Use of breeding to detect and explain instabilities in the global ocean

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[1] The breeding method of Toth and Kalnay finds the perturbations that grow naturally in a dynamical system like the atmosphere or the ocean. Here breeding is applied to a global ocean model forced by reanalysis winds in order to identify instabilities of weekly and monthly timescales. This study extends the method to show how the energy equations for the bred vectors can be derived with only very minimal approximations and used to assess the physical mechanisms that give rise to the instabilities. Tropical Instability Waves in the tropical Pacific are diagnosed, confirming the existence of bands of both baroclinic and barotropic energy conversions indicated earlier by Masina et al. and others. In the South Atlantic Convergence Zone, the bred vector energy analysis shows that there is kinetic to potential ocean eddy energy conversion, suggesting that the growing instabilities found in this area are forced by the wind.


1. Introduction

[2] Previous examinations of the structure and causes of flow instabilities in the ocean have generally required consideration of time averages of the kinetic and potential energy equations [e.g., Pinardi and Robinson, 1986; Ducet et al., 2000] or even a full linear instability analysis [Huck and Vallis, 2001]. However, the process of time averaging reduces the ability of this approach to discriminate among several concurrent instabilities with differing time-evolutions. Here we explore the potential of the breeding method to isolate and identify the aspects of time-dependent ocean flows that are unstable to small perturbations. The method was originally developed both for application to data assimilation (to identify the growing component of the analysis error [Yang et al., 2008]) and to provide a set of plausible initial conditions for ensemble forecasting of atmospheric motions representative of the growing errors in the analysis [Toth and Kalnay, 1993, 1997]. Here we extend the method to provide an alternative method of computing the energetics of the global upper ocean which also has advantages of being simple to implement and computationally inexpensive.

[3] The breeding method begins with an arbitrary small perturbation of the initial state of an unstable system, such as the ocean, represented by a numerical model. This model is integrated forward for a time interval, ∆t, beginning from both the perturbed and unperturbed (or control) initial state. The vector difference in ocean state variables between the two resulting nonlinear forecasts is called the bred vector. At ∆t this bred vector is rescaled to the size of the initial perturbation and then is added to the control simulation to form the perturbed initial state for a new simulation. Examples of norms for rescaling are the root mean square difference of sea surface temperature (SST) or the kinetic energy of the perturbations. Twin simulations beginning with the control and newly perturbed control initial state at ∆t are then integrated forward from ∆t to 2∆t to create a new simulation pair. The bred vector at time 2∆t is then computed, rescaled, and the process is repeated. After a short spinup and when carried out over many cycles, the resulting time series of bred vectors has been shown by Toth and Kalnay to isolate and identify the components of the system that grow most rapidly on a time-scale of ∆t or longer, and to separate them from other rapidly growing components that saturate in times shorter than ∆t. By varying ∆t (hereafter known as the “breeding interval”), Peña and Kalnay [2004] showed how to isolate instabilities of different temporal scales. The bred vectors created by this process are essentially non-linear generalizations of Lyapunov vectors and, like Lyapunov vectors, they are independent of the norm chosen for rescaling [Toth and Kalnay, 1997; Kalnay, 2004]. Equivalent results can be obtained with any norm; however, given a norm, rescaling the BVs to different sizes as measured by that norm controls the degree of nonlinearity in the BV evolution. In addition, the time interval can be chosen to select phenomena which have not reached error saturation in that interval. It is through tuning these two parameters that the breeding method can be used to isolate different types of instabilities [Peña and Kalnay, 2004; Chikamoto et al., 2007; Vikhlaev et al., 2007].

[4] Like the atmosphere, upper ocean currents are subject to a variety of flow instabilities. These instabilities are concentrated in regions of strong currents such as the western sides of subtropical gyres and the deep tropics where eddy kinetic energy may exceed 4500 cm²s⁻² [Ducet et al., 2000]. Many currents, such as the Agulhas, Kuroshio, Gulf Stream, Brazil, Malvinas, and Antarctic Circumpolar Currents, have a fairly constant level of eddy variability year round. Others, such as the North Equatorial Counter Current (NECC) of the tropical Pacific, show strong seasonality. For the NECC instability generated eddy kinetic energy reaches a maximum at 10°N in summer [Ducet et al., 2000]. A little south of the NECC, in the latitude range 3⁰N–6⁰N, tropical instability waves (TIWs) occur in the longitude band between 180⁰ and 120⁰W longitude. These TIWs also have a well defined seasonal cycle, with activity beginning in August and continuing through March of the next year [Masina and Philander, 1999; Masina et al., 1999].
Beginning with Philander [1976] there has been a long running discussion in the literature regarding the relative importance of baroclinic, barotropic, and frontal instabilities in providing the energy source for these TIWs. On the equator, most of the wave energy has been observed just above the Equatorial Undercurrent (EUC) [Weisberg, 1984; Luther and Johnson, 1990; Qiao and Weisberg, 1998]. Using observations from moored current meters, Weisberg [1984] calculated that the barotropic conversion in the cyclonic shear region of the EUC in the Atlantic was enough to account for the growth of the TIWs there and argued for a similar mechanism in the Pacific. In another observation experiment, Luther and Johnson [1990] argued for three distinct sources of TIW energy: the barotropic conversion in the EUC at the equator and two regions of baroclinic conversion between $3^\circ$N and $6^\circ$W and between $5^\circ$N and $9^\circ$N. More recently, numerical model simulations have been used to diagnose the instabilities. Masina and Philander [1999] and Masina et al. [1999] suggest that there are two distinct locations of energy conversion with baroclinic conversion occurring between $3^\circ$N and $5^\circ$N and barotropic conversion occurring further equatorward. Perturbation energy budgets performed on a 2.5 layer model by McCready and Yu [1992] emphasized the importance of barotropic instability and introduced a frontal instability mechanism as an energy source for TIWs. Baroclinic instability, however, was found to be an energy sink. Using data from a 28-year long run of a coupled general circulation model, though, Yu and Liu [2003] found that baroclinic instability associated with the northern SST front was of major importance in generating Pacific TIWs around both $2^\circ$N and $2^\circ$S. The strength of the TIWs is closely tied to the phase of ENSO, with the diminished strength of SST front in El Niño years associated with a decrease in TIW production and the increased SST front of La Niña causing stronger TIW activity [Contreras, 2002].

To explore the potential of the breeding approach in examining fluid instabilities in the ocean we extend the breeding method by defining the potential and kinetic energy equations for the perturbations and use these to explore instabilities in an ocean general circulation model driven by observed winds. In our discussion, we compare the results of this new approach with more traditional methods.

### 2. Model and Methods

The primitive equation Geophysical Fluid Dynamics Laboratory (GFDL) Modular Ocean Model v.2 is used in a domain extending from $62.5^\circ$S–$62.5^\circ$N with $1^\circ \times 1^\circ$ horizontal resolution in midlatitudes reducing to $1^\circ \times 1/2^\circ$ at the equator in order to resolve the intense equatorial current systems. The model has 20 fixed depth levels in the vertical with 15m resolution near the surface expanding to 737m near the bottom. Horizontal and vertical mixing and heat and salt diffusion parameters are set as described by Carton et al. [2000a, 2000b] in order to reproduce the mean circulation. Initial conditions of climatological temperature and salinity are obtained from the World Ocean Atlas 1994 [Levitus and Boyer, 1994], while monthly winds are provided by the National Centers for Environmental Prediction (NCEP) reanalysis [Kalnay et al., 1996]. Surface heat and freshwater flux are calculated using a simple Haney-type relaxation to climatological monthly temperature and salinity.

To begin the breeding process, a random perturbation sampled from a flat distribution between $-0.5^\circ$C and $0.5^\circ$C is introduced into the initial conditions for the sea surface temperature (SST) field. The remaining experiments reported here all use this same initial perturbation. Other experiments using different randomly chosen initial perturbations or perturbations in the velocity field yield similar results, confirming the earlier observation by Toth and Kalnay [1993, 1997] that the structure of the bred vectors is independent of the initial perturbation.

The derivation of the bred vector conservation of kinetic and potential energy equations resembles that of the more common perturbation energy equations [Orlanski and Katzfey, 1991; Oczkowski et al., 2005] although, due to the fact that both the control and the perturbed nonlinear runs satisfy exactly the model equations, the bred vector equations are exact and do not require Reynolds averaging. A complete derivation of the bred vector energy equations can be found in the auxiliary material.  

1. Bred vector kinetic energy is defined as $KE_b = \rho_b \tilde{V}_b \cdot \tilde{V}_b / 2$ where $\tilde{V}_b$ is the bred vector horizontal velocity. Substituting this definition into the momentum equations, where $\tilde{V}_p$ is the control run horizontal velocity, leads to:

$$
\frac{\partial KE_b}{\partial t} = - \left[ \nabla \cdot \left( \tilde{V}_b \partial KE_b \right) \right] + \frac{\partial (w_c KE_b)}{\partial z}
- \nabla \cdot \left[ \tilde{V}_b \cdot \left( \rho_b \partial p_b \right) \right] - w_b \partial g \rho_b - \rho_0
+ \tilde{V}_b \cdot \left( \rho_b \partial \tilde{V}_b \right) + \rho_b \tilde{V}_b \cdot \tilde{F}_b
$$

where $w_b$ and $w_c$ are bred and control vertical velocities, $p_b$ is bred vector pressure, and $\rho_b$ is bred vector density. The first bracketed term is horizontal and vertical divergence of the kinetic energy transport, and vanishes when integrated over the whole domain. The second is the work of the pressure force, the third is the baroclinic energy conversion from perturbation potential to perturbation kinetic energy, the fourth is barotropic energy conversion from background kinetic to bred kinetic energy, and the fifth term is a friction term. The friction term and vertical transports are generally negligible in the problems considered here.

Similarly, bred vector potential energy, defined as $PE_b = \rho_b \tilde{g} \tilde{V}_b^2 / 2 \rho_0 N^2$, is governed by the following equation:

$$
\frac{\partial PE_b}{\partial t} = - \left[ \nabla \cdot \left( \tilde{V}_b \partial \tilde{V}_b \right) \right] + \frac{\partial (w_c PE_b)}{\partial z}
+ w_b \partial g \rho_b - \rho_b \tilde{g} \tilde{V}_b^2 / 2 \rho_0 N^2
+ \tilde{F}_b \cdot \left( \rho_b \partial \tilde{V}_b \right)
- \tilde{V}_b \cdot \tilde{F}_b
$$

The first bracketed term is horizontal and vertical divergence of the potential energy transport, and vanishes when integrated over the whole domain. The second term is baroclinic energy conversion and has the opposite sign of the corresponding term in the kinetic energy equation. The third term is negligible, since it is proportional to the perturbations.  

1Auxiliary materials are available in the HTML. doi:10.1029/2009GL037729.
in density times a term that vanishes when integrated over the whole domain. The last term is also negligible. In this Letter we focus mainly on interpreting the bred vector kinetic energy equation.

Results from control and bred vector simulations spanning two periods are examined, a multi-decadal period beginning January 1951 through December 1979 and a shorter, observation-rich period spanning the eight year period January 1985 through December 1992. Due to space limitations we will focus on the latter period, although the longer run is used to compute climatological monthly averages. A monthly breeding interval is used for the shorter simulation, while a 10 day breeding interval is used for the longer simulation to better isolate the period of TIWs.

3. Results

We begin by considering the bred vector energy balance on 11 November 1988, a time when the tropical Pacific was in a late developing La Niña (with a Southern Oscillation Index of 21.0 and a Nino3.4 Index value of $-2^\circ$C). The bred vector shows a dipole pattern off the coast of South America and a wave pattern in the Tropical Pacific (Figure 1a) which successive bred vectors show to have a period of $24^\circ$ days and to propagate westward at 0.46 m/s. Examination of the bred vector energetics shows that baroclinic processes are causing an increase in bred vector kinetic energy along the equator (Figure 1b). In the region of the dipole pattern off the coast of South America, by contrast, there is a conversion from bred vector kinetic to potential energy consistent with a transfer of bred vector kinetic energy from the atmosphere to the ocean, which is then converted into bred vector potential energy.

Tropical waves in this simulation first appear in August, are seen to strengthen through the winter, and then dissipate by May of the following year. Bred vector barotropic energy conversion exhibits the same seasonal cycle, with conversion increasing in August between 150°W and 5°N latitude, and between the surface and 150 m depth. The vertical axis has units $10^{-7}$ kg m$^{-1}$ s$^{-3}$.
120°W and shifting westward and extending to approximately 200°W by the end of the year (Figure 1b). Bred vector baroclinic energy conversion is maximum along the northern edge of the Pacific equatorial cold tongue, between 3°N and 5°N (Figure 2a). Positive barotropic conversion is seen in two latitude bands. The maximum of barotropic energy conversion occurs just north of the Equator, while bred vector barotropic conversion can also be seen in the same 3°N to 5°N latitude band as baroclinic conversion (Figure 2a).

[14] We next consider the interannual dependence of the energy conversion terms (Figure 2b). Both baroclinic and barotropic energy conversion terms spike during August through January, with the size of the spike varying by year. The strongest spike in energy conversion occurs in the La Niña period of 1988–1989 when the TIWs, NECC, and Equatorial Undercurrent are all anomalously strong. During this spike in energy conversion, baroclinic energy conversion is positive, indicating a conversion from bred potential to bred kinetic energy, while the barotropic conversion is negative, indicating a transfer from bred kinetic energy to background kinetic energy. In contrast to the La Niña period, bred vector energy conversion is weak during the 1991–1992 El Niño when the Equatorial Undercurrent has reduced transport and TIWs are weak.

[15] Finally, we examine the vertical structure of the bred vector energy conversion. The majority of the baroclinic conversion occurs above the thermocline, with the strongest conversion taking place in the upper 100 meters. This can be seen in the October average, which is qualitatively representative of the location and pattern of the relative amplitude of the conversion during fall and winter months (Figure 3a). The longitude of this maximum baroclinic conversion corresponds to the location of the tongue of cool SSTs (which can be seen at the surface of the middle of Figure 3a) and consequent strong meridional SST gradient. Barotropic conversion also occurs in this region and the equator between 160°W and 125°W (Figure 3b) although it takes place deeper than baroclinic conversion, with the strongest conversion at and just below the shear zone between the westward South Equatorial Current and the eastward Equatorial Undercurrent.

4. Summary

[16] The purpose of this Letter is to apply bred vectors, an idea developed in the context of atmospheric data assimilation, to stability analysis of ocean circulation. As part of this application we introduce the bred vector energy equations, which are analogous to the more traditional eddy energy equations but are obtained without averaging or approximations (other than the neglect of terms shown to be small by scale analysis) due to the fact that both the control and the perturbed runs satisfy the model dynamical equations. We find, consistent with findings reported by Yang et al. [2006], that changes in bred vector energy reflect important aspects of the growth of flow instabilities. Thus, breeding, the process by which the bred vectors are constructed, is able to identify ocean instabilities effectively and inexpensively. Because they span the state space described by key ocean processes, bred vectors also have potential applications in the construction of ensembles of model states for ensemble data assimilation and forecasting.

[17] Our examination of bred vectors in the global ocean focuses on instabilities of tropical Pacific currents because of their intensity, their importance for coupled air-sea interactions, and because of the extensive literature describing them. Examination of the bred vector energy equations shows that there are two locations of energy conversion for the tropical instability waves which dominate intraseasonal variability in this region. Between 3°N and 5°N, both baroclinic and barotropic energy conversion occurs along the northern edge of the cool tongue. A separate region of barotropic conversion is detected just north of the equator in the shear zone between the Equatorial Undercurrent and the shallower South Equatorial Current, e.g., in agreement with Massina et al. [1999]. Both types of energy conversion have interannual variations due to changes in the currents and stratification, which are themselves closely tied to the phase of ENSO.
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