downstream of Cape Hatteras, instead of the elongated EKE band observed between 70°W and 50°W. As in most geopotential-coordinate model simulations (even at the present resolution), this EKE maximum is associated with an unrealistic pulsating standing eddy located at the Gulf Stream separation point, and with an unrealistic overshoot of the Gulf Stream. The NAC path and associated EKE field also exhibits some discrepancies in the central basin. Instead of flowing north and creating an EKE maximum between 50°W and 40°W as observed, the simulated NAC and associated EKE maximum extend northeastwards across the Mid-Atlantic Ridge, leading to an overestimated EKE in the eastern basin. However, as argued later, this mean bias appears to not adversely affect the EKE variability studied here.

** Meridional redistribution of EKE between 1993 and 1996 (Fig. 4)*

The upper panel in Figure 4 shows the relative change between years 1993 and 1996 in annually-averaged near-surface EKE computed from T/P data. As expected, it does not differ much from SW98's Figure 1-c due to our similar data processing techniques. EKE has substantially decreased (by about 20-30%) north of about 45°N and increased over the subtropical gyre over this period at large scale. SW98 have shown that this change is statistically significant. Figure 4-b shows the same quantity computed from the HF simulation outputs. Local model-data accordance is good north of 45°N, except in the Irminger Current. Local model-data differences are more numerous in the subtropical gyre, especially along the eastern boundary and in the band [65-40°W, 20-28°N] where the model simulates a seemingly unrealistic increase of EKE between 1993 and 1996. Despite these local differences, the model reproduces fairly well the large-scale dipolar EKE pattern mentioned by SW98.

** Temporal evolution of subpolar and subtropical EKE (Figs. 5-a and 5-b)*

The model simulated and T/P observed running EKEs averaged within the two boxes shown in Figure 4-a are presented in Figure 5. This comparison comments on the real temporal evolution of the signal, the model's skill in capturing this signal and connections between these series and the NAO. In the following discussion, the meridional contrast of a variable refers to the difference between these subpolar and subtropical areal averages.

In the northern box (Fig. 5-a), the observed EKE fluctuates around 100 cm^2s^{-2}. It decreases during 1993, increases until mid-1995, decreases until early 1998 and finally increases until late 1999. With the exception of the 1995-1996 period when the model simulates a non-observed EKE decrease, the magnitude of subpolar EKE and its overall temporal evolution (especially the marked EKE minimum observed between 1996 and 1998) is realistically simulated over this 7-year period.

In the southern box (Fig. 5-b), T/P data exhibit an increase of EKE between 1994 and 1996 and a decrease between 1998 and 2000, modulated by 2 or 3 oscillations with a timescale of about 2 years. The simulated EKE level in this box is weaker than the measured level by about 30%. Simulated EKE increases in 1995, i.e. one year after observed, and then exhibits the same kind of 2-year modulation, approximately in phase with the observations. In accordance with T/P data, the model produces the highest EKE values in the subtropical area between 1995 and 1998.

Observations suggest that the mean level of EKE is stronger in the subtropical box than in the subpolar box, whereas both levels are of the same order in the simulation. Figure 3 shows that this comes from the overestimated level of simulated EKE in the eastern part of the North Atlantic Current (northern box), and, probably to a lesser extent, from the underestimation of EKE in the Azores Current (southern box). Despite these differences in mean EKE, the interannual variability of EKE turns out to be rather well simulated in both regions with observations during the T/P mission.

** Temporal evolution of the EKE meridional contrast (Fig. 5-c)*

Observations show that a negative meridional EKE contrast anomaly slowly builds up between 1993 and 1995, reaches -40 cm^2s^{-2} with respect to its initial value, and vanishes between 1998 and 1999 (thick line in Fig. 5-c). Superimposed over this 7-year evolution, a modulation at shorter period (1.5 to 2 year) is clearly visible between 1993 and the end of 1997..
This panel exhibits a remarkable model/data agreement in terms of meridional EKE contrast, phase and magnitude, both for the slow and shorter-term evolutions. The linear trends visible in both subpolar and subtropical simulated EKEs before 1994 cancel. These trends are also present in the PF simulation and thus correspond to an intrinsic long-term adjustment of the model energy. Two model-data misfits appear locally in Figure 5-c, though: the aforementioned absence in the model of the observed increase of subtropical EKE explains the model-data difference in 1995, and the unrealistic drop in the simulated subpolar EKE explains it in 1996. However, these differences can be considered as slight distortions within a globally consistent picture.

Interestingly, the 1993-1996 decrease in EKE contrast is well sampled from the dataset that was available to SW98: the difference between the 1996 and 1993 EKE fields nicely exhibits this strong signal. This change is also the strongest that occurred over the integration period, since the northern and southern EKEs were out of phase. A similar event occurred in the model in 1982-1985 and led to another noticeable fluctuation of the EKE meridional contrast. However, the realism of this latter event cannot be confirmed from observations because our altimeter dataset starts in 1993.

5. Link between the NAO and the meridional contrast of EKE

The NAO represents the leading mode of atmospheric variability over the basin on interannual to decadal timescales. This mode modulates the sea-level pressure (SLP) difference between Iceland and the Azores from which the NAO index is derived (Fig. 7). NAO-related atmospheric fluctuations directly affect the meridional gradient of SLP, and thus the location and intensity of the westerlies and its associated storm track (Hurrel, 1995; Rogers 1990). SW98 notice that both the westerlies and the storm track have shifted southwards between 1993 and 1996 in the North Atlantic (their Figure 14), in accordance with the NAO index evolution. They suggest the existence of a link between the NAO cycle and the EKE contrast observed via the energy input by the wind into the ocean. We showed in the previous section that the CLIPPER HF simulation reproduces well this observed large-scale EKE dipole, and its observed temporal variability during the whole T/P mission. This allows us to investigate SW98's hypothesis over the model 20-year integration, and to address the latter two questions raised in the introduction.

** General comments*

In the HF simulation, the meridional contrast of EKE seems to follow the evolution of the NAO index with a few months lag over the periods 1982-1984 and 1994-1999 (Fig. 6-b). The plain lines in Figure 6-c show the cross-correlations *C(lag)* between the unfiltered time series of NAO index and EKE contrast (shown in Fig. 6-a). In order to distinguish simply among the aforementioned periods of the HF simulation, *C(lag)* is shown for three successive time intervals of equal durations. The first and last thirds include the 1982-1984 and 1994-1999 events, respectively. The plain black and green lines in this figure show that in the HF run, *C(lag)* is maximum around 12 months before 1987 and around 9-10 months after 1994. The significance level of *C(lag)* may be estimated from the dashed lines shown in Figure 6-c. These lines show *C(lag)* for the three thirds of the PF experiment, in which no NAO cycle is forced and thus no significant cross-correlation is expected. Only the plain green line in Figure 6-c reaches +0.4 and exceeds the "noise" level reached by the dashed lines (after low-pass filtering, this cross-correlation is much stronger). The EKE contrast is thus significantly correlated with the NAO index with a 9-10 months lag after 1994. It is thus possible, if not proven, that the evolution of the EKE contrast was forced by the NAO cycle over the period 1994-1999 (which includes the time interval studied by SW98). In contrast, the EKE contrast anomaly observed in 1982-1985 may not have been forced by the NAO cycle, despite the resemblance and phase relationship between the corresponding blue and red peaks in Figure 6-b.

The existence of this NAO-EKE connection is supported by an additional feature: The 1993-1999 EKE contrast anomalies deduced from T/P observations (thick black line in Figure 5-d) and simulated in the HF run (thin black line in Figure 5-d) are clearly stronger than the standard deviation (σ_{PF} = 7.4 $cm^2 s^{-2}$) of the intrinsic variability generated by the model forced by a perpetual seasonal cycle (blue line in Figure 5-d). In 1996, these EKE contrast anomalies exceed 3 σ_{PF} and 4.5 σ_{PF} (from T/P and HF data, respectively).

The meridional contrast of the energy input by the wind is defined as the difference of $[\underline{u}.\underline{\tau}]$ averaged over the subpolar and the subtropical gyres. This term exhibits a strong monthly variability and seems to follow the NAO index with no lag (Fig. 6-a). Indeed, there is a clear zero-lagged correlation between these monthly time series over the 20 years of integration, slightly stronger after 1994 (Fig. 6-d).

SW98's hypothesis is thus very well supported by our results. The NAO index fluctuations modulate [i] the meridional contrast of energy transferred by the wind to the ocean with no significant lag, and, in turn, [ii] the meridional contrast of oceanic EKE with a 9-10 months lag after 1994. Some of these features are now investigated in more detail.

** Apparent absence of the link between NAO index and EKE contrast before 1994*

This NAO-EKE link is not always clearly present in both the observations and the model results. There are two possible explanations. First, it is possible that NAO-related, interannual atmospheric fluctuations may need to exceed a threshold in order to trigger the observed and simulated events, or make the response dominant. This is consistent with the fact that periods of possible NAO driving (i.e. 1994-1999, and perhaps 1982-1984) coincide with two major fluctuations of the NAO index. Second, and alternatively, the NAO-EKE link may actually exist over the whole integration period in the real world, but the forcing used in the HF simulation might be inaccurate in its NAO structure. Of these two, we prefer the first explanation, because the change in forcing (January 1st, 1994) did not affect the correlation found throughout the HF integration between the observed NAO index and the model wind forcing at important NAO locations (as shown in Fig. 7)

To summarize at this point, it appears that NAO-related atmospheric changes induce a quasi-immediate meridional redistribution of the energy provided by the wind to the ocean. Only strong NAO transitions are likely to trigger (or make emerge from other signals) observed meridional redistributions of surface EKE with a 9-10-months lag.

** Role of the westerlies and of the GS/NAC system in the fluctuations of the EKE contrast*

Figure 4 and SW98's Figure 1-c highlight the gyre-scale extension of the relative change in surface EKE between 1993 and 1996. Absolute changes in EKE are, however, confined along the GS/NAC system. This shows that the modulations of the EKE meridional contrast mostly reflect the difference in eddy activity along the southwestern and northeastern parts of this current system. The EKE contrast is therefore likely to be controlled by the GS/NAC dynamics, itself responding to the NAO cycle. This view is supported by two other facts.

High-frequency wind fluctuations such as those present in the storm track may contribute to force part of the oceanic EKE (Müller and Frankignoul, 1981), although baroclinic instability is known to dominate throughout most of the ocean (Stammer, 1998) and in particular along the main currents. Fluctuations of the NAO index correspond to simultaneous meridional migrations of the westerlies and of the associated storm track. Both latter features may contribute (through distinct mechanisms) to the EKE contrast evolution. The red line in Figure 5-d shows that between 1993 and 1999, the LF forcing, devoid of high-frequency atmospheric fluctuations, generates an evolution of the EKE contrast comparable to those derived from T/P and from the HF simulation. The non-synoptic, slow evolution of the atmosphere is thus the main forcing of the observed fluctuations of the EKE contrast. These EKE fluctuations probably follow the adjustment of the GS/NAC system to the NAO cycle through modifications of baroclinic instability (the main EKE source in this strong current). Nevertheless, the model-data agreement in terms of EKE contrast is better at timescales shorter than about 1 year in the HF simulation than in the LF simulation (Fig. 5-d), showing that high-frequency wind fluctuations contribute to the observed EKE contrast.

The magnitude of the lag (9-10 months) found between the NAO index and EKE contrast also supports the proposed scenario. If the direct, local forcing of oceanic eddies by atmospheric eddies in the storm track was dominant, NAO index fluctuations would induce oceanic EKE responses with essentially zero lag. The most plausible scenario is thus the following: the intergyre limit (the GS/NAC system) adjusts to the slow displacement of the westerlies, this affects the distribution of baroclinic instability and thus of EKE along the current, and ultimately the EKE meridional contrast.

** On the connection between the NAO cycle and EKE contrast*

The 9-10 months lag involves at least two timescales. The former is required for the ocean to adjust to changes in the wind forcing, and the second is likely to match the growth rate of mesoscale eddies. This latter timescale should not exceed 3 weeks, according to the linear baroclinic instability analyses done by Beckmann (1988) and Beckmann et al (1994). Therefore, the oceanic adjustment to changes in the wind forcing probably involves timescales of several months. This timescale is intermediate between the fast barotropic and slow baroclinic linear adjustment timescales of the basin, respectively on the order of days and years. This suggests that more complex processes are at work.

Unfortunately, only few studies are focused on the timescales and mechanisms involved in the large-scale horizontal circulation adjusting the NAO-related interannual wind changes. White and Heywood (1995) have noticed that, in accordance with the Sverdrup balance, the EKE veins associated with the branches of the NAC migrate in accordance with the line of zero wind stress curl on timescales shorter than one year. This simple Sverdrupian argument apparently explains the sign relationship we found between the NAO index and the EKE meridional contrast. Also based on simple Sverdrupian arguments, Bersch et al (1999) suggest, on the contrary, that the weak westerlies observed during negative NAO situation induce a shrinking of the subpolar gyre, i.e. a northwestward drift of the NAC path. The timescale of this oceanic response is not particularly discussed in this latter study but is again shorter than one year, as seen from the rapid evolution of the eastern North Atlantic stratification along a Greenland-Ireland section. This observed, rapid poleward shift of the NAC in low NAO situation is confirmed by numerical simulations (Eden and Willebrand, 2000) but is not consistent with White and Heywood (1995) observations and explanations. The Sverdrup argument thus leads to opposite expectations for the NAC path, and might thus not be relevant to interpret our results. In addition, this linear theory does not explain a NAO-EKE lag as long as 10 months.

The dynamics of the real ocean (and that simulated by the present model) are much more nonlinear and turbulent than simulated by Eden and Willebrand (2000). From idealized, eddy-resolving simulations, Dewar (2003) describes nonlinear modes involved in the adjustment of a turbulent ocean to NAO-like wind fluctuations. The most rapid mode corresponds to a barotropic-like adjustment process of the separated jet that starts just after the wind perturbation is applied. This nonlinear advective mechanism then needs a few months interval to complete the adjustment of the jet through a redistribution of the wind-forced potential vorticity (PV) anomaly. If the velocity is on the order of 0.2 m/s within O(1000 km)-wide western boundary recirculation cells, the jet should need about 2 months to adjust to the modified PV field. This time estimate is done in the context of an idealized western boundary current regime. In contrast, the ocean dynamics in the real world and in our simulations are influenced by many other factors. However, the actual, realistic geometry might not influence the estimated timescale for two reasons. Firstly, recirculation cells do exist in the real ocean and, secondly, the suggested adjustment mechanism is regional (not wavelike) thus its timescale does not depend on the domain extension. The 10-months lag we found is somewhat longer than the 2 months estimated above, but we might need a few of these adjustment cycles for changes in the general circulation (and subsequent EKE) to develop downstream of the western boundary.

6. Summary

Seven years of T/P data have been processed in the North Atlantic to investigate the interannual variability of EKE, in particular the behavior of a dipolar pattern previously mentioned by SW98. As showed by these authors, the subpolar (subtropical) EKE field was significantly stronger (weaker) than usual in 1993, and this contrast had changed signs in 1996.

We computed a simple EKE contrast index every ten days from T/P data. This showed that the EKE maximum associated with the NAC slowly shifted southward between 1993 and 1996, and this trend reversed after 1997 until the end of 1999. In other words, the 1993 and 1996 situations described by SW98 were part of a more general, interannual fluctuation of the North Atlantic eddy field. The same index was computed from the outputs of the 1/6° CLIPPER numerical simulation: the model realistically reproduces the evolution of the meridional EKE contrast during the T/P mission.

Model integrations were diagnosed to investigate the possible connection between this EKE evolution and the NAO cycle (as suggested by SW98). The observed and simulated EKE contrast indexes were shown to follow the evolution of the NAO index with a 9-10-months lag over the T/P period. This link is not visible before 1994 in the simulation, suggesting that only strong NAO index changes (such as those observed after 1993) can trigger such EKE meridional redistributions. Using two other model simulations, we showed that the 1993-1999 slow EKE evolution (i) was significant, i.e. largely exceeding the intrinsic interannual variability of EKE obtained with a repeated seasonal cycle, and (ii) forced by the slow migration of the westerlies associated with the NAO cycle, i.e. not the atmospheric high-frequency fluctuations.

The value of the lag itself suggests that the NAO-related atmospheric fluctuations first induce an adjustment of the NAC system, which then affects the meridional contrast of EKE. The timescale (9-10 months) of the adjustment is slower (faster) than that of a linear barotropic (baroclinic) adjustment process. Linear arguments (such as those developed by Eden and Willebrand, 2000) do not provide adequate explanations for the 9-10 months lag. In contrast, the nonlinear adjustment process studied by Dewar (2003) might support it. In this scenario, NAO-related, wind-forced potential vorticity anomalies are advected in the inertial recirculations and affect the large-scale circulation within a few months.

Additional idealized investigations are necessary to understand these dynamical links in more detail. This work is currently underway.

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TABLE

Table 1: Model integrations

Integration	Forcing	Initial State	Integration period
Spinup	Climatological ERA15 ECMWF climatology	Reynaud et al (1998) climatology, state of rest	8 years
HF (high frequency)	Daily ECMWF re-analysis (1979-1993) and analysis (1994-2000)	End of spinup (February 14, 1979)	1979-2000
LF (low frequency)	Low-passed HF forcing (cutoff period: 1 month)	HF state on March 21, 1992	March 21, 1992 until December 31, 1999
PF (periodic forcing)	Same as spinup	End of spinup (1979)	Spinup forcing repeated until December 31, 1999

FIGURES

Figure 1: Model bathymetry (m) and domain. White lines locate the four open boundaries.

Figure 2: Zonal component (Pa) of the wind stress at 38.67°W, 16.68°N used to force the HF (rough line) and LF (smooth line) runs during year 1994.

Figure 3: Multi-year averages of yearly eddy kinetic energies (EKE) in cm^2s^{-2} (a) from TOPEX/Poseidon data on the period 1993-2000 and (b) from the HF simulation at 55 m on the period 1980-2000. These EKEs are computed over individual years (running EKE on July, 1st) and averaged over these years. The color scale is the same.

Figure 4: Difference of EKE between 1996 and 1993 normalized by the averaged EKEs between 1993 and 1996: (EKE1996-EKE1993) /[(EKE1993 + EKE1994 + EKE1995 + EKE1996) / 4]. The resulting quantity was smoothed to highlight large-scale structures. (a) TOPEX/Poseidon data, (b) HF simulation outputs. Whites rectangles represent the northern and southern boxes over which EKE is computed and displayed in Figure 5.

Figure 5: Temporal evolution of T/P and HF simulation EKE in cm^2s^{-2}, horizontally averaged (a) in the southern box and (b) in the northern box shown in Figure 4. (c) Anomaly of the EKE meridional contrast (EKEnorth - EKEsouth, in cm^2s^{-2}) from T/P and from the HF model simulation. (d) Same as (c) with EKE contrasts diagnosed from simulations LF and PF. Differences in panels (c) and (d) are presented in anomaly with respect to their mean over years 1993-1999.

Figure 6:

(a) Normalized anomalies (with respect to the whole time series) of the following quantities in the HF simulation: monthly NAO index (red, Hurrel 1995), meridional contrast of EKE (blue) and of the energy input by the wind, i.e. $[\underline{u}.\underline{\tau}]_{North}$ - $[\underline{u}.\underline{\tau}]_{South}$ where $\underline{u}$ and $\underline{\tau}$ are the monthly surface model velocity and wind-stress, respectively. The 1994-1999 period from which panels (c), (d) and (e) were built is highlighted by thick lines. The period 1982-1985 is also highlighted.

(b) Blue and red lines from (a) after low-pass filtering.

(c) Cross-correlations *C(t)* between the monthly NAO index and the EKE contrast over the 3 thirds of the period 1980-1999 in the HF simulation. NAO index fluctuations lead at positive lags. Plain and dashed lines correspond to HF and PF runs, respectively.

(d) Same as (c) but cross-correlation between the NAO index and the meridional contrast of the energy input by the wind during the three thirds of the HF integration.

(e) Cross-correlation between the NAO index and the EKE contrast (blue line) and between the NAO index and the meridional contrast of the energy input by the wind in the HF run over the period 1994-1999 (black, corresponds to the latest thick lines in panels a and b).

Figure 7: Evolution of the normalized NAO index (red) and the normalized anomaly of the zonal wind stress at 55.7°N, 25°W throughout the HF integration (black). The NAO index is taken from Hurrel (1995).

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