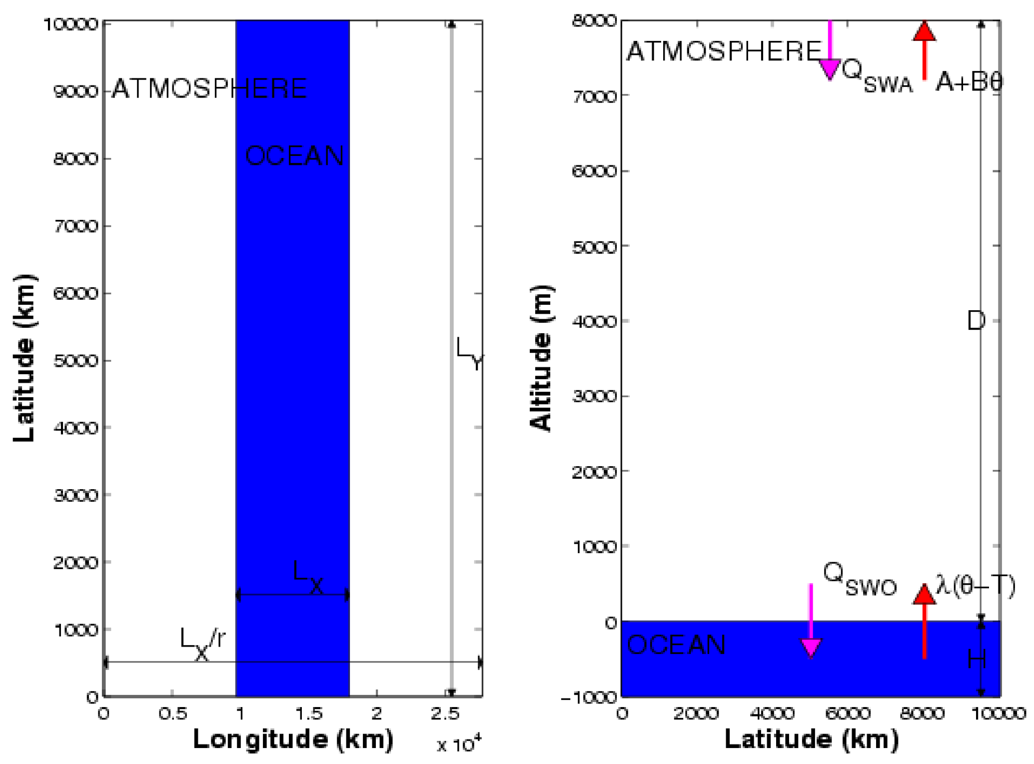
Decadal Oscillations in a Simplified Coupled Model due to Unstable interactions Between Zonal Winds and Ocean Gyres

O. Arzel and T. Huck

LPO/UBO UFR Sciences, 6 avenue Le Gorgeu, BP 209, 29285 BREST CEDEX FRANCE
contact: oarzel@univ−brest.fr

Motivation: *Coupled climate models show variability in the North Pacific due to unstable air−sea interactions between the subtropical gyre circulation and the Aleutian low pressure system (Latif and Barnett (1994). Cessi (2000) and Gallego and Cessi (2000) exhibit a simple coupled model that produces decadal oscillations due to the coupling between ocean gyres heat transport and wind stress where the key element is the Rossby wave propagation. We look hereafter into the physical mecanism of such an oscillation.*

The model



The domain geometry consists in one hemisphere with a single rectangular basin which dimensions are approximately the ones of the North Pacific. The only external forcing is the prescribed incoming solar radiation at the top of the atmosphere Q_{SWA} and at the ocean surface Q_{SWO}.

The diagnostic atmospheric model

Zonally averaged one layer of fixed height D, with constant stratification S in potential temperature ϑ, in energy balance between horizontal eddy diffusion, incoming short wave, outgoing longwave, and surface heat flux.

➡ Surface air temperature $\vartheta_s(y)$ determined through heat balance: $-C_{pa}\rho_s k_s d_e \partial_y^2 \vartheta_s = Q_{SWA}(y) - A - B\vartheta_s - r[Q_{SWO}(y) + \lambda(\vartheta_s - \overline{T}_s)]$

➡ Zonal wind $\tau(y)$ determined through momentum balance: $\tau - \frac{d_e k_s}{\gamma}\partial_y^2 \tau = -\rho_s k_s d_e d^{-1}[\beta d + \frac{f}{S}(\partial_y \vartheta_s + L_\rho^2 \partial_y^3 \vartheta_s)]$

where $\int_0^{L_y} \tau\, dy = 0$ is used to determine the vertical eddy diffusion scale height d in the atmosphere, L_ρ is the baroclinic radius of deformation.

The prognostic ocean model

One « thermocline » of constant depth H, homogeneous, anisotropic horizontal diffusion, forced by surface wind stress and heat fluxes.

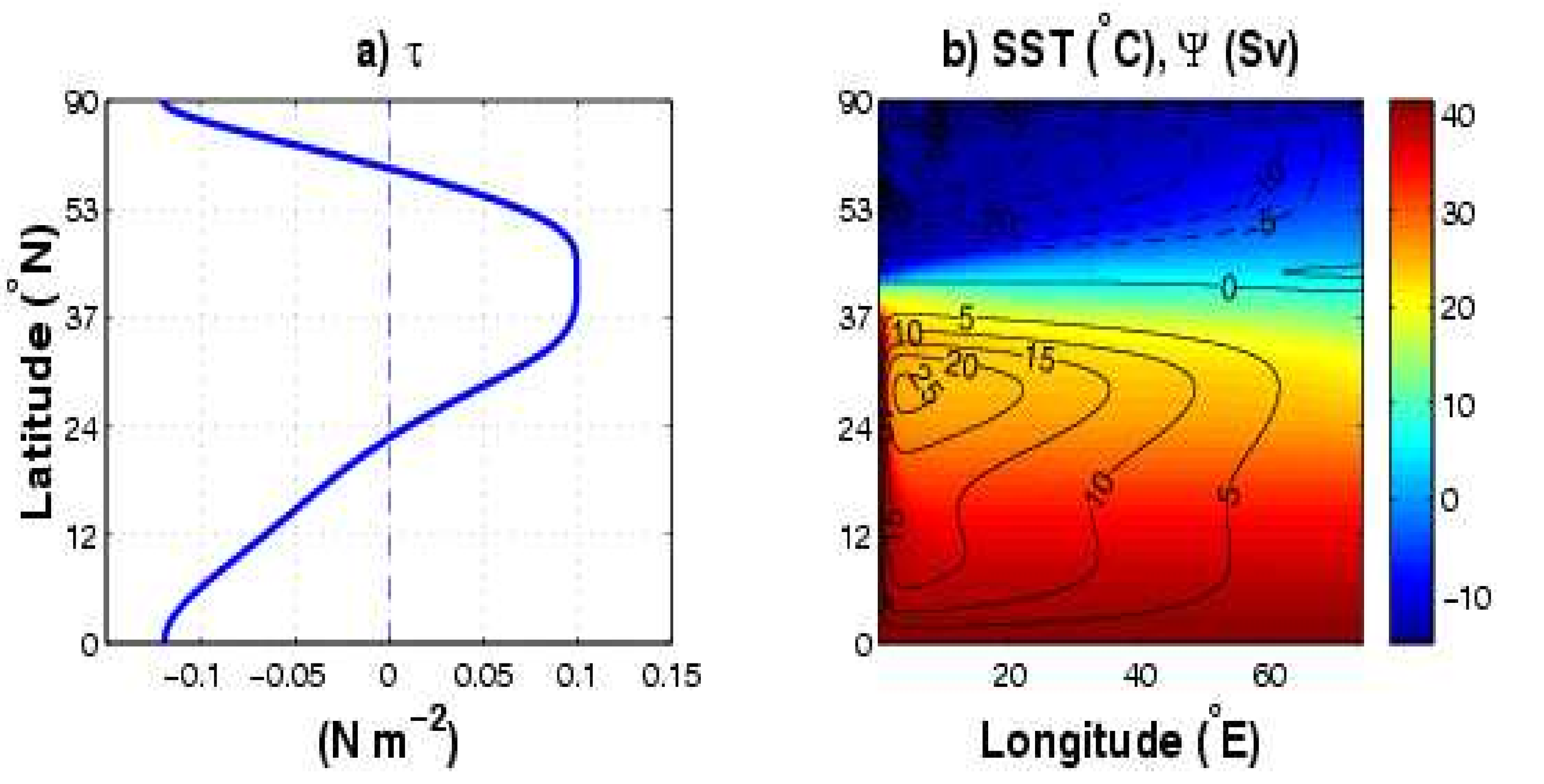
➡ Vertically homogeneous temperature T(x,y) integrated through heat balance: $C_{pw}\rho_w[H\partial_t T + J(\Psi, T)] = F(y) + \lambda(\vartheta_s - T) + C_{pw}\rho_w \nabla(K_h \nabla T)$

➡ Horizontal streamfunction $\Psi(x,y)$ integrated from large−scale limit of quasi−geostrophic equivalent barotropic vorticity equation:

$$\partial_t \Psi - c\partial_x \Psi = \frac{R^2}{\rho_w}\partial_y \tau + A_{hx}\partial_x^2 \Psi + A_{hy}\partial_y^2 \Psi$$

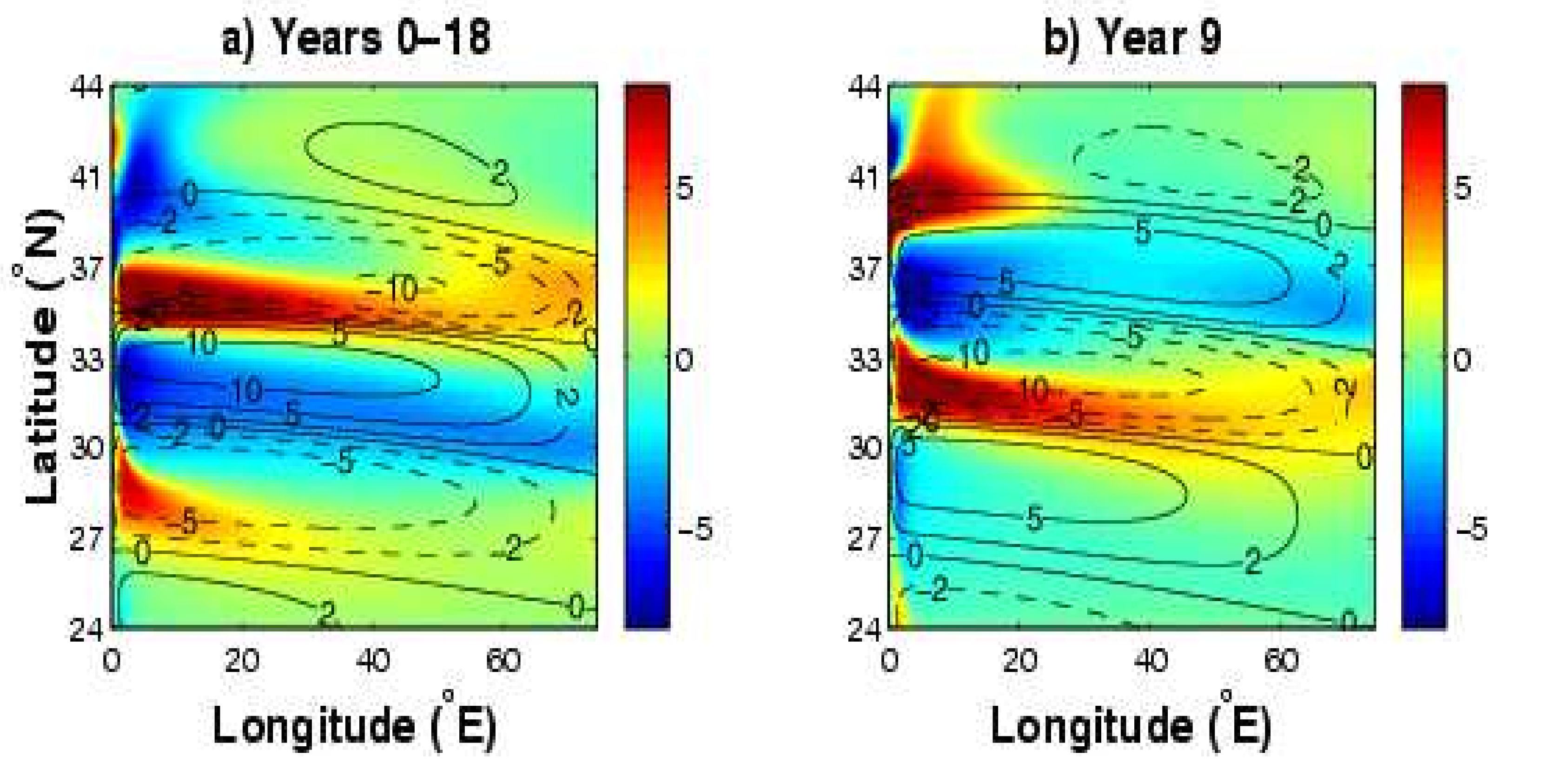
, where R is the Rossby radius of deformation and $c = \beta R^2$ is the long Rossby wave velocity.

The mean state



a) In spite of the simplicity and crudeness of the model, the wind structure is relatively well represented with trade winds in the tropics, westerlies in the mid−latitudes and easterlies poleward of 65°N.
b) The corresponding transport streamfunction contours in Sverdrups (Sv, 1 Sv = $10^6\, m^3 s^{-1}$) superimposed on the oceanic temperature (SST) field. The maximum transport is 50 Sv (cyclonic dashed) in the subpolar gyre (relatively homogeneous SST) and 25 Sv in the subtropical gyre. The circulation produces a strong thermal front at the intergyre boundary (40°N).

The model variability



For a certain range of parameters, decadal oscillations emerge. Here we display the transport streamfunction anomalies contours (Sv) superimposed on SST anomalies (°C) south of the intergyre boundary during one oscillation period (18.2 years). The SST anomalies propagating southward (~2.6 mm/s) are reinforced by the circulation anomaly slightly shifted southward. A positive SST anomaly leads very quickly to the formation of a negative (dashed) circulation anomaly which slows down the western boundary current. Then a cold SST anomaly appears and generates a positive circulation anomaly which leads to the formation of a warm SST anomaly. In this way, the cycle repeats itself. The SST anomalies take about 35 years to reach the tropics where they vanish near the western boundary.

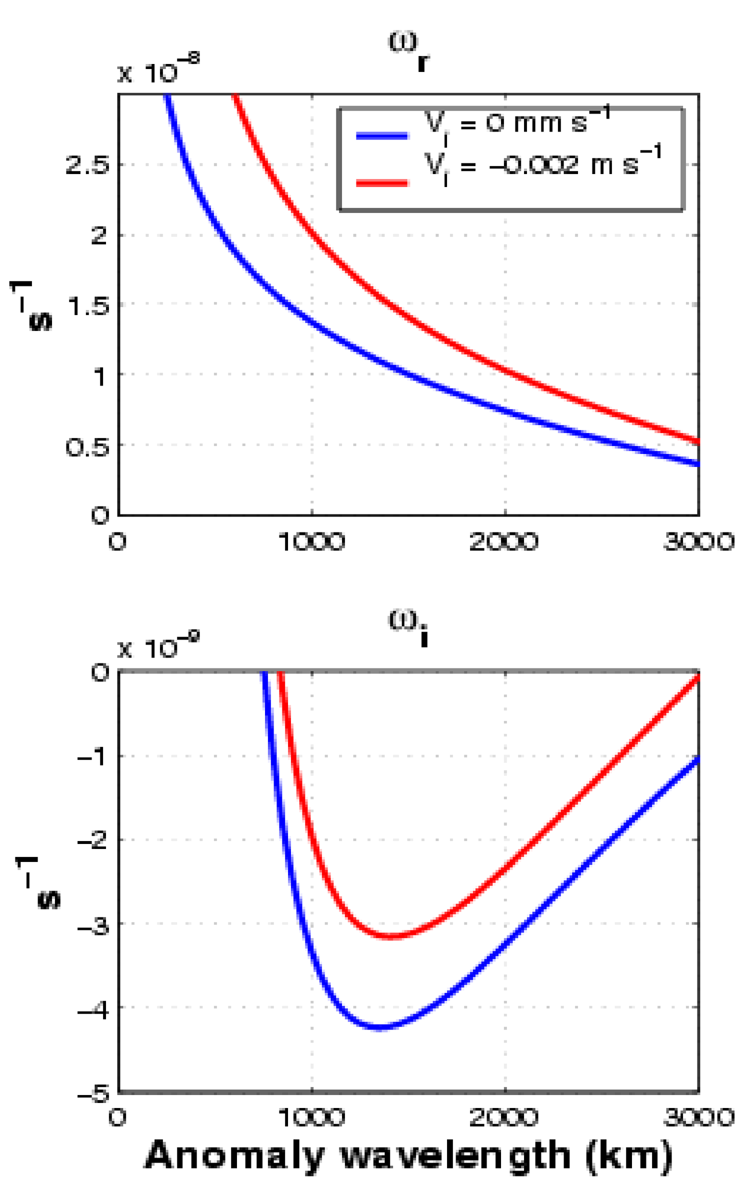
A simplified model

The zonality of streamfunction and SST anomalies lead us to make a zonally averaged simplified model, in order to understand the origin of the southward propagation and the growth rate of SST anomalies. In the zonally−averaged heat balance anomaly equation, we approximate the two dominant advective terms:

$$\overline{\partial_y \Psi' \times \partial_x T} \approx \partial_y \overline{\Psi'} \times \overline{\partial_x T} \quad , \quad \overline{\partial_x \Psi \times \partial_y T'} \approx H V_i \times \partial_y \overline{T'}$$

where the bar represents the zonal mean, and V_i the mean meridional velocity in the subtropical gyre. Hence, we obtain a set of equations for the zonally averaged anomalies. We assume that all the anomalies $(T', \tau', \Psi', \vartheta')$ are harmonic perturbations in latitude on uniform mean state. Then we obtain a dispersion relation for the zonally averaged SST anomalies.

We display here, for the solution propagating southward and for differents values of V_i the real and imaginary parts of the wave frequency, obtained through the resolution of the dispersion relation: It shows a positive growth rate that reaches a maximum for an anomaly wavelength around 1300 km, which is close to the global model oscillation wavelength. The solution propagating northward has a negative growth rate, hence only the solution propagating southward emerges.



Note that the propagation term of baroclinic Rossby waves does not appear in the zonal mean equivalent barotropic vorticity equation, and yet a positive growth rate with a southward propagation appears. This figure shows that the mean current V_i does not modify significantly the growth rate and the southward propagation of the solution.

Hence, for more simplicity, we neglect the mean meridional advection of temperature anomalies and the diffusive and viscosity terms which are not essential for the positive growth rate and southward propagation of the anomalies. Moreover, for small scale perturbations (in the meridional direction) the atmospheric heat balance and momentum balance *(1)* equations are simplified. This leads to a new set of equations reduced to an equation similar to a wave equation *(2)*:

$$(1)\ \tau' \sim \frac{\rho_s d_e \gamma g}{f\Theta}\partial_y \vartheta_s' \quad , (2)\ \partial_t^2 \overline{T'} - \eta^2 \partial_y \overline{T'} = 0 \quad \textit{with} \quad \eta^2 = \frac{R^2 |\overline{\partial_x T}| g \gamma r \lambda}{C_{pw}\rho_w H k_s f \Theta}$$

Its resolution provides two solutions: a damped solution propagating northward, and an amplified solution propagating southward, again only the solution propagating southward emerges. For a wavelength anomaly of 1300 km (corresponding to the maximum growth rate), and a mean zonal gradient $\overline{\partial_x T} = -2\times10^{-6}\, K\, m^{-1}$ in the subtropical gyre, the southward phase velocity is 2.7mm/s.

Conclusions: The mean zonal gradient $\overline{\partial_x T}$ is the energy source term of the positive growth rate of the SST anomalies. The southward propagation is partly due to the mean meridional current V_i, but mainly to a mechanism of air−sea coupled waves out of phase in latitude which self−advect.

References : Cessi, P., 2000: Thermal feedback on wind−stress as a contributing cause of climate variability. *J. Climate.*, **13**, 232−244 ; Lacasce, J. H., 2000: Baroclinic Rossby waves in a square basin. *J. Phys. Oceanogr.*, **30**, 3161−3178 ; Latif, M. and Barnett, T. P., 1996: Decadal variability over the north Pacific and north America: dynamics and predictability. *J. Climate.*, **9**, 2407−2423.