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On the role of ocean-atmosphere interaction in midlatitude interdecadal variability

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Abstract. A simple midlatitude ocean-atmosphere model is used to investigate possible roles of coupled feedback in interdecadal climate variability over the North Atlantic and North Pacific oceans. Stochastic forcing by atmospheric internal variability maintains variance in both uncoupled and coupled cases. Coupling contributes to creating an interdecadal mode with distinct spatial pattern and preferred time scale, seen as a broad spectral peak. A near-analytic solution for the coupled interdecadal mode suggests that the most important parameter in determining the period is the zonal length scale of the atmospheric wind stress feedback over the region. Subject to this scale, the period is then determined by oceanic Rossby wave dynamics, which tends to give westward propagation in subsurface fields. The dipolar sea surface temperature (SST) anomalies are generated primarily by the advection of climatological SST by geostrophic current. Although the magnitude of the feedback of SST to atmospheric response is much smaller than atmospheric internal variability, its effects are significant.

1. Introduction

A number of studies have documented climate variability on time scales of decades over the North Atlantic and North Pacific (see Trenberth, 1990; Wallace et al., 1990; Deser and Blackmon, 1993; and references therein). The challenge is to distinguish among various hypothesized mechanisms. The "null hypothesis" is atmospheric white noise stochastic forcing being converted to a red spectrum by upper ocean thermal inertia (Hasselmann, 1976; Frankignoul and Hasselmann, 1977; Battisti et al., 1995). Dynamically more complex sources of variability include internal ocean variability or ocean dynamical response to atmospheric variability (e.g., Weaver et al., 1991; Delworth et al., 1993; Speich et al., 1995; Saravanan and McWilliams, 1997 and references therein). The question has been raised whether active atmospheric coupling plays a role, with effects of SST on the midlatitude atmospheric circulation feeding back onto the ocean circulation, at least in some coupled general circulation models (GCMs) (Latif and Barnett, 1994; von Storch, 1994; Robertson, 1996) and perhaps in observations (Kushnir, 1994). Atmospheric GCM experiments suggest that the midlatitude response to SST is modest compared to atmospheric internal variability, may have complex seasonal and nonlinear dependences, and varies among models (Palmer and Sun, 1985; Kushnir and Lau, 1992; Ferranti et al., 1994; Peng et al., 1995). Given this, it is an open question whether such coupled feedbacks play any significant role in interdecadal variability and, if so, what that role is. Simple models can assist in defining what this role might be (Liu, 1993; Jin, 1997; M. Münnich et al., 1997, pers. comm.; N. Schneider, 1997, pers. comm.).

Here, we investigate possible physical mechanisms of interdecadal variability in a simple ocean-atmosphere model for idealized North Atlantic and North Pacific basins, including near-analytical solutions, which provide prototypes for discussing how midlatitude ocean-atmosphere feedbacks may behave.

2. The coupled model

We consider a linearized perturbation system with quasi-geostrophic shallow water upper ocean dynamics and an SST equation for a surface mixed-layer. The linearized vorticity and SST equations on a β-plane are

$$[\partial_t(\nabla^2 - \lambda^{-2}) + \beta\partial_x + \epsilon_c\nabla^2 - \nu\nabla^4]\psi_g = -\frac{\partial_y \tau_x}{\rho H} \quad (1)$$

$$\partial_t T = (\partial_x \bar{T}\partial_y - \partial_y \bar{T}\partial_x)\psi_g + \frac{\tau_x}{f\rho h}\partial_y \bar{T} + \frac{Q}{c_w \rho h} \quad (2)$$

where $\nabla^2 = \partial_x^2 + \partial_y^2$; β is the latitudinal derivative of the Coriolis parameter (fixed at the 40°N value); ψ_g is the perturbation geostrophic streamfunction; T the perturbation SST; $\bar{T}$ the climatological SST; λ the Rossby deformation radius, here 32 km; f the Coriolis parameter; ϵ_c a Rayleigh damping rate of ocean current, varied between 0 and 1 year^{-1}; ν the horizontal turbulent viscosity coefficient, between 10 and 10^4 m^2 s^{-1}; ρ the sea water density; H the depth of the upper layer, about 100 m; h the depth of surface mixed-layer, here 50 m; c_w the specific heat of sea water; τ_x is the zonal component of wind stress at the sea surface; Q the heat flux at sea surface. Meridional wind stress is neglected.

The atmospheric model represents atmospheric internal variability as stochastic wind stress (ξ_τ) and heat

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flux (ξ_q) forcing, and approximates the atmospheric response to the ocean through wind stress and heat flux feedbacks associated with an SST basis function, θ_1. The spatial wind stress and heat flux feedback patterns are estimated based on atmospheric GCM response to the SST pattern θ_1. In the case of wind stress, the GCM patterns are approximated by fitting the sinusoid form in (3). Seasonal dependence can be important (Peng et al., 1995) but is neglected for simplicity. We use

$$\tau_x = \mu\tau_A \cos(l_a y + \gamma)\cos(k_a x + \alpha)\frac{<\theta_1 T>}{<\theta_1^2>} + \xi_\tau \quad (3)$$

$$Q = -c_w \rho h \epsilon_T T + \mu Q_{fb}(x,y)\frac{<\theta_1 T>}{<\theta_1^2>} + \xi_q \quad (4)$$

where τ_A is the amplitude; k_a and l_a the wavenumbers; α and γ the phases; μ a coupling coefficient for sensitivity studies. The projection of SST on θ_1 over the basin is written $<\theta_1 T>$. Q_{fb} is a non-local heat flux feedback. The local dependence of heat flux on SST is treated separately in the first term of (4) with SST decay time scale ϵ_T, about 2.7 year^{-1}, estimated from Hamburg ECHAM-2 T21 GCM surface heat fluxes.

For the Atlantic case, $\alpha = 0.5\pi$, $\gamma = -0.8\pi$, $\tau_A = 0.05$ dyne/cm^2, $2\pi/k_a$ is about 10,000 km and $2\pi/l_a$ about 6,700 km. These parameters, and the fields θ_1 (Fig. 1a), and Q_{fb} (Fig. 1b) are estimated based on results from the Max-Planck-Institut für Meteorologie's (MPIM) ECHAM-2 T21 atmospheric GCM, in an analysis similar to Graham et al. (1994), Zorita et al. (1992), and Kharin (1995). In the Pacific, $2\pi/k_a$ is about 13,600 km. We estimate the spatial patterns of zonal wind stress feedback, θ_1, and Q_{fb} based on Latif and Barnett's (1994) Figs. 3c, 1b, and 3b, respectively, or based on Lau and Nath's (1990) Figs. 12c, 13a, and 13c, respectively.

Fourier transforming the coupled system (1) - (4) in time, as $e^{i\omega t}$, we solve two versions of the problem: a stochastically forced response; or an eigenvalue problem for complex frequency ω. For the eigenvalue problem, the solution for $\tilde{T}$ becomes

$$\tilde{T}(x,y) = \mu(i\omega + \epsilon_T)^{-1}\frac{<\theta_1 \tilde{T}>}{<\theta_1^2>}\left(\frac{V_P + V_H}{A^2 + \beta^2 k^2}\right.$$

$$\left. + \frac{\tau_A \cos(l_a y + \gamma)\cos(k_a x + \alpha)}{f\rho H}\partial_y \bar{T} + \frac{Q_{fb}}{c_w \rho h}\right) \quad (5)$$

where $A = i\omega(K^2 + \lambda^{-2}) + \epsilon_c K^2 + \nu K^4$, with $K^2 = k_a^2 + l_a^2$. The quantity $V_P(x,y,\omega) = l_a\tau_A(\rho H)^{-1}[l_a \cos(l_a y + \gamma)\partial_x \bar{T} - \sin(l_a y + \gamma)\partial_y \bar{T}\partial_x][\beta k_a \sin(k_a x + \alpha) - A\cos(k_a x + \alpha)]$ is associated with the advection by geostrophic currents from a particular solution directly due to the wind stress feedback, and $V_H(x,y,\omega)$ is a similar term from the homogeneous solution required to match basin boundary conditions. To obtain a frequency equation for ω, take the inner product of (5) with θ_1, and cancel the factor $<\theta_1 \tilde{T}>$ from both sides. After rearrangement of terms, the frequency equation is then solved numerically or by approximation method.

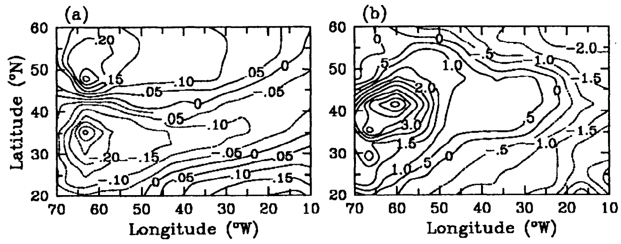

Figure 1. North Atlantic patterns based on MPIM atmospheric GCM results, stretched to a rectangular basin (a) The SST basis function, θ_1; (b) the heat flux feedback, Q_{fb}, in W m^{-2}; signed positive downward.

When temporally white atmospheric noise is included, the power spectral density of the response can be obtained by computing $<\theta_1 \tilde{T}>$ from the frequency equation for a set of given values of ω. Then Fourier-transformed geostrophic streamfunction and SST anomalies can be obtained.

3. Model results

We solve the eigenvalue problem for both coupled ($\mu = 1$) and uncoupled ($\mu = 0$) systems. Two types of eigenmodes are strongly affected by coupling: a rapidly decaying non-oscillating SST mode and a weakly decaying interdecadal mode. The coupled interdecadal mode is more interesting because of its time scale and its weak decay rate. This mode has a period of around 18 years in the North Atlantic and around 24 years in the North Pacific. The streamfunction of the interdecadal mode in the North Atlantic (Fig. 2a) exhibits large-scale, westward propagating oscillations. The SST anomaly is characterized by a large-scale north-south dipolar pattern (Fig. 2b). The basic features of the two fields found in Fig. 2 are also found in the North Pacific with longer zonal length scale. The interdecadal mode is not very sensitive to modifications of the nonlocal heat flux feedback; the period does not depend strongly on modest changes in phase parameters in the wind stress. Thus the Pacific case results with feedbacks approximating Latif and Barnett (1994) and Lau and Nath (1990) were qualitatively similar, aside from shifts in the dipole pattern of SST, and the longer period.

The interdecadal mode would have to be maintained by stochastic forcing from the midlatitude atmosphere. Major questions are whether the signature of this mode can be distinguished from other stochastically maintained variability, and whether the coupled case is significantly different than the uncoupled ocean response. When stochastic forcing, white in time, is added, the power spectral density of the streamfunction and SST anomaly (sum of squares over the basin) in the North Atlantic ocean are shown in Fig. 3 for two values of the coupling coefficient. For uncoupled or low coupling cases (Fig. 3a), there is no power spectral peak. This is because all the oceanic modes simply contribute to the background reddened response, akin to the Hasselmann (1976) hypothesis. As the coupling coefficient increases,

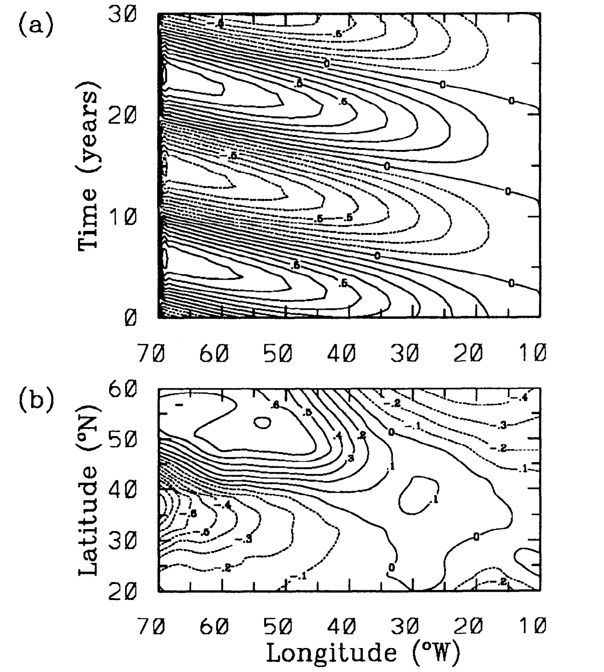

Figure 2. Patterns from the eigensolution for the interdecadal mode. (a) The time-longitude dependence of the geostrophic streamfunction; (b) the spatial form of the SST anomaly in the North Atlantic at the phase when feedback is maximum, about t=0 and t=18 in (a).

a spectral peak for each variable starts to rise. For the standard coupling value (Fig. 3b), this peak is modest but quite distinct (note linear scale). The exact value of the peak (17 years here) is sensitive to parameters and the peak values can be shifted by changes in the red background, but the peak is due to the interdecadal mode. The sharpness of the peak also exhibits sensitivity to the form of the stochastic forcing. For the case shown, the spatial form of the wind stress and heat flux forcing are the same as the wind stress and nonlocal heat flux feedbacks, respectively. This is chosen to demonstrate that even if the spatial form identical, the case with feedback behaves differently than the case with only noise forcing. In the North Pacific case, we noted much stronger dependence on the spatial patterns in the stochastic forcing. This more complex relationship between the eigenvalue and stochastically forced problems is being pursued in further work.

4. Physical mechanism

To understand the physical processes behind the interdecadal mode, we seek a near-analytic solution. Analyzing the contributions to the frequency equation term by term, we find that the frequency of the interdecadal mode is dominated by $\beta k_a/(k_a^2+l_a^2+\lambda^{-2})$, which looks like the oceanic Rossby wave frequency. It is not simply a Rossby wave because the wavenumber k_a is given by the atmosphere through the coupling. The interdecadal mode frequency is mainly determined by the Rossby wave frequency that most closely matches the typical wavelength of the atmospheric feedback. The westward propagation is associated with long Rossby-wave-like dynamics acting subject to constraints by the coupled feedback in wind stress. During its propagation, SST anomalies are generated by the advection of climatological SST by the geostrophic current. This SST anomaly produces a weak but persistent feedback to the atmosphere. Whereas an uncoupled long Rossby wave would tend to dissipate upon reaching the western boundary, in the coupled case the atmosphere carries some information back eastward. This tends to re-excite the wave in the interior of the basin, and creates a distinct oscillation time scale.

The decay rate depends on the choice of ocean damping parameters, but the interdecadal mode always decays more slowly than the uncoupled ocean modes because of its larger spatial scale. For free ocean wave solutions, matching at the western boundary to short Rossby waves gives a larger effect of diffusion.

5. Summary and discussion

Although the atmospheric response to midlatitude SST anomalies is complex, prototypes for how such feedbacks might behave in the coupled system can be useful. The principal postulate here is that in addition to atmospheric internal variability, represented as a stochastic process, there is an atmospheric response to SST. This is represented by nonlocal feedbacks associated with a specific SST pattern, and crudely estimated from GCM results, neglecting seasonal cycle effects. When coupled to a simple geostrophic ocean model with a fixed-depth mixed layer, this feedback produces an interdecadal oscillation that differs significantly from modes of the uncoupled ocean. An approximate analytic expression for the frequency suggests a view of how ocean-atmosphere interaction sets the time scale and spatial form of this mode. The wind stress feedback in the atmosphere tends to set the dominant zonal scale of the mode; subject to this scale, the period is then determined by oceanic Rossby wave dynamics. For an estimate of the wavelength over both oceans, the period is less than 20 years in the North Atlantic, and between 20 to 30 years in the North Pacific. However, the period is sensitive to the length scale of the

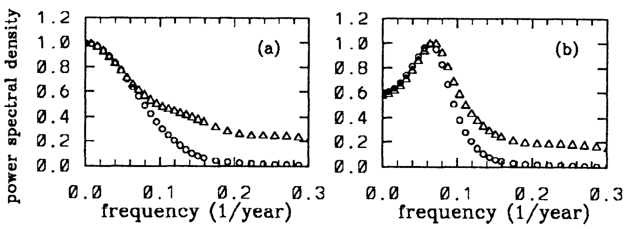

Figure 3. The power spectral density of geostrophic streamfunction (open circles) and SST anomaly (open triangles) in the North Atlantic for a stochastically forced case. (a) For low coupling, μ = 0.1; (b) for standard coupling, $\mu = 1$.

atmospheric wind stress feedback, tending to increase with this length scale. Our values are roughly fit to atmospheric GCM results, but we know that these patterns can be sensitive to the model used and estimation procedure. We thus emphasize the physical mechanism rather than the precise period. However, we note that Latif and Barnett (1994, 1996) and Robertson (1996) find time scales on the order of 20 years in the North Pacific, while A. Grötzner et al. (1997, pers. comm.) note an 18 year time scale in the North Atlantic in two MPIM coupled models, and F.M. Selten et al. (1997, pers. comm.) find 18 years in their coupled model. Our results suggest that if the zonal scale of the response in the atmospheric GCM were shorter, the coupled GCM periods would be closer to the observed 12 year period (Deser and Blackmon, 1993; Kushnir, 1994).

Atmospheric stochastic wind stress forcing excites this mode, along with a broad spectrum of oceanic modes. The power spectrum of the stochastically forced case exhibits a reddened background spectrum, on top of which the interdecadal mode can appear as a spectral peak. Although the amplitude of the atmospheric feedback is weak compared to the internal atmospheric variance, coupling plays an essential role in producing a mode whose signature is distinct from other modes in the system. It does so by setting a preferred length scale and providing an eastward return mechanism to complement westward long Rossby wave propagation. This prototype suggests that while active atmospheric feedbacks may be subtle, they can potentially play a role in midlatitude interdecadal variability.

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