Decadal Oscillations in the Mid-Latitude Ocean-Atmosphere System

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1 Introduction

Climate variability at decadal time scales has been an emerging area of interest in recent years. Evidence reported in [1, 2, 3] suggests that the North Pacific ocean-atmosphere system exhibits variability of an oscillatory nature at the time scale of a few decades; it is hypothesized that the ocean-atmosphere coupling plays an important role in the generation of this decadal variability. Similar oscillatory modes have also been observed in the Atlantic [4]; however in this case the existence of a strong thermohaline circulation makes unclear the role of the wind-driven flow in the generation of the oscillations.

The North Pacific decadal oscillations have also been found in various coupled GCMs (Global Climate Models) (e.g. [5, 3]). The sea surface temperature anomalies 1) are large-scale, extending almost across the entire Pacific basin, and 2) have dipolar structure, with the positive and the negative anomalies flipping signs every half-period.

In this work, an idealized model for the large-scale, coupled dynamics of the mid-latitude wind-driven oceanic circulation and the atmosphere has been derived. In this model, the ocean and the atmosphere are coupled through wind stress and heat fluxes at the air-sea interface. The equations are based on global heat and global momentum balances, along the same lines as in [6]. However, the model is significantly simpler than in [6], and allows a good understanding of the physics involved. In particular, it is able to capture the sustained oscillations at decadal time scales, consistent with the above mentioned observations and numerical models.

2 The Model

The geometry of the model consists of an oceanic box with $0 < x < L_x$ as the zonal coordinate, $0 < y < L_y$ as the meridional coordinate and $-H < z < 0$ as the vertical coordinate. The atmospheric box representing the troposphere has the same meridional extent.

In what follows the heat and mechanical balances of the atmospheric and oceanic parts are considered, which lead to the set of equations that describe the model.

2.1 Oceanic Mechanical Balance: Potential Vorticity Equation

The quasi-geostrophic equivalent barotropic potential vorticity equation on a mid-latitude beta plane is considered. After integrating in z and neglecting the relative vorticity term (whose contribution is small for large-scale motions), the ocean flow is governed by,

$$\frac{\partial \psi}{\partial t} - c\frac{\partial \psi}{\partial x} = \frac{R^2}{\rho_w H}\frac{\partial \tau_s}{\partial y} + c\delta\frac{\partial^2 \psi}{\partial x^2}. \tag{1}$$

Here, ψ represents the depth averaged streamfunction, R is the deformation radius, $c = \beta R^2$ is the speed of the long Rossby waves, ρ_w is the sea-water density, τ_s represents the zonal component of the atmospheric surface wind stress and $\delta << L_x$ is the width of the western boundary layer.

Taking τ_s to be constant in x, $\tau_s = \bar{\tau}_s(y,t)$, where the overbar stands for zonal average, and imposing the boundary conditions $\psi = 0$ at $x = 0, L_x$, the solution of (1) can be expressed in the form:

$$\psi = \psi_I(x,y,t) - \psi_I(0,y,t)\exp(-x/\delta) + O(\delta). \tag{2}$$

Here, ψ_I is the depth-averaged streamfunction in the interior and is determined by the wind-stress curl through the relation,

$$\psi_I = \frac{R^2}{\rho_w H}\int_{t-\frac{L_x - x}{c}}^{t} \frac{\partial \bar{\tau}_s}{\partial y} dt'. \tag{3}$$

Therefore, the zonally averaged meridional velocity in the interior has the form:

$$\bar{v}_I = \frac{1}{L_x}\int_{\delta}^{L_x} \frac{\partial \psi_I}{\partial x} dx = -\frac{R^2}{L_x \rho_w H}\int_{t-\frac{L_x}{c}}^{t} \frac{\partial \bar{\tau}_s}{\partial y} dt'. \tag{4}$$

Note that $\bar{v}_I$ at a given time t, is fully determined by the wind stress curl at the times from $t-\frac{L_x}{c}$ to the actual time t, where $\frac{L_x}{c}$ represents the time that it takes a Rossby wave to cross the basin. This equation represents the baroclinic response of the ocean to an imposed wind-stress through propagation of Rossby waves from the eastern boundary. In this idealized formulation, the time that it takes the ocean to adjust to a given wind stress is given by the Rossby wave crossing time $\frac{L_x}{c}$ and is of the order of ten years. If the wind stress undergoes changes at shorter time scales, the ocean-atmosphere system finds itself in a dynamical situation in which the ocean is continuously trying to be in Sverdrup balance with the overlying wind. In doing so, Rossby waves are continuously being generated in the eastern boundary and moving westward carrying with them information about the wind that was blowing at the times they were formed. Therefore, this is the equation that contains the memory of the system.

The following section details how to determine this wind-stress curl, $\frac{\partial \bar{\tau}_s}{\partial y}$.

2.2 Atmospheric Momentum Balance: Zonal Momentum Equation

The natural way of representing large-scale, mid-latitude, atmospheric dynamics in a simple fashion, is to consider the modification by the baroclinic eddies of the mean zonal flow. Specifically, we consider the zonally averaged x-momentum equation, which for quasi-geostrophic motions on a mid-latitude beta plane is given by,

$$\frac{\partial \bar{u}}{\partial t} = f_o \bar{v} - \frac{\partial(\overline{u'v'})}{\partial y} + \frac{1}{\rho_a}\frac{\partial \bar{\tau}}{\partial z}. \tag{5}$$

The overbar stands for the zonal average and the prime stands for the departure from the average, f_o and ρ_a are the Coriolis parameter and atmospheric density respectively and the last term represents vertical transfer of zonal momentum by the stress, $\bar{\tau}$. Integrating over the vertical direction and assuming steady equilibrium, one finds:

$$-\int_0^\infty \rho_a \frac{\partial(\overline{u'v'})}{\partial y} dz = \bar{\tau}_s. \tag{6}$$

This equation represents the zonally and vertically averaged momentum budget for the atmosphere. The zonally averaged wind stress is balanced by the horizontal convergence of momentum flux by the eddies in the entire troposphere.

Using the definition of quasi-geostrophic potential vorticity,

$$q = f_o + \beta y + v_x - u_y + \frac{f_o}{\rho_a}\frac{\partial}{\partial z}(\frac{\rho_a}{S}\theta) \tag{7}$$

where q and θ represent potential vorticity and potential temperature respectively and S is the static stability of the atmosphere which is taken to be constant for simplicity, it is possible to relate the zonally averaged eddy fluxes of momentum, heat and potential vorticity. The relation between these three fluxes is given by,

$$\rho_a(\overline{q'v'}) = \nabla \cdot \vec{F}_{E-P} = -\frac{\partial\left(\rho_a(\overline{u'v'})\right)}{\partial y} + f_o \frac{\partial}{\partial z}\left(\frac{\rho_a}{S}(\overline{\theta'v'})\right), \tag{8}$$

where $\vec{F}_{E-P}$ is the Eliassen-Palm flux. This relation allows the wind stress curl to be expressed in terms of θ and q, which, unlike the momentum u, are conserved following the particle trajectories. This is a desired property when considering the parameterization of eddy fluxes.

2.2.1 Parameterization of Eddy Fluxes

The classical way of parameterizing eddy fluxes uses an analogy with molecular diffusion, considering them proportional to the mean gradients. Thus, the eddy flux divergence is parameterized as:

$$\frac{\partial(\overline{q'v'})}{\partial y} = -k\frac{\partial^2 \bar{q}}{\partial y^2} \tag{9}$$

$$\frac{\partial(\overline{\theta'v'})}{\partial y} = -k\frac{\partial^2 \bar{\theta}}{\partial y^2} \tag{10}$$

where the parameter k represents the eddy diffusivity, i.e. the rate of change of mean square displacement of $\bar{\theta}$ and $\bar{q}$ by baroclinic eddies. In the present model this parameterization is simplified by considering relaxation to the planetary average, that is,

$$\frac{\partial(\overline{q'v'})}{\partial y} = \nu(\bar{q} - \bar{q}_A) \tag{11}$$

$$\frac{\partial(\overline{\theta'v'})}{\partial y} = \nu(\bar{\theta} - \bar{\theta}_A). \tag{12}$$

The subscript “A” stands for the meridional average and ν is the eddy relaxation rate. Following [7] and having in mind the fact that eddy activity is higher near the surface, ν is taken of the form: $\nu = \nu_o \exp(-z/d)$. For the purpose of the present work, this parameterization is qualitatively similar to the classical diffusion analogy, while being simpler.

Because large scales are the focus of this work, relative vorticity is neglected compared to planetary vorticity and vortex stretching, similarly as in the ocean momentum budget.

Finally, considering the change with height of the atmospheric density $\rho_a = \rho_o \exp(-z/D)$, and defining the effective eddy scale-height: $\frac{1}{d_e} = \frac{1}{D} + \frac{1}{d}$, the equation for the wind-stress curl is given by:

$$\frac{\partial \bar{\tau}_s}{\partial y} = d_e \rho_o \nu_o \left(\beta (y - \frac{L_y}{2}) + \frac{f_o}{Sd} (\bar{\theta}_s - \bar{\theta}_{sA}) \right). \tag{13}$$

The boundary conditions are no net wind stress: $\int_0^{L_y} \bar{\tau}_s = 0$, in addition to $\bar{\tau}_s = 0$ at $y = 0, L_y$. The requirement of no net wind stress determines the constant d. The wind stress curl as expressed in the previous equation depends on the zonally average surface potential temperature $\bar{\theta}_s$. Finding an expression for this variable is, therefore, the next step in the derivation of the model.

2.3 Atmospheric Heat Balance: Potential Temperature Equation

The zonally averaged thermodynamic energy equation for the atmosphere is:

$$\frac{\partial \bar{\theta}}{\partial t} = -\frac{1}{\rho_a} \frac{\partial (\rho_a \overline{\theta w})}{\partial z} - \frac{\partial (\overline{\theta' v'})}{\partial y} + \frac{1}{C_{pa} \rho_a} \frac{\partial \bar{Q}}{\partial z} \tag{14}$$

where C_{pa} represents specific heat and $\bar{Q}$ is the diabatic heat flux. Since the typical time scales of atmospheric potential temperature changes, in response to variations in the sea surface temperature, are of the order of weeks, and therefore, much shorter than the typical time scales of evolution of the latter (of the order of years), it is reasonable to neglect the time derivative of $\bar{\theta}$ and assume that the atmosphere adjusts instantaneously to the ocean. Integrating vertically, the remaining equation gives a balance between the horizontal heat flux divergence by eddies in the troposphere and the net diabatic heat fluxes at the vertical boundaries, that is,

$$C_{pa} \int_0^{\infty} \rho_a \frac{\partial (\overline{\theta' v'})}{\partial y} dz = F - r\lambda (\bar{\theta}_s - \bar{T}_s). \tag{15}$$

Here F represents the net radiative incoming flux at the top of the atmosphere and it is given by the difference between the net incoming solar radiation, F_{SOLAR} and the outgoing long-wave radiation, $F_{LONG-WAVE}$. The first one is a prescribed function of latitude chosen to be,

$$F_{SOLAR} = F_o + \tilde{F}_1(y) = F_o + F_1 cos(\frac{\pi y}{L_y}), \tag{16}$$

the latter is estimated by linearizing the gray Stefan-Boltzmann law and is given by,

$$F_{LONG-WAVE} = A + B\bar{\theta}_s. \tag{17}$$

The second term of the right hand side of (15) denotes the exchange of heat flux at the air-sea interface. It is a typical bulk formulae equation which represents relaxation to the zonally averaged sea surface temperature $\bar{T}_s$ with λ being the bulk transfer coefficient and r being the fraction of the earth's surface that is covered by oceans.

Using the eddy flux parameterization in (12), the final expression for the zonally averaged atmospheric potential temperature is,

$$(\bar{\theta}_s - \bar{\theta}_{sA}) = (1-a)\left((\bar{T}_s - \bar{\theta}_{sA}) + \frac{\tilde{F}_1}{r\lambda}\right) \tag{18}$$

where $0 < a < 1$, is given by $a = 1 - \frac{r\lambda}{C_{pa}\nu_o\rho_o d_e + B + r\lambda}$ and contains the contribution from the baroclinic eddies.

The next and final step needed to close the system is to find an expression for the sea surface temperature as a function of the known variables.

2.4 Oceanic Heat Balance: Upper Ocean Heat Content Equation

It is assumed, for simplicity, that the temperature T of the model oceanic layer is independent of depth ($T = T_s$). The vertically integrated heat balance equation is:

$$\frac{\partial T}{\partial t} + \frac{\partial(uT)}{\partial x} + \frac{\partial(vT)}{\partial y} = \frac{\lambda}{C_{pw}\rho_w H}(\bar{\theta}_s - T) + \epsilon\nabla^2{}_h T. \tag{19}$$

Where ϵ is the heat diffusivity coefficient and (u, v) is now the oceanic wind-driven velocity. Note that, except from the western boundary layer balance, this is the only equation where a viscous or diffusive term has been introduced explicitly. In the absence of this term, the warm waters that are transported poleward by the subtropical gyre and the cold waters that are transported equatorward by the subpolar gyre would meet somewhere in the middle forming a discontinuous front. Therefore, this term, represents the only way of direct communication between the gyres which in the present formulation mostly "feel" each other through the atmosphere.

Since the atmospheric model only "sees" the zonally averaged oceanic temperature, the natural way to proceed is to separate T into its zonally averaged part, $\bar{T}$ and its departure from the average, T'. The upper ocean heat content (19) can then be split into the two equations:

$$\frac{\partial\tilde{T}}{\partial t} + \frac{\partial(\overline{vT'})}{\partial y} = \frac{\lambda}{C_{pw}\rho_w H}(-a\tilde{T} + \frac{\tilde{F}(1-a)}{r\lambda}) + \epsilon\frac{\partial^2\tilde{T}}{\partial y^2} \tag{20}$$

$$\frac{\partial T'}{\partial t} + u\frac{\partial T'}{\partial x} + v\frac{\partial T'}{\partial y} + v\frac{\partial\tilde{T}}{\partial y} - \frac{\partial(\overline{vT'})}{\partial y} = -\frac{\lambda}{C_{pw}\rho_w H}T' + \epsilon\frac{\partial^2 T'}{\partial y^2}. \tag{21}$$

Here (18) has been used to write $\bar{\theta}_s$ as a function of $\bar{T}$, and a new variable $\tilde{T}$ has been introduced for convenience. It represents the zonally averaged oceanic temperature after removing the meridionally averaged value of the atmospheric surface potential temperature, so that $\tilde{T} = \bar{T} - \bar{\theta}_{sA}$.

The first thing to notice is that the relaxation rate for $\tilde{T}$ is smaller than that for T' by the factor $a(< 1)$. This is because the atmospheric potential temperature depends on $\tilde{T}$ not on T'. Therefore, on the time scales over which $\tilde{T}$ changes, T' rapidly reaches equilibrium. The next step is to look for an expression that diagnostically relates T' to $\tilde{T}$. In the limit $\frac{v}{L_y} << \frac{\lambda}{C_{pa}\rho_w H}$, the term $v\frac{\partial \tilde{T}}{\partial y}$ balances the relaxation term $\frac{-\lambda}{C_{pw}\rho_w H}T'$. This approximation is not strictly valid in the range of parameters used later, however it is qualitatively correct.

This diagnostic relation allows to express the advective term in the $\tilde{T}$ equation in the following way,

$$\frac{\partial(\overline{vT'})}{\partial y} = -\frac{C_{pw}\rho_w H}{\lambda}\frac{\partial}{\partial y}(\overline{v^2}\frac{\partial \tilde{T}}{\partial y}) \tag{22}$$

so that the effect of meridional advection by the gyres on the zonally averaged temperature is down mean gradient temperature diffusivity proportional to $\overline{v^2}$.

Moreover, it can be shown that the largest contribution to $\overline{v^2}$ comes from the western boundary layer velocity, v_{BL}, so that at first order in δ, $\overline{v^2}$ can be expressed as,

$$\overline{v^2} = \frac{1}{L_x}\int_0^{L_x} dx({v_I}^2 + {v_{BL}}^2 + 2v_I v_{BL}) \simeq \overline{v_{BL}^2} \simeq \frac{\bar{v}_I^2 L_x}{2\delta}. \tag{23}$$

The final expression for the evolution of the oceanic temperature is,

$$\frac{\partial \tilde{T}}{\partial t} - \frac{C_{pw}\rho_w H}{\lambda\Delta}\frac{\partial}{\partial y}(\bar{v}_I^2\frac{\partial \tilde{T}}{\partial y}) = \frac{\lambda}{C_{pw}\rho_w H}\left(-a\tilde{T} + \frac{\tilde{F}_1(1-a)}{r\lambda}\right) + \epsilon\frac{\partial^2 \tilde{T}}{\partial y^2} \tag{24}$$

with boundary conditions of no normal heat flux, $\frac{\partial \tilde{T}}{\partial y} = 0$ at $y = 0, L_y$. Here $\Delta = \frac{2\delta}{L_x}$.

This last equation, together with (4), (13) and (18) compose the closed set of equations of the model.

2.5 Final set of equations

In order to simplify the notation, the following redefinition of variables will be used from here on: $T = \tilde{T}$, $\theta = \bar{\theta}_s - \bar{\theta}_{sA}$, $\tau = \bar{\tau}_s$, and $v = \bar{v}_I$. The system of equations is non-dimensionalized using,

$$y = L_y y^* \tag{25}$$

$$t = \alpha^{-1}t^* \quad (\alpha = \frac{\lambda}{C_{pa}\rho_w H}) \tag{26}$$

$$(\theta, T) = \frac{F_1}{r\lambda}(\theta^*, T^*) \tag{27}$$

$$v = \frac{d_e\rho_o\nu_o L_y}{\rho_w H}v^* \tag{28}$$

$$\tau = d_e\rho_o\nu_o\beta {L_y}^2\tau^*. \tag{29}$$

The set of equations in their non-dimensional form are given by,

$$\theta = (1-a)[T + f(y)] \tag{30}$$

$$\frac{\partial \tau}{\partial y} = y - \frac{1}{2} + \gamma\theta \tag{31}$$

$$v = -\frac{1}{t_o}\int_{t-t_o}^{t} \frac{\partial \tau}{\partial y} dt' \tag{32}$$

$$\frac{\partial T}{\partial t} - \mu\frac{\partial}{\partial y}(v^2\frac{\partial T}{\partial y}) = -aT + (1-a)f(y) + \sigma\frac{\partial^2 T}{\partial y^2} \tag{33}$$

where

$$f(y) = cos(\pi y) \tag{34}$$

$$\gamma = \frac{f_o}{Sd}\frac{F_1}{r\lambda\beta L_y} \tag{35}$$

$$\mu = (\frac{d_e\rho_o\nu_o}{\rho_w H\alpha})^2\frac{1}{\Delta} \tag{36}$$

$$\sigma = \frac{\epsilon}{\alpha(L_y)^2}, \tag{37}$$

$$t_o = \frac{L_x}{c}\alpha \tag{38}$$

where t_o is the non-dimensional delay time given by the ratio of the Rossby wave crossing time and the oceanic temperature decay time, and the asterisks have been dropped.

Combining (30), (31) and (32) into a single equation for the velocity v as a function of the temperature T, one finds,

$$v = -(y - \frac{1}{2}) - \gamma(1-a)f(y) - \gamma(1-a)\frac{1}{t_o}\int_{t-t_o}^{t} T(y,t')dt'. \tag{39}$$

This relation, given v at $t = 0$, can also be expressed as:

$$\frac{\partial v}{\partial t} = -\gamma(1-a)\left(T(y,t) - T(y,t-t_o)\right) \tag{40}$$

The final model set of equations is reduced to the equations (33) and (39). The advantages of this formulation are:

- Simplicity (only one dimension in space, and only two dependent variables, v and T)
- No need to resolve the boundary layer explicitly

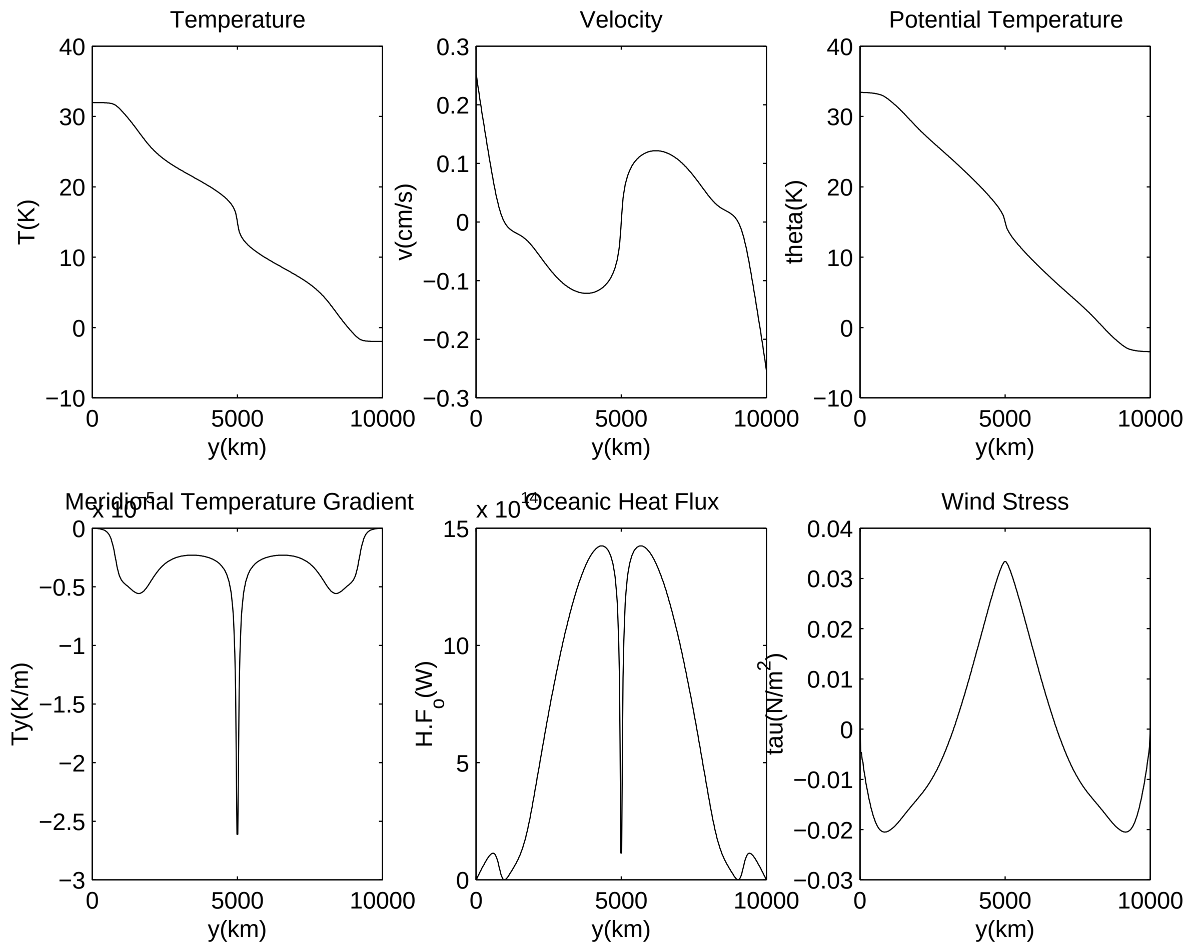

Figure 1: Latitudinal dependence of various physical magnitudes at a given time

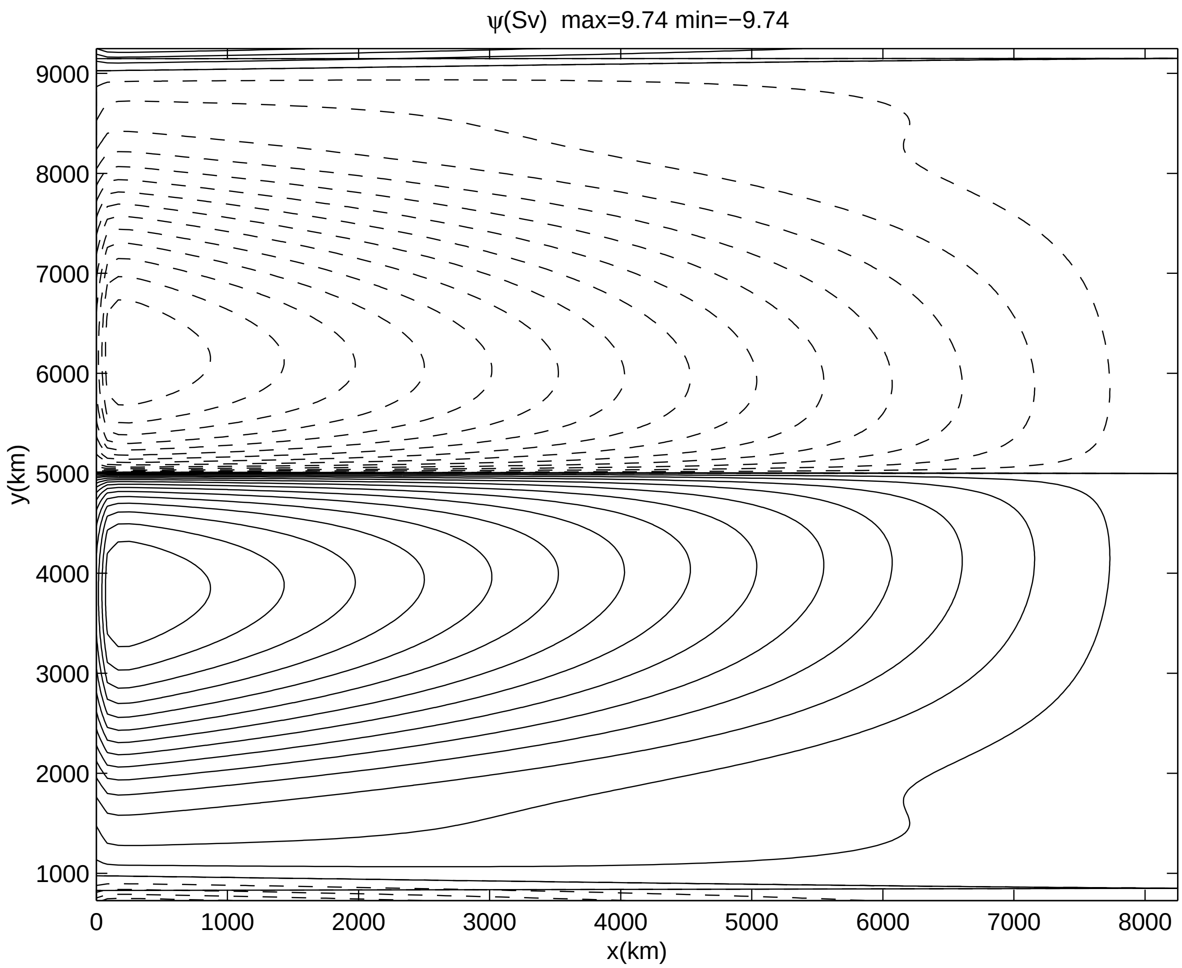

Figure 2: Streamfunction at a given time

$L_x = 0.825 \times 10^7 m$	$\rho_o = 1.25\text{kg } m^{-3}$	$A = 200 W m^{-2}$	$S = 5 \times 10^{-3}\ K\ m^{-1}$
$L_y = 10^7 m$	$\rho_w = 1000\text{kg } m^{-3}$	$B = 2.475 W m^{-2} K^{-1}$	$\beta = 2 \times 10^{-11} m^{-1} s^{-1}$
$H = 10^3 m$	$C_{pa} = 1000 J K^{-1}\text{kg}^{-1}$	$F_1 = 125 W m^{-2}$	$f_o = 10^{-4} s^{-1}$
$\Delta = 0.015$	$C_{pw} = 4000 J K^{-1}\text{kg}^{-1}$	$\lambda = 23 W m^{-2} K^{-1}$	$R = 2.8 \times 10^4 m$
$D = 10^4\ m$	$\nu_o = 10^{-6} s^{-1}$	$r = 0.3$	$\sigma = 3 \times 10^{-3}$

Table 1: Parameter values

3 Results and Discussion

The system of equations presented in the previous section is solved numerically using Crank-Nicholson scheme for the integration of the temperature equation, centered differential scheme for the spatial derivatives and the trapezoidal rule to solve the integral in time for the velocity equation. At each time step, $t + dt$, the knowledge of the temperature field at times from $t - t_o$ to t is required. The radiative equilibrium temperature is used as the initial guess and it is assumed that the previous history of the temperature field, when unknown, is constant in time and equal to the initial condition.

A Hopf bifurcation is found when moving in the parameter space increasing the importance of the non linearity with respect to the diffusivity. This corresponds to an increase in the ratio $\frac{\mu}{\sigma}$ in (33). The results presented below have been obtained at a point in the parameter space near the bifurcation point and inside the region of sustained oscillations.

A list of the parameter values used in this calculation is presented in table 1.

Figure 1 shows a snapshot of the oceanic temperature and its meridional gradient,of the wind-driven oceanic velocity and of the oceanic heat flux, as well as the atmospheric potential temperature and wind stress fields as a function of latitude for a given time. They all have realistic values, although the equator to pole oceanic temperature difference, the oceanic velocity and the wind stress are weaker than observed. A contour plot of the corresponding oceanic streamfunction is presented in figure 2. Note that in this figure (and from here on), the displayed meridional extent is the central region between the two minima of wind stress, where the subpolar and subtropical gyres are enclosed.

3.1 The oscillations

The evolution of the oceanic temperature and velocity fields during about 3/2 of an oscillation period is presented in figure 3. Their corresponding anomalies are shown in figure 4. The basic features of the oscillations are:

- They are decadal oscillations with a period of about 40 years (slightly more than twice the delay time (17 years)).
- They are large-scale, extending over most of the basin.

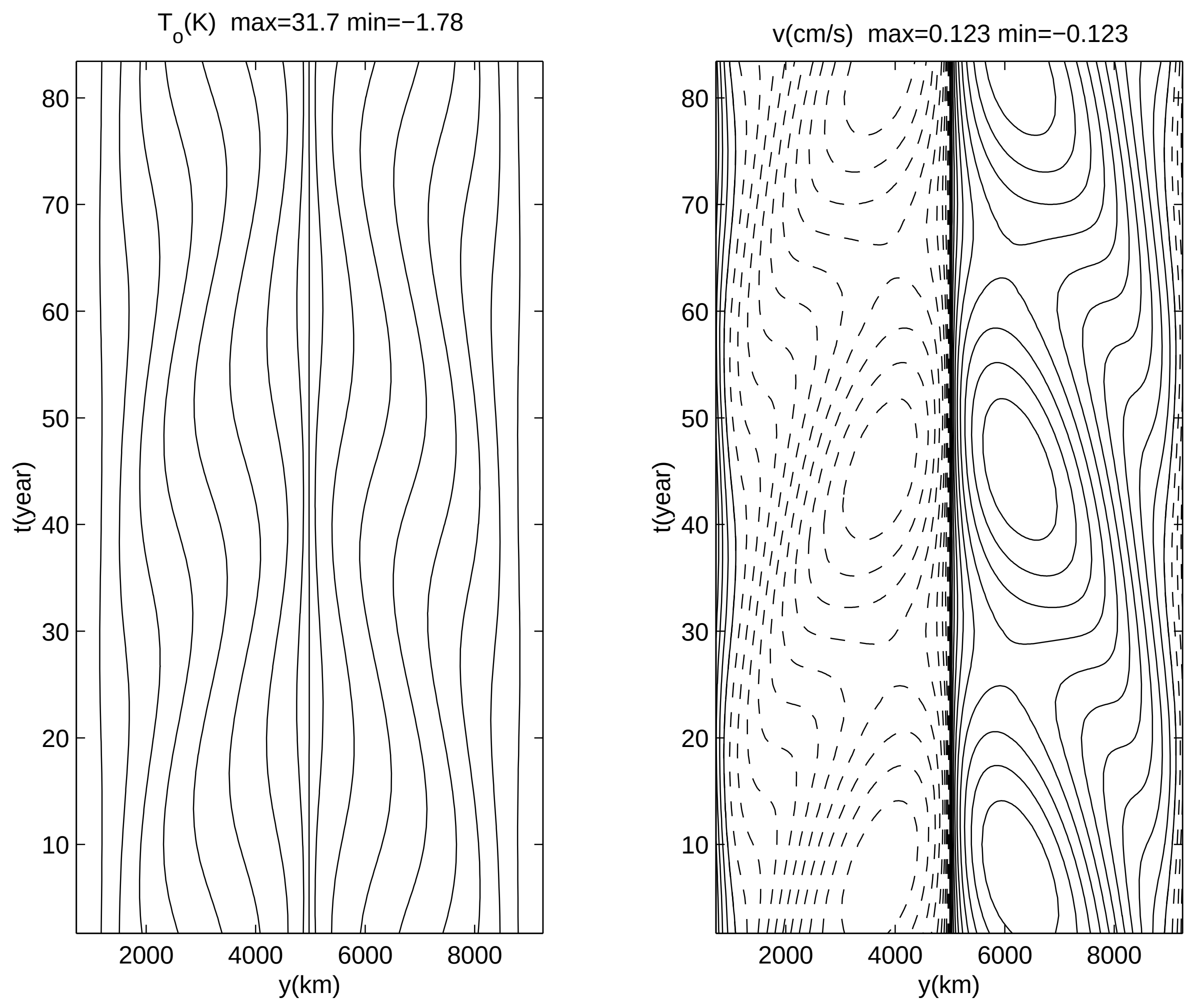

Figure 3: Time evolution of oceanic temperature and velocity

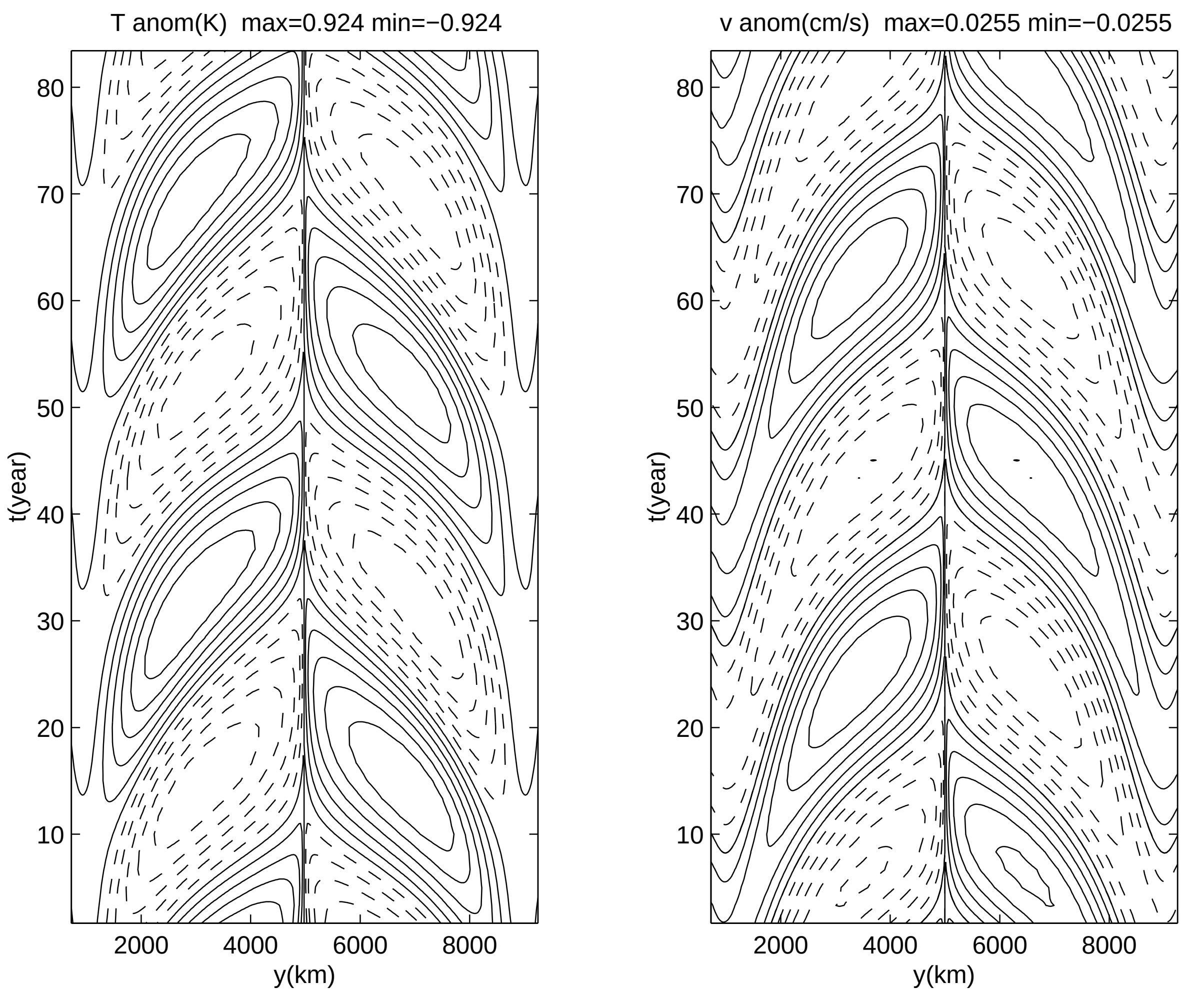

Figure 4: Time evolution of oceanic temperature and velocity anomalies

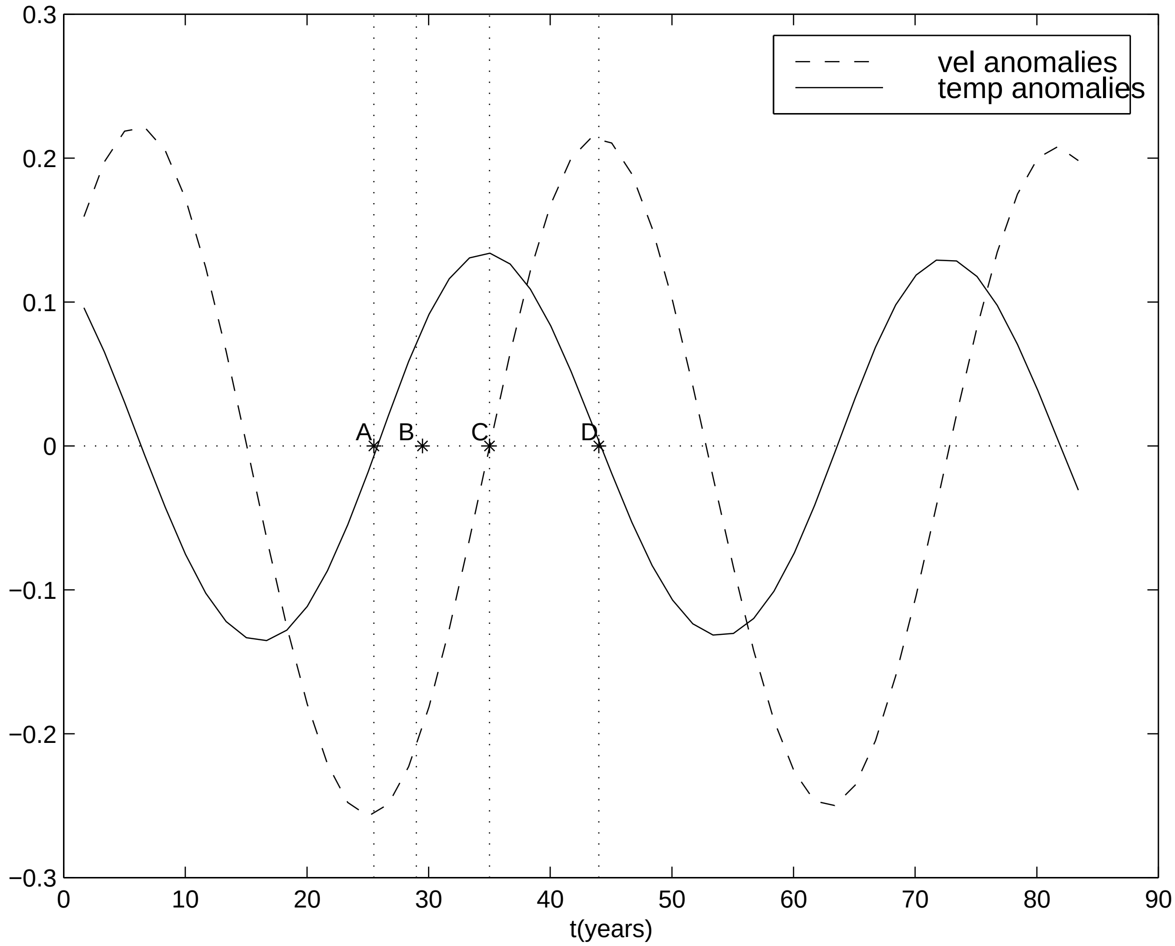

Figure 5: Time evolution of normalized oceanic temperature and velocity anomalies at latitude $y = 3500km$

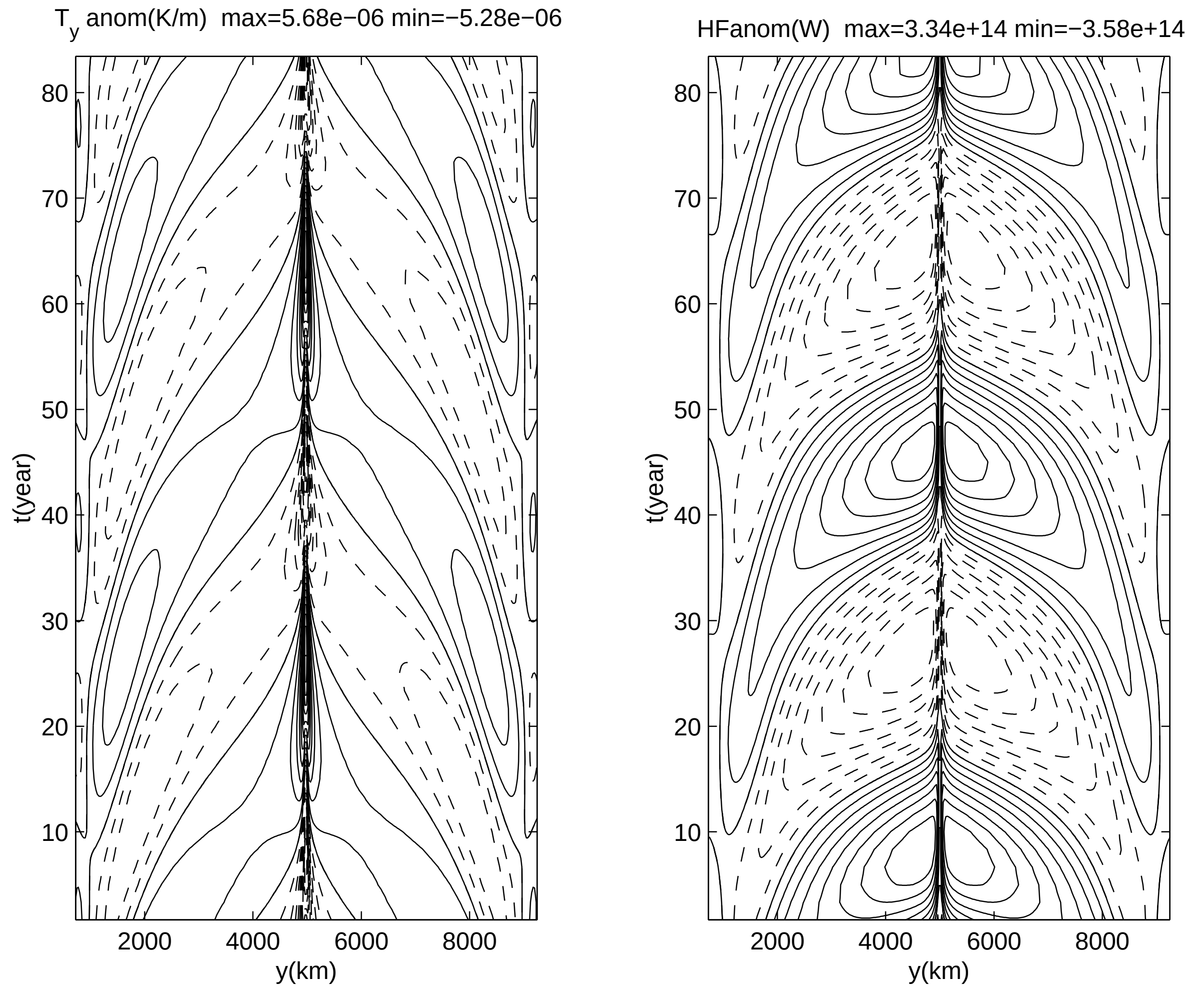

Figure 6: Time evolution of oceanic meridional temperature gradient and heat flux anomalies

- They are small in amplitude, with a root meansquare variance of about 1% in the oceanic temperature and about 10% in the oceanic velocity.
- They are characterized by temperature anomalies, which are antisymmetric about the zero wind stress curl line, and which flip signs in each phase of the oscillation.
- The oceanic temperature and velocity fields are 90 degrees out of phase.

Figure 5 shows the time evolution of the temperature and the velocity anomalies (normalized by their mean values), at a fixed latitude ($y = 3500km$) in the subtropical gyre. Consider the point in time in which there are no temperature anomalies and the circulation in the gyres is weakest. This corresponds to point A in figure 5. As time goes on, the subtropical gyre becomes warmer while the subpolar gyre becomes colder,(point B in figure 5). As a consequence, the meridional temperature gradient is reinforced, which causes an increase in the wind-stress curl. If the ocean responded instantaneously, the resulting stronger circulation in the gyres would advect temperature opposing to the temperature gradient anomaly until equilibrium is reached. However, in the presence of a lag time, the velocity in the gyres depends not only on the actual, increased wind-stress but also on the wind-stress which was forcing the ocean in the past 17 years. At that time, there was a negative temperature anomaly in the subtropical gyre and a positive one in the subpolar gyre and therefore a reduced wind-stress curl. The reduced velocity allows the temperature anomaly to grow until about 10 years later, when the effect of the increased temperature spins up the gyre. Thus, it is not until some years, that the oceanic circulation anomaly starts changing sign in response to the variations in temperature (point C in figure 5). As the circulation becomes stronger, it begins to reduce the temperature anomaly and eventually changes its sign. Then, the second phase of the oscillation starts, (point D in figure 5).

Note that the storage term in the oceanic temperature equation plays an important role. In its absence an initial perturbation in meridional distribution of oceanic temperature would be "instantaneously" damped through heat transport and fluxes at the air-sea interface, arresting the oscillations. Also note that this would not be the case if, in the delay equation (40), the temperature modes at time $(t - t_o)$ had a larger amplitude than those at time t.

The evolution of the oceanic meridional temperature gradient and of the oceanic heat flux anomalies are shown in figure 6.

4 Conclusions and Future Work

- An idealized model for the coupled large-scale dynamics of the midlatitude atmosphere and the wind-driven ocean circulation has been derived.
- In this model, the ocean and the atmosphere are coupled through wind stress and heat fluxes at the air-sea interface.

- The formulation can be reduced to two equations. One describes the time evolution of the oceanic temperature and the other represents the transient oceanic velocity response to the wind-stress curl.

- Sustained oscillations at decadal time scales have been found which resemble those found in the observations and in the results of more comprehensive numerical models. They have small amplitude and large-scale meridional extent.

- These oscillations owe their existence to the time lagged response of the ocean to changes in the wind stress curl associated with Rossby wave propagation. An adjustment time of the oceanic temperatures to changes in the circulation is also required.

Areas of future work could include:

- Further exploration of the parameter space.

- Analysis of the solutions in comparison with observations as well as with results from coupled GCMs.

- Addition of an asymmetric component in the radiative forcing. Can we obtain chaotic behavior?.

- Multiple equilibria?. Equations (33) and (39) can be simplified further by considering the spatial extension in Fourier series of T and v and retaining only the mode that is directly forced by radiation. As a result, the system is reduced to two ordinary differential equations in time. A calculation of the steady states of this system for different parameter values led to multiple equilibria for sufficiently small values of Δ, i.e. for the case of a narrow western boundary layer. An exploration of the possible existence and causes of multiple equilibria in the non-truncated system could be a next step.

- Teleconnections. The mechanisms by which the ocean basins communicate with each other through the atmosphere are still not well understood. The model presented here can be expanded to represent two ocean basins (which would have different delay times) so that it can be used to explored these mechanisms.

5 Acknowledgments

This work has been done in collaboration with Paola Cessi. Steve Meachan, Phil Morrison, Michael Shelley and George Veronis are gratefully acknowledged for their valuable input.

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