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

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Abstract

Decadal oscillations are obtained in a simple coupled model, consisting in a one-layer bidimensional ocean model and a one-layer unidimensional energy-balance atmospheric model, including a parameterization of the zonal winds 'à la' Green, interacting through heat and momentum fluxes. The range of parameters leading to oscillations is determined through several numerical experiments: ocean eddy-diffusion smaller than estimations from observations is required for oscillations to be sustained. The essential physical processes are clarified, but the spatial pattern of the oscillation remains too complex for a simple low-order conceptual mechanism to emerge.

Motivation

- Decadal variability is found in surface observations in the North Paci
Atlantic (Sutton and Allen 1997, Mann *et al.* 1998), and the subduct
propagation of mid-latitude surface anomalies may influence tropical
like ENSO (Gu and Philander 1997).

- Coupled climate models show variability in the North Pacific due to u
interactions between the subtropical gyre circulation and the Aleutia
(Latif and Barnett 1994).

- Cessi (2000) and Gallego and Cessi (2000) exhibit a simple coupled m
conservation principles, that produce decadal oscillations due to the c
gyres heat transport and wind-stress, the key element being the delay
circulation to changes in the wind-stress.

The diagnostic atmospheric model

• zonally-averaged one layer of fixed height D, with constant stratification
between horizontal eddy diffusion, incoming shortwave, outgoing longwav

→ surface temperature $\theta(y)$ determined through heat balance:

$$-C_{pa}\rho_a k_s d_e \partial_y^2\theta = Q_{\mathrm{SWA}} - (A + B\theta) - r[Q_{\mathrm{SWO}} + \lambda(\theta$$

→ zonal wind $\tau(y)$ determined through momentum balance:

$$\tau - \frac{d_e k_s}{\gamma}\partial_y^2\tau = -\frac{\rho_a k_s d_e}{d}\left[\beta d + \frac{f}{S}(\partial_y\theta + L_\rho^2\partial_y^3\theta)\right.$$

where $\int_0^{L_y}\tau dy = 0$ is used to determine the vertical eddy diffusion scale d

$$\text{effective scale height } d_e = \frac{dD}{d + D} \ (\sim 3600\ \mathrm{m})$$

$$\text{baroclinic radius of deformation } L_\rho = \left(\frac{dd_e gS}{f^2\Theta}\right)^{1/2} \ (\sim$$

The prognostic ocean model

- one ”thermocline” layer of fixed depth H with wind-driven dynamics

→ vertically homogeneous temperature $T(x, y)$ integrated through heat b

$$C_{pw}\rho_w[H\partial_t T + J(\Psi, T)] = Q_{\mathrm{SWO}} + \lambda(\theta - T) + C_{pw}\rho_w\nabla$$

→ horizontal streamfunction $\Psi(x, y)$ integrated from large-scale limit of q
barotropic vorticity equation:

$$\partial_t\Psi - \beta R^2\partial_x\Psi = \frac{R^2}{\rho_w}\partial_y\tau + \nabla\cdot(A\nabla\Psi)$$

where R is the Rossby radius of deformation (first baroclinic mode),
K and A are the eddy diffusion coefficients for temperature and momentu

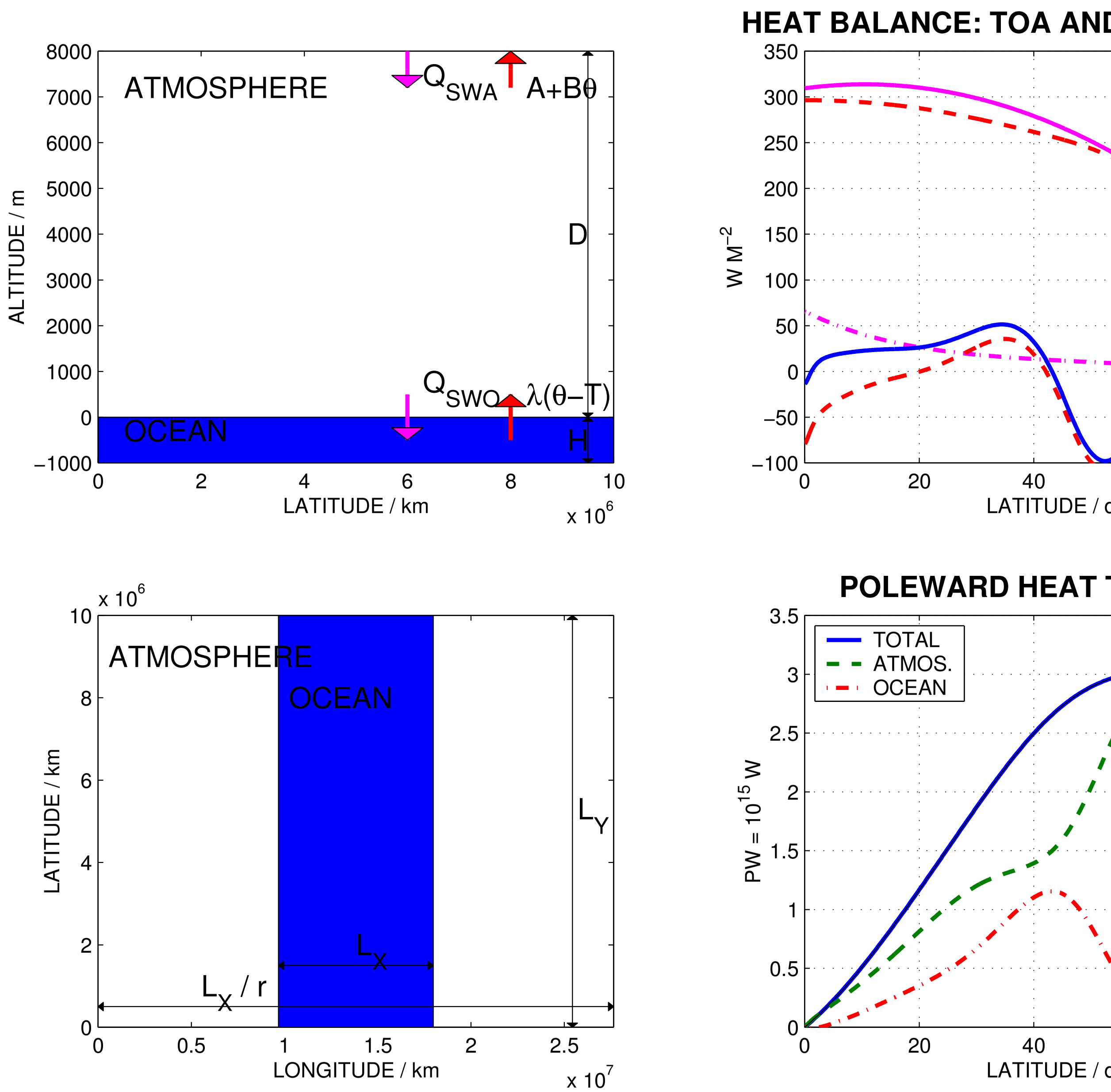
Coupling through surface heat flux $Q(x, y) = Q_{\mathrm{SWO}}(y) + \lambda\,(\theta(y) - T(x,$

Configuration and parameters

- Cartesian coordinates single-hemisphere β-plane
- atmosphere of terrestrial width at 45°N (28000 km), ocean of North I
  both extending from Equator to Pole

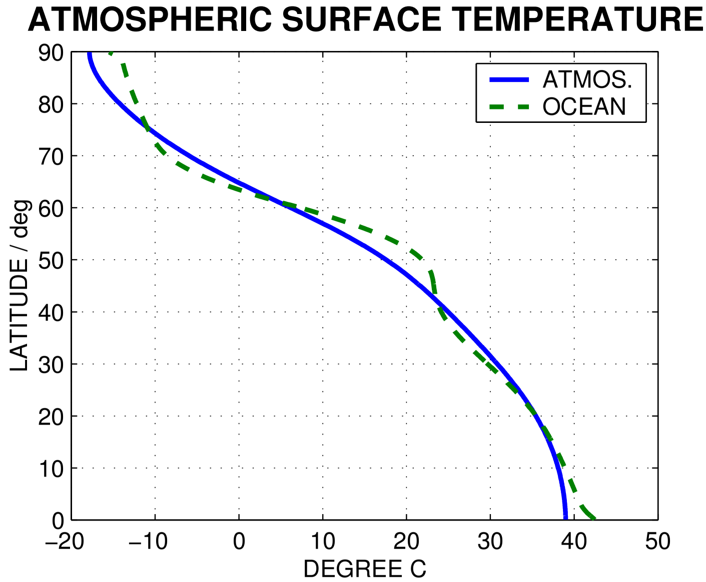
The only external forcing is the prescribed incoming solar radiation at th
and at the ocean surface. In spite of the simplicity and crudeness of the m
relatively well represented with trade winds in the tropics, westerlies in th
easterlies poleward of 80°N.

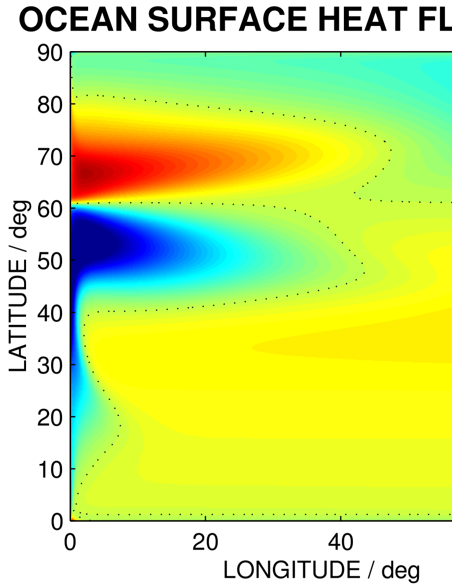
Relatively low eddy-diffusivity is used in the ocean ($K = A = 200$ m^2 s^{-1}
sustained. However a proper representation of the frictional western boun
viscosity in the zonal direction (2000 m^2 s^{-1}), what is not necessary in co
observations.

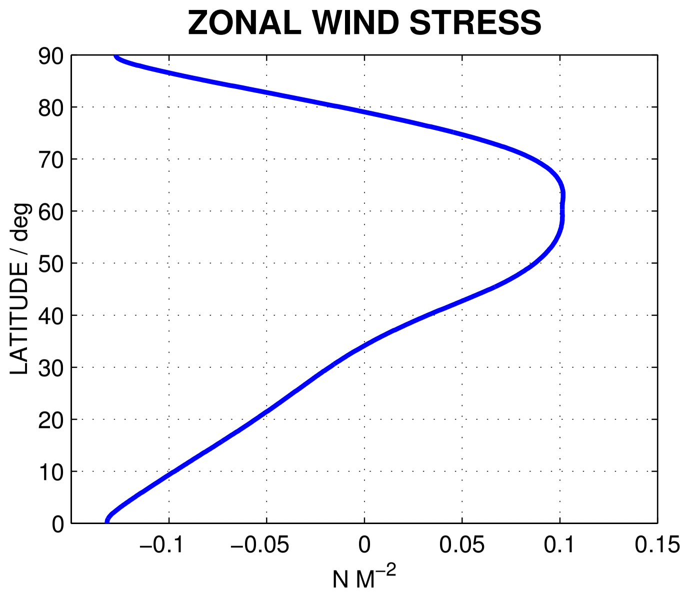


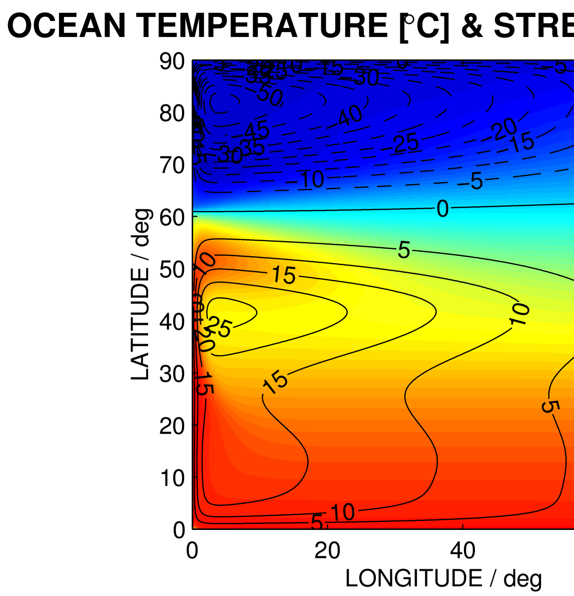
ATMOSPHERE
OCEAN
Q_{SWA}
A+Bθ
D
Q_{SWO}
λ(θ−T)
H
ALTITUDE / m
LATITUDE / km
x 10^6
HEAT BALANCE: TOA AND
W M^-2
LATITUDE / d
ATMOSPHERE
OCEAN
L_X
L_X / r
L_Y
LATITUDE / km
LONGITUDE / km
x 10^7
POLEWARD HEAT T
TOTAL
ATMOS.
OCEAN
PW = 10^15 W
LATITUDE / d

The mean state









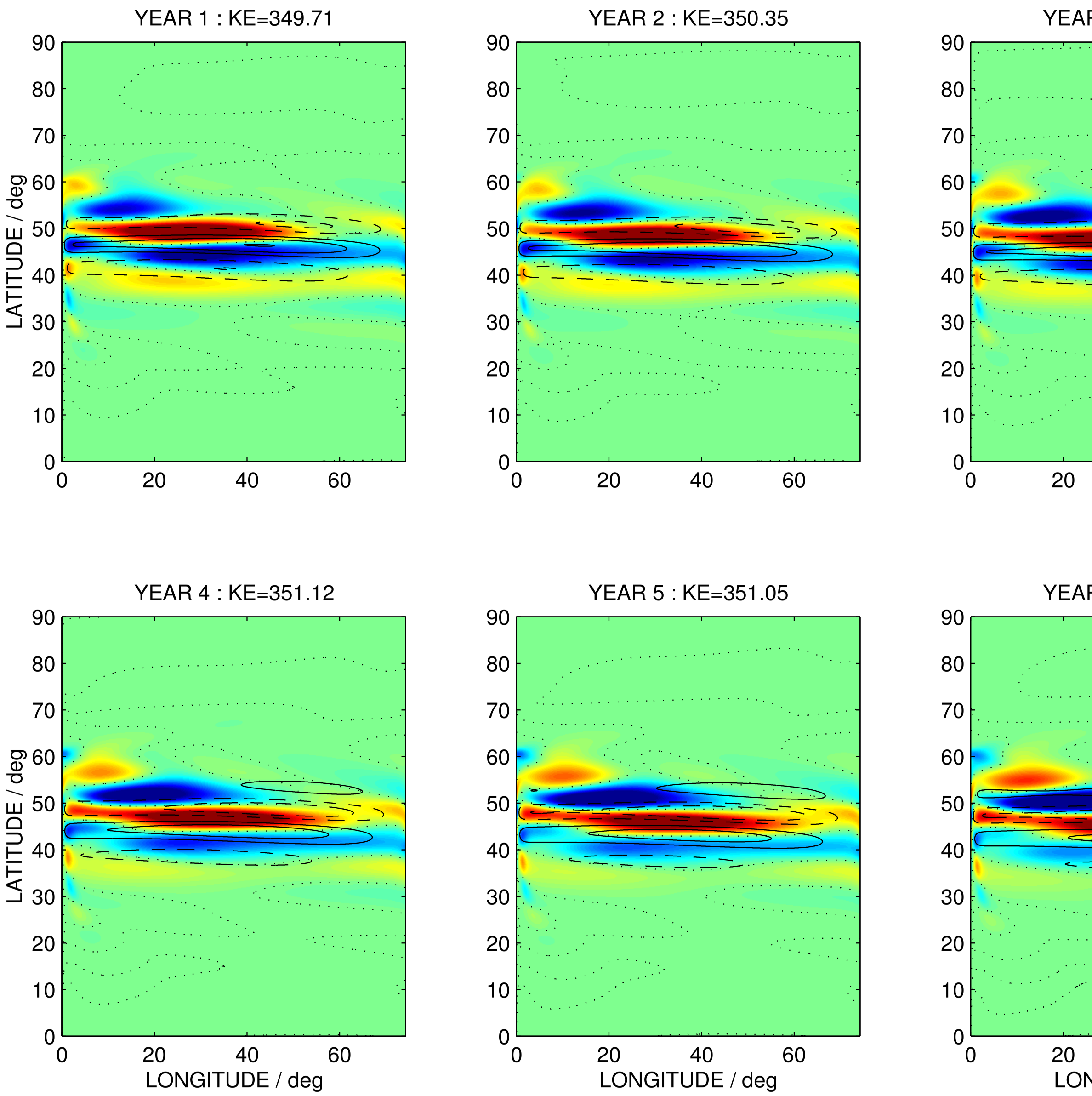
Decadal oscillations

For a limited range of parameters, consisting in low atmospheric and ocea
oscillations settle in place of a steady state. Periods are typically in the r
Temperature anomalies reach several degrees C in the ocean, but hardly

The variability is located south of the intergyre boundary, between 30°N
position of the zero wind-stress curl line remains stationnary.

The anomalies structure is complex with a high meridional wavenumber.
are initiated at the separation point of the western boundary current, jus
Then, they extend easterly and move slowly southward. They die off afte
into the boundary current, almost 2 periods after their formation. The di
variables follow the ocean anomalies along their meridional course.

However, the development of a positive SST anomaly at the western bou
coincident with a minimum of the subtropical gyre intensity, because at t
streamfunction anomaly is of positive sign south of the zero wind stress c



Anomalies of ocean temperature (blue:-2°C, red:+2°C) and streamfunctio
Sverdrup, solid≥0, dashed<0) every year over half an oscillation period.

The oscillation period

In this experiment, the oscillation period $T = 12.5$ yr is easily related to t
baroclinic Rossby waves across the basin:

$$\frac{L_X}{\beta R^2} = \frac{8000\,\mathrm{km}}{1.6 \times 10^{-11}\,\mathrm{m}^{-1}\mathrm{s}^{-1}\,(35\,\mathrm{km})^2} = 12.9\,\mathrm{yr}$$

However Cessi (2000) shows that it is not that simple since the advection
well. Indeed the same experiment with a slightly different boundary cond
generate oscillations with a period of 19.5 yr.

Comparison of experiments with various parameters and boundary condit
determine exactly how the oscillation period is established.

The oscillation mechanism

Response of the wind stress τ' to temperature anomalies θ'

For large scale perturbations ($l \gg L_\rho$ and $l \gg L_d$, $L_d = (d_e k_s/\gamma)^{1/2} \sim 55$

$$\tau' \propto -\frac{\rho_a f d_e k_s}{d\ S} \partial_y \theta'$$

For small scale harmonic perturbations ($l \ll L_\rho$ and $l \ll L_d$):

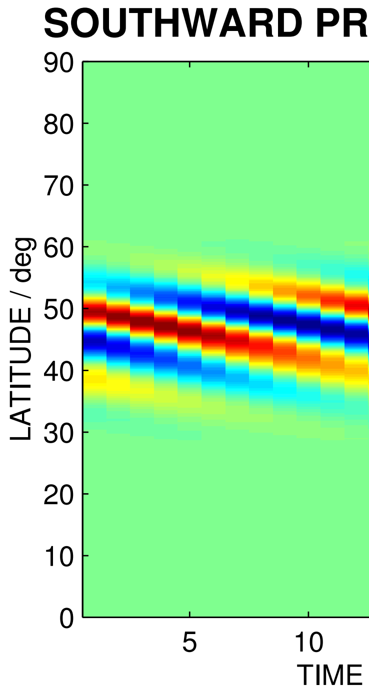
$$\tau' \propto \frac{\gamma \rho_a f L_\rho^2}{d\ S} \partial_y \theta' \propto \frac{\gamma g \rho_a d_e}{f \Theta} \partial_y \theta$$

Note the opposite sign, but rather similar coefficient in amplitude (2.2 ×

Propagation of temperature anomalies
Southward propagation of ocean temperature anomalies from 60°N to 30°N follows a passive advection by the mean flow, with a mean velocity of 2.6×10^{-3} m s^{-1}.

Remaining questions:

- what sets the meridional wavenumber?
- what is exactly the instability mechanism?



Discussion

Differences with Cessi (2000)

- boundary condition for wind speed at $y = 0$ and $y = L_y$: $\partial_y \tau = 0$ (inste
- several resolutions and parameters have been experimented and compar
previously used seems necessary
- resolution and numerics do influence the occurence of oscillations
- a numerical experiment with isotropic viscosity is now running ($A_x = A$

The paradox

It appears that the parameter range leading to oscillations is rather limit
coefficients $O(200\ \mathrm{m}^2\ \mathrm{s}^{-1})$ that are smaller than estimations from observa
s^{-1}), such that there is finally some incoherence with the choice of the dy
sophisticated ocean model should then be used. Following Cessi (2000), it
attribute the variability to coupled processes or to simple oceanic turbule
clear criteria (like the phase relationship in poleward heat transport chan
ocean). This is the challenge of climatic data analysis, so this might just

Conclusion

- Simplest setting for studying interactions between ocean gyres and zona
- Potentially interesting mechanism for decadal oscillations observed in th
- Further developments include the use of an ocean models that allow act
temperature anomalies

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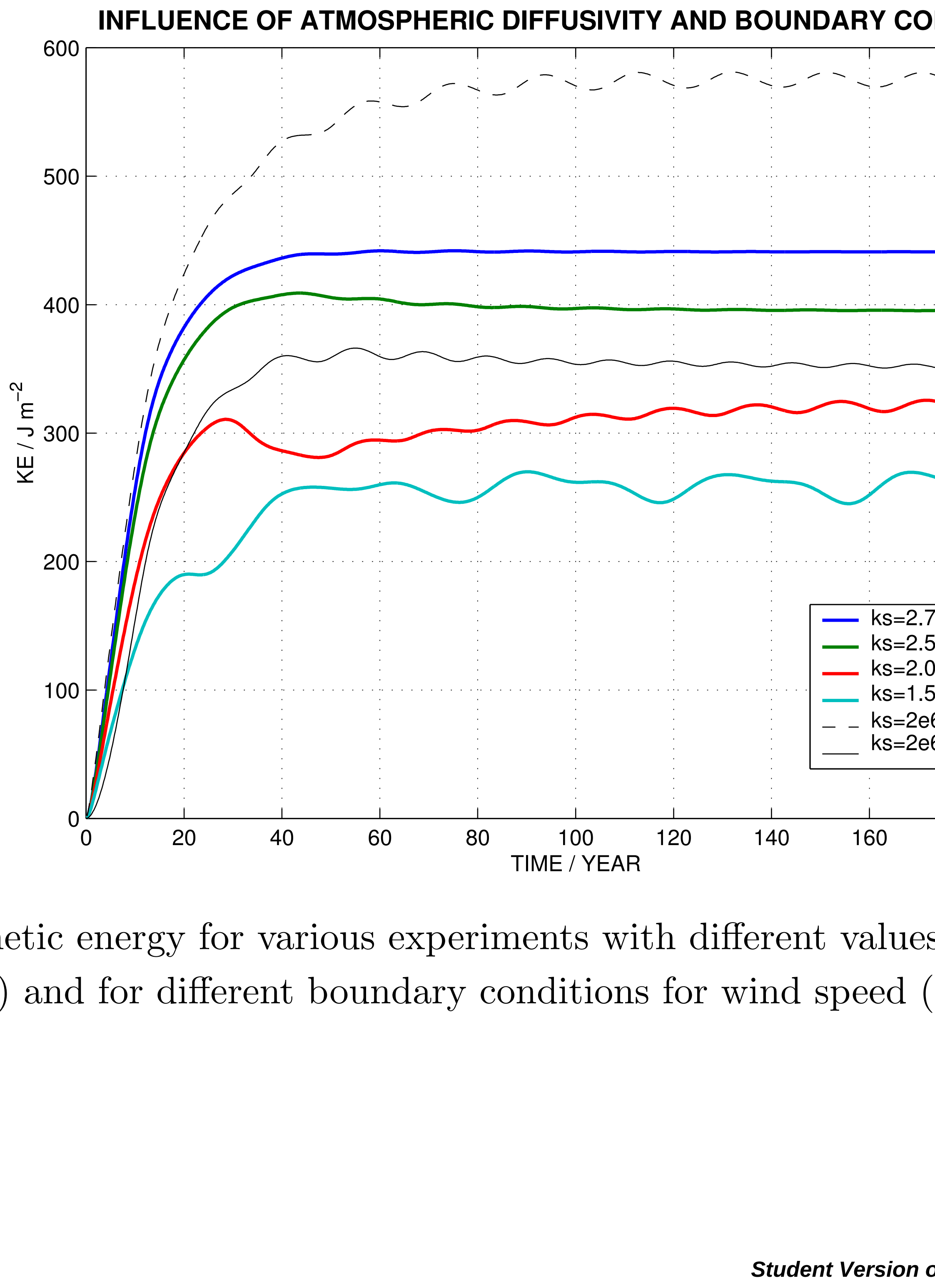
pas [s]	time [yr]	period [yr]	KE_{min} [J m^{-2}]	KE_{max} [J m^{-2}]	SST_{min} [°C]	SST_{max} [°C]
8640	640	17.9	454	465	-51.28	121.18
8640	500	18.4	539	550	-40.29	91.49
7200	280	18.9	553	565	-34.41	77.16
3800	210	19.2	563	574	-30.74	68.49
2000	202	19.5	569	581	-28.01	62.82
8640	620	EQ.	517	517	-40.86	114.53
8640	668	EQ.	557	557	-31.72	89.43
7200	270	EQ.	573	573	-27.15	75.48
3800	916	EQ.	583	583	-24.18	67.02
1920	200	EQ.	590	591	-22.13	61.54
1080	200	EQ.	605	605	-18.04	51.33

periments summary: $A_x = 2000$ m^2 s^{-1}, $A_y = K = 200$ m^2 s^{-1}, $\tau = 0$
s in Cessi (2000). Note the influence of resolution on the extrema in

iption	**Typical value**
s parameter	10^{-4} s^{-1}
ary β-effect	1.6×10^{-11} m^{-1} s^{-1}
heat for air	1000 J K^{-1} kg^{-1}
heat for sea warter	4000 J K^{-1} kg^{-1}
zonal extent	$0.825\times10^{7}m$
meridional extent	1.0×10^{7} m
pheric height	8×10^{3} m
thermocline depth	10^{3} m
air density	1.25 kg m^{-3}
ter density	1000 kg m^{-3}
atmosphere exchange coefficient	23 W m^{-2} K^{-1}
drag coefficient	2.4×10^{-2} m s^{-1}
surface gravity	9.8 m s^{-2}
nesq temperature	273 K
pheric stratification	5×10^{-3} K m^{-1}
raction of atmosphere above ocean	0.3
ve radiation (LW) for $\theta=0°C$	200 W m^{-2}
$/\partial\theta$	2.475 W m^{-2} °C^{-1}
pheric surface baroclinic eddy diffusivity	$2-3\times10^{6}$ m^{2} s^{-1}
eddy-diffusivity	$200-2000$ m^{2} s^{-1}
eddy-viscosity	$200-2000$ m^{2} s^{-1}

Table 2: Parameters used in the coupled model



Time series of kinetic energy for various experiments with different values
diffusivity (colors) and for different boundary conditions for wind speed (