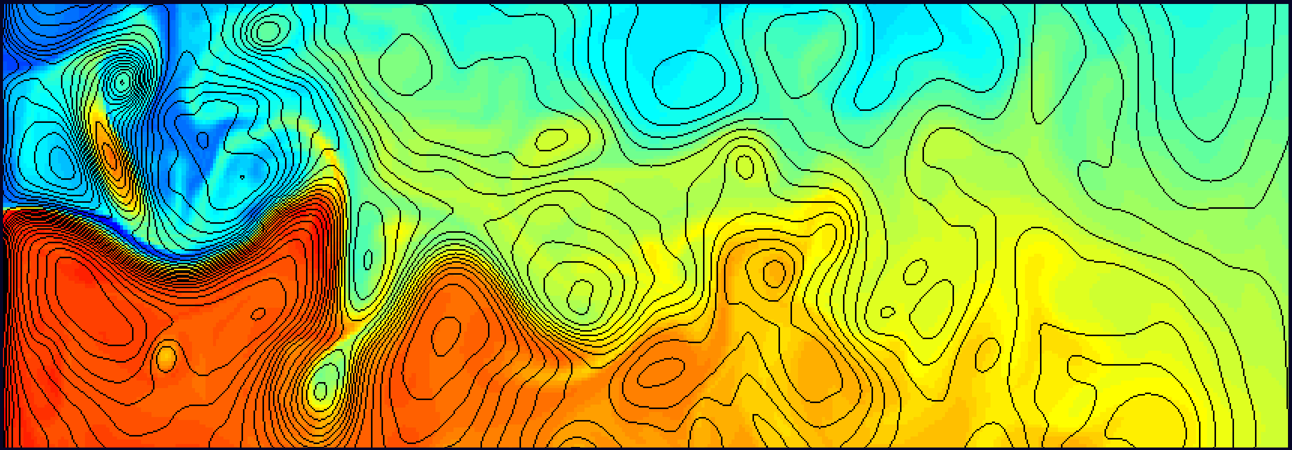
The Quasi-Geostrophic Coupled Model (Q-GCM): Studying Ocean-Atmosphere Interaction With An Eddying Ocean



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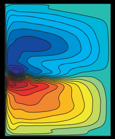
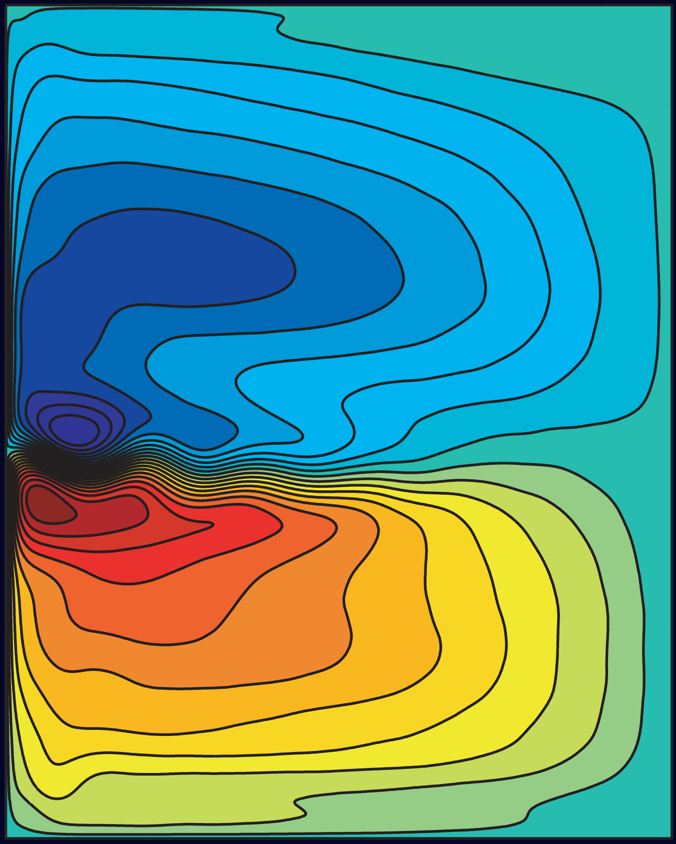


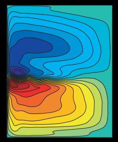
1. Motivation

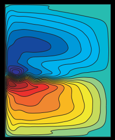
Numerical studies of wind-driven ocean gyres demonstrate tendency for spontaneous midlatitude variability:

- Inclusion of the mesoscale is essential in these models to produce the variability: thus it may not be captured by GCMs;
- Low frequency (decadal) variability has been observed;
- Variability in circulation forces changes in meridional heat flux, and may thereby affect climate.



Hypothesis: Intrinsic ocean variability contributes to midlatitude climate variability.





2. Model Development

Q-GCM: General

Quasigeostrophic dynamics:

- Includes non-linear effects;
- Includes planetary waves;
- Can be solved efficiently even at high resolution;
- In the standard implementation there is poor representation of vertical processes (especially heat transport);
- Usually has constant temperature layers, in which variations in thickness act as a proxy for temperature variations.

A QG coupled model will need to include:

1. Wind stress τ as a drag on the atmosphere, and driver of the ocean;
2. Exchange of heat between the ocean and atmosphere;
3. Atmosphere driven by prescribed incoming solar radiation.

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Q-GCM: Inclusion of wind stress

Define an atmospheric mixed layer of thickness ${}^aH_m$. It will be embedded within the lower layer of the atmosphere.

$$ {}^a v_m = {}^a v_1 + \frac{{}^a\tau^x}{{}^aH_m f_0} $$

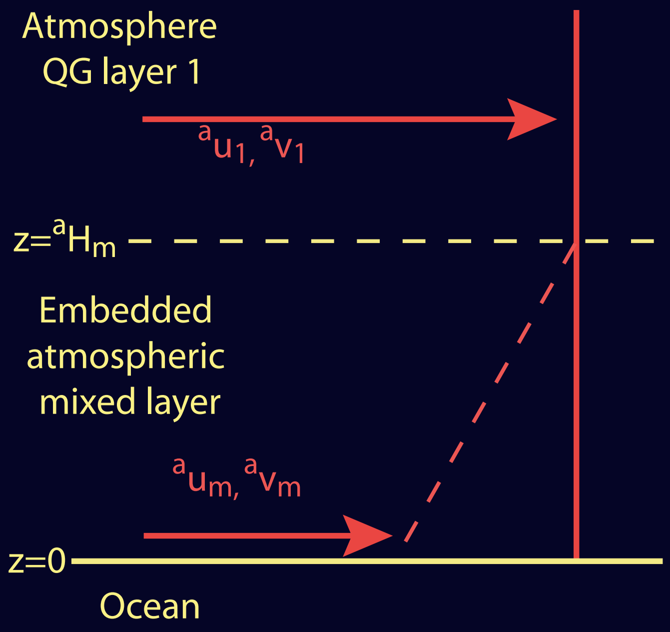
$$ {}^a u_m = {}^a u_1 - \frac{{}^a\tau^y}{{}^aH_m f_0} $$

and quadratic stress,

$$ \left({}^a\tau^x, {}^a\tau^y\right) = C_D \left|\mathbf{{}^a u_m}\right| \left({}^a u_m, {}^a v_m\right), $$

allows us to solve for stress as a function of ${}^a u_1$ and ${}^a v_1$. Ekman pumping is

$$ {}^a w_{ek} = \frac{{}^a\tau^y_x - {}^a\tau^x_y}{f_0}, $$



In the ocean, $\quad {}^o\tau = \frac{{}^a\rho}{{}^o\rho}{}^a\tau$

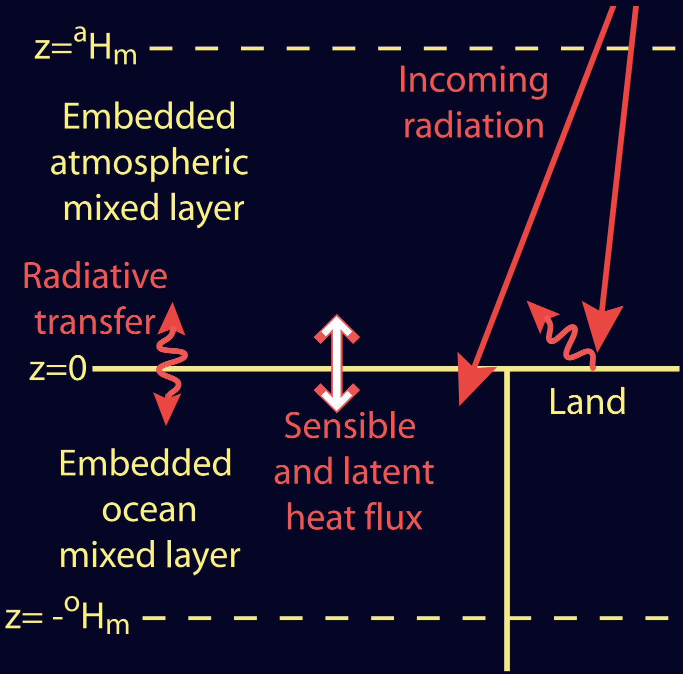
$$ {}^o w_{ek} = \frac{{}^o\tau^y_x - {}^o\tau^x_y}{f_0} $$

Q-GCM: Exchange of heat between ocean and atmosphere

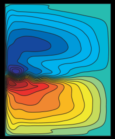
Mixed layers are also used to communicate heat. Three components of heat at the ocean–atmosphere interface:

1. Radiative transfer;
2. Sensible and latent heat flux;
3. Incoming solar radiation.

This requires knowledge of mixed layer temperature, for which we need an evolution equation:



$$ {}^oT_{mt} + ({}^ou_m{}^oT_m)_x + ({}^ov_m{}^oT_m)_y - \frac{{}^ow_{ek}{}^oT_m}{{}^oH_m} = {}^oK_H\nabla_H^2\,{}^oT_m + \frac{-{}^oF_0 + {}^oF_m}{{}^o\rho{}^oC_p{}^oH_m} $$



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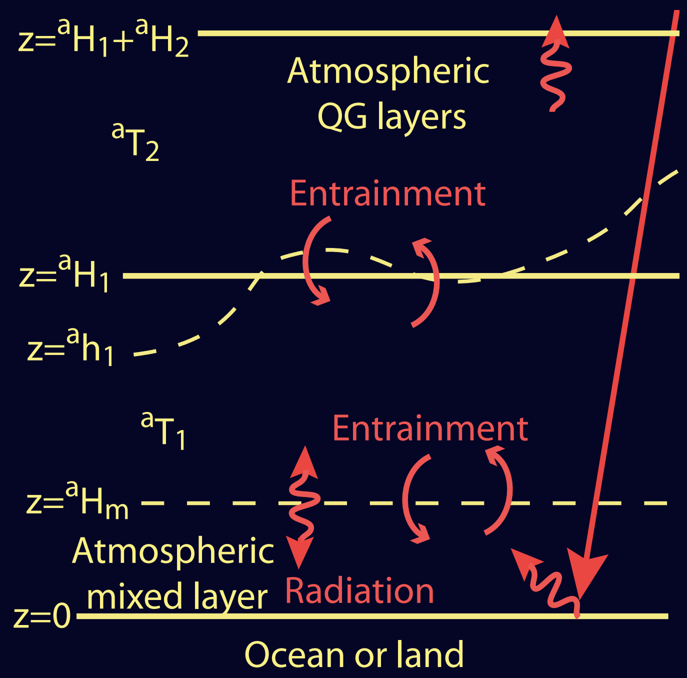
Q-GCM: Atmosphere driven by radiation

Since the atmospheric mixed layer receives the driving radiation, we must allow this to be communicated to the constant temperature QG layers (our 'troposphere').
This heat can only be transferred by altering the relative thicknesses of the two QG layers and is achieved by an entrainment

$$ {}^a e_1 = \frac{\text{Net Forcing}}{{}^a\rho\, {}^aC_p \Delta_1^a T} $$

Entrainment is also employed to handle the vertical flux of heat in the ocean which is produced from Ekman pumping.



Q-GCM: Equations

The mixed layers and heat transport require a modification to the QG equations. For example, in the upper ocean layer we define potential vorticity

$$^{o}q_1 = \frac{\nabla_H^2 \, {}^{o}p_1}{f_0} + \beta y + \frac{f_0 \, {}^{o}\eta_1}{{}^{o}H_1}$$

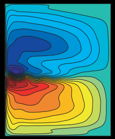
where $\quad {}^{o}\eta_1 = -\frac{{}^{o}p_1 - {}^{o}p_2}{{}^{o}g'} = {}^{o}H_1 - {}^{o}h_1$. The evolution equation is then

$$^{o}q_{1t} + ({}^{o}u_1 \, {}^{o}q_1)_x + ({}^{o}v_1 \, {}^{o}q_1)_y = \frac{f_0}{{}^{o}H_1}({}^{o}w_{ek} - {}^{o}e_1) - \frac{{}^{o}A_H}{f_0}\nabla_H^6 \, {}^{o}p_1$$

Compare this with the usual QG equation

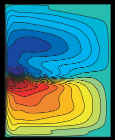
$$q_t + (uq)_x + (vq)_y = \frac{\nabla \times \tau}{\rho H} + \nu \nabla^4 p.$$

The only real difference here is the inclusion of the entrainment term in Q-GCM.



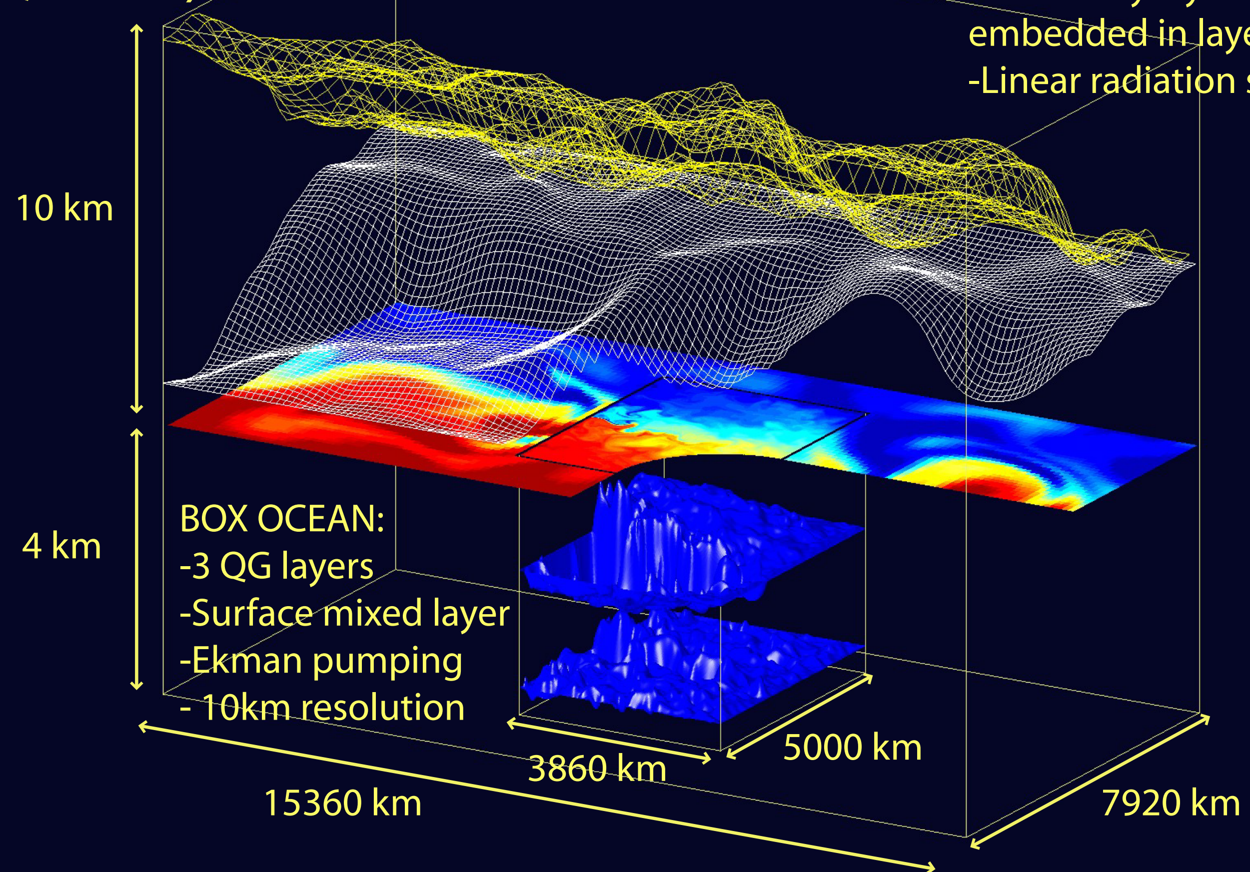
Q-GCM: Other features

- Convection allowed in both the atmosphere and ocean.
- Heat fluxes are linearised about a stationary mean state.
- Atmospheric mixed layer thickness is variable, but the ocean mixed layer thickness is held constant. (Explanation of this is deferred until later.)
- Ekman drag at the bottom of the ocean is included.
- Boundaries are partial slip – usually we run in a configuration close to no slip.
- Atmosphere is 8000km in north-south extent – larger than most plausible β-planes. (This is because the vertical sidewalls of the atmosphere acted to stabilise baroclinic instability.)
- We use hyper-viscosity in both of the QG layers.
- 66 CPU hours on a linux (1600MHz) single processor for one century of integration.

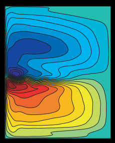


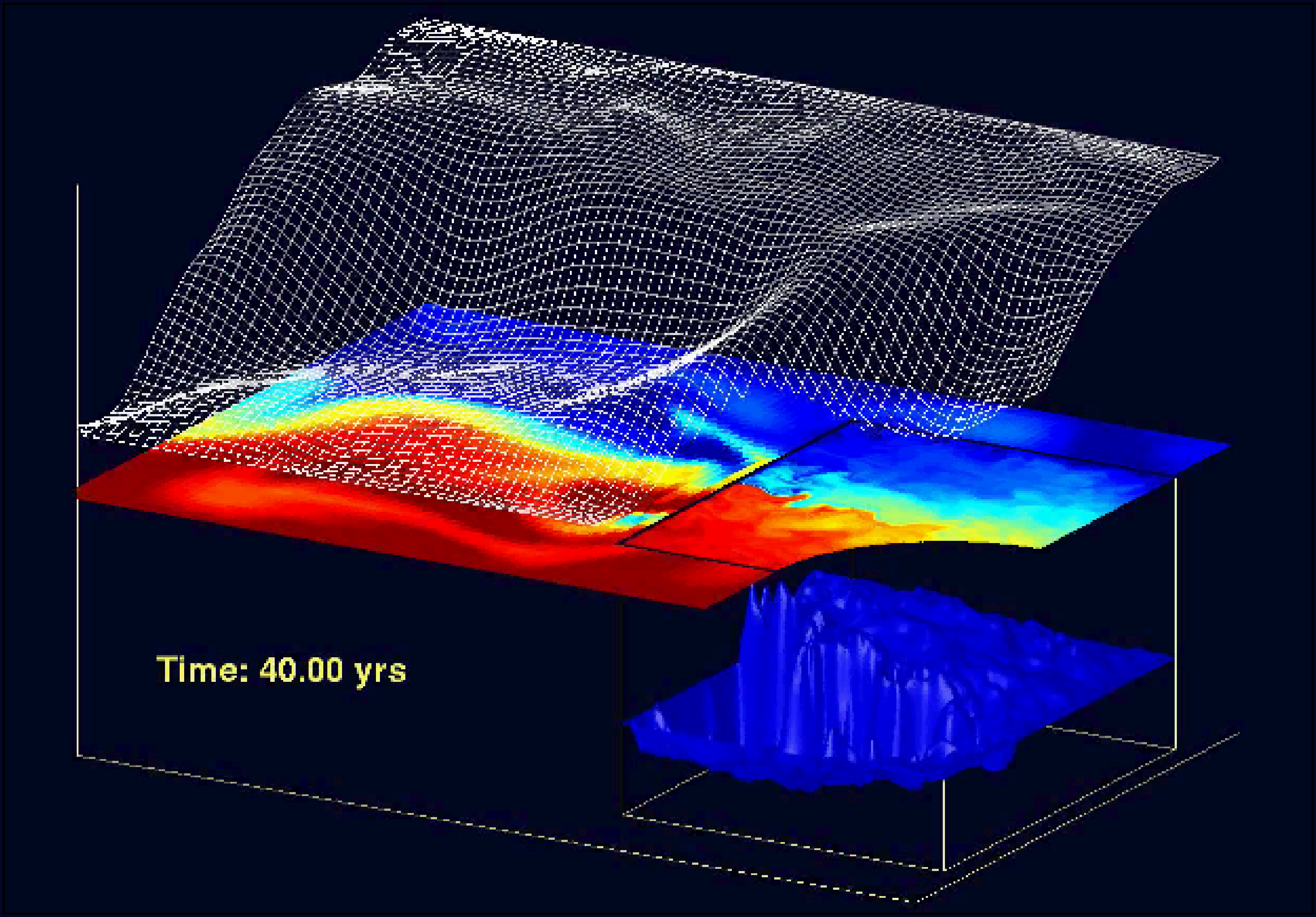
Quasi-Geostrophic Coupled Model (Q-GCM)

CHANNEL ATMOSPHERE :
-3 QG layers
- Boundary layer embedded in layer 1
-Linear radiation scheme



BOX OCEAN:
-3 QG layers
-Surface mixed layer
-Ekman pumping
- 10km resolution

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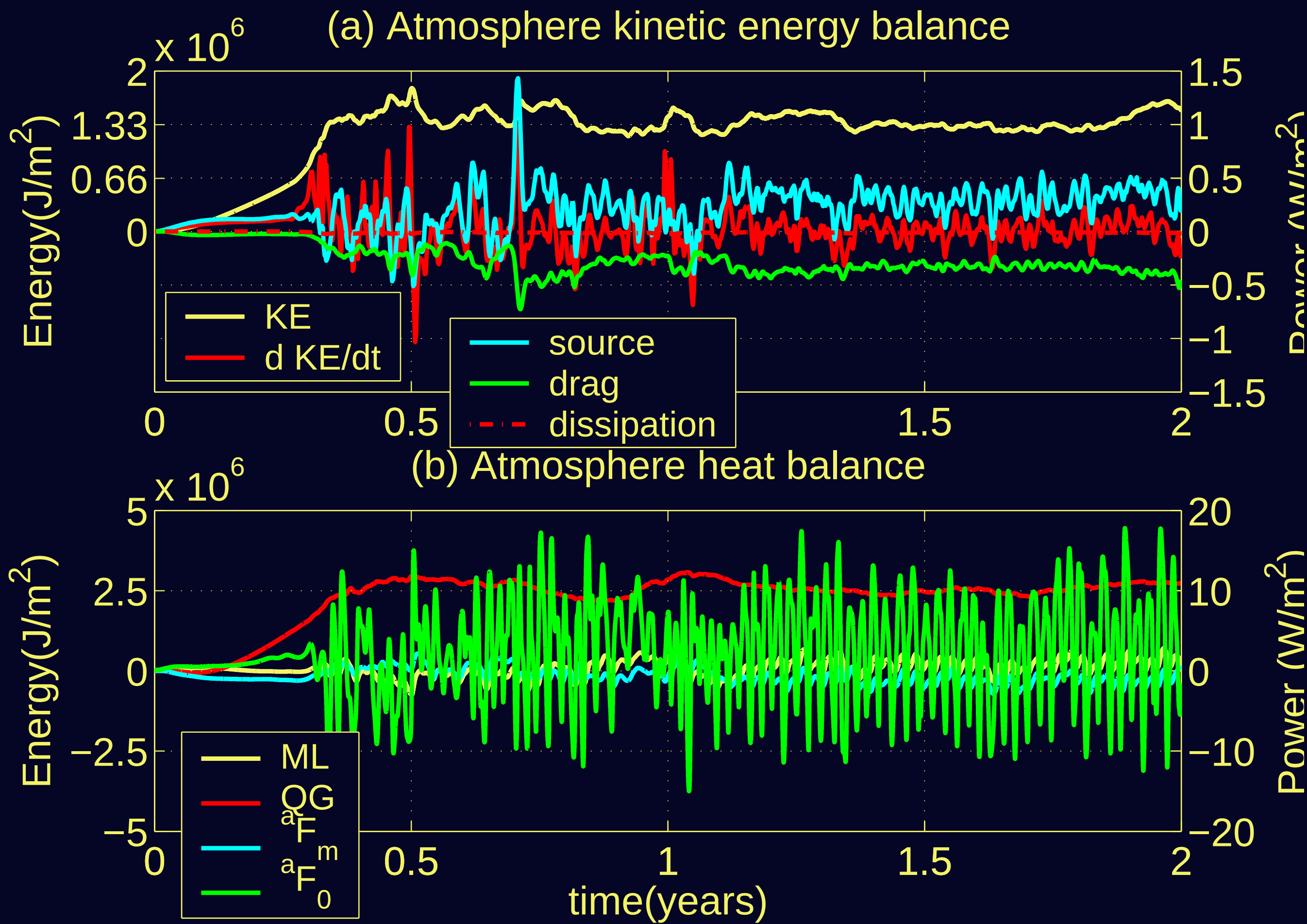
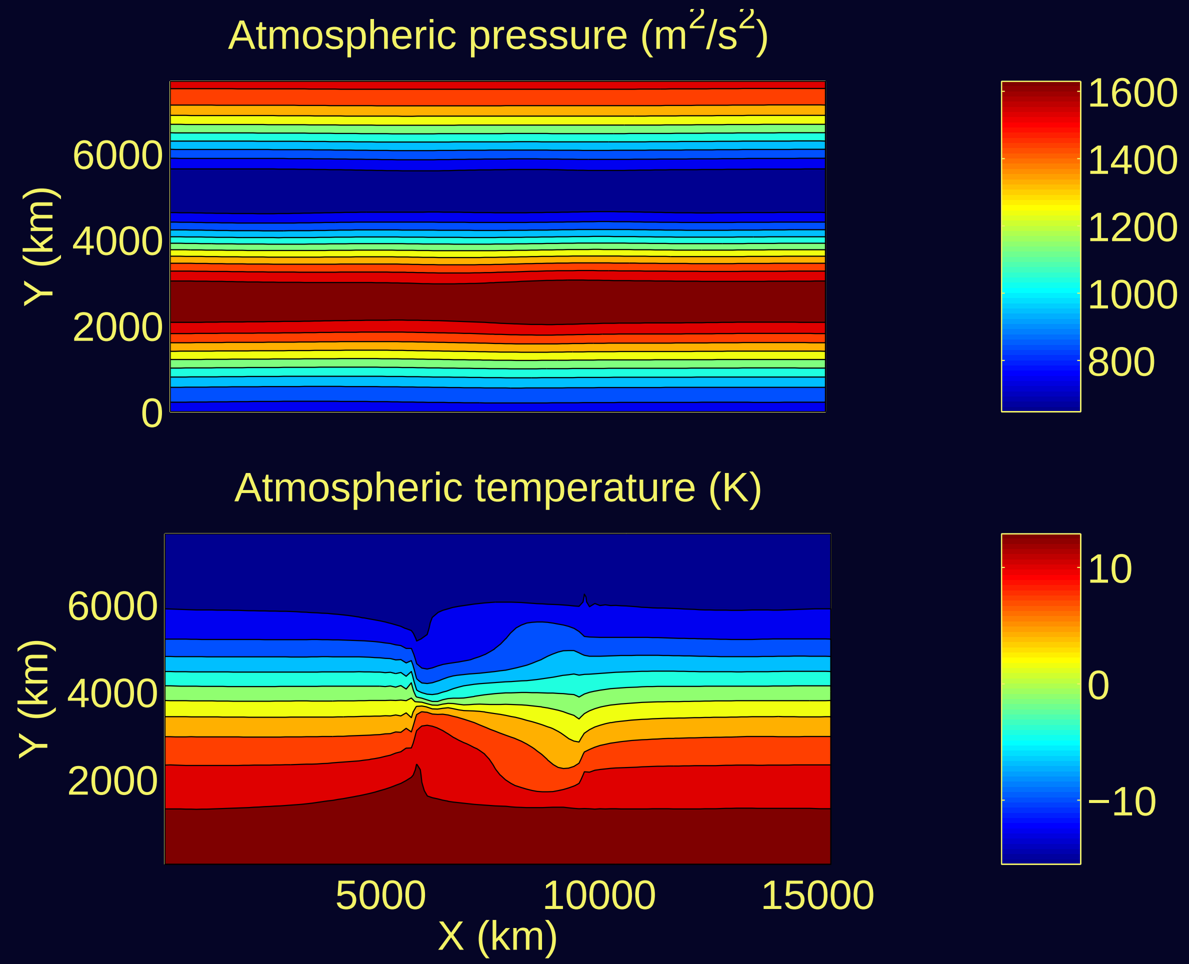
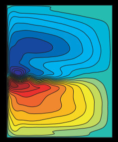
3. Results: Spinup

Spinup – energy balance in the atmosphere



Spinup – atmosphere mean state





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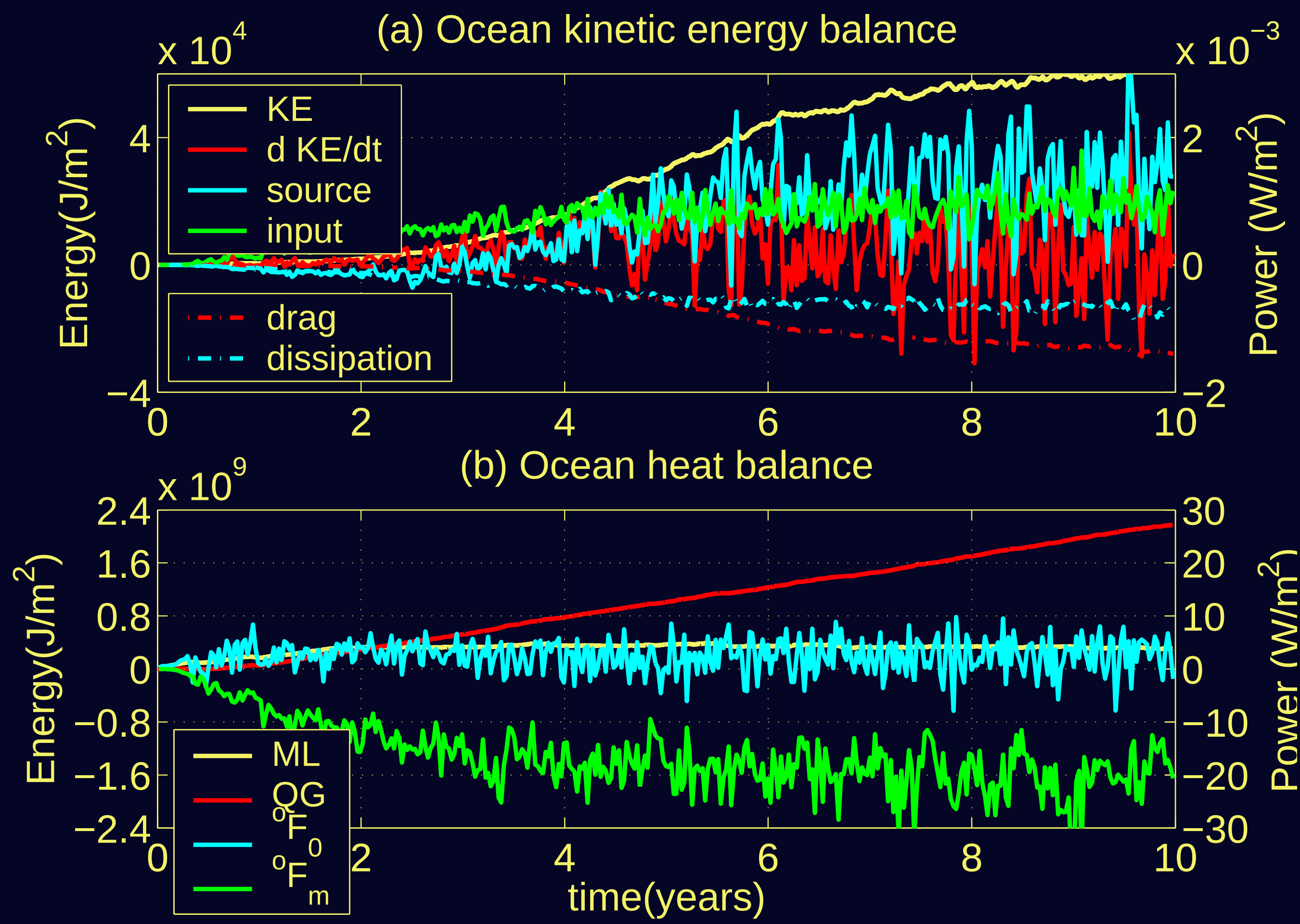
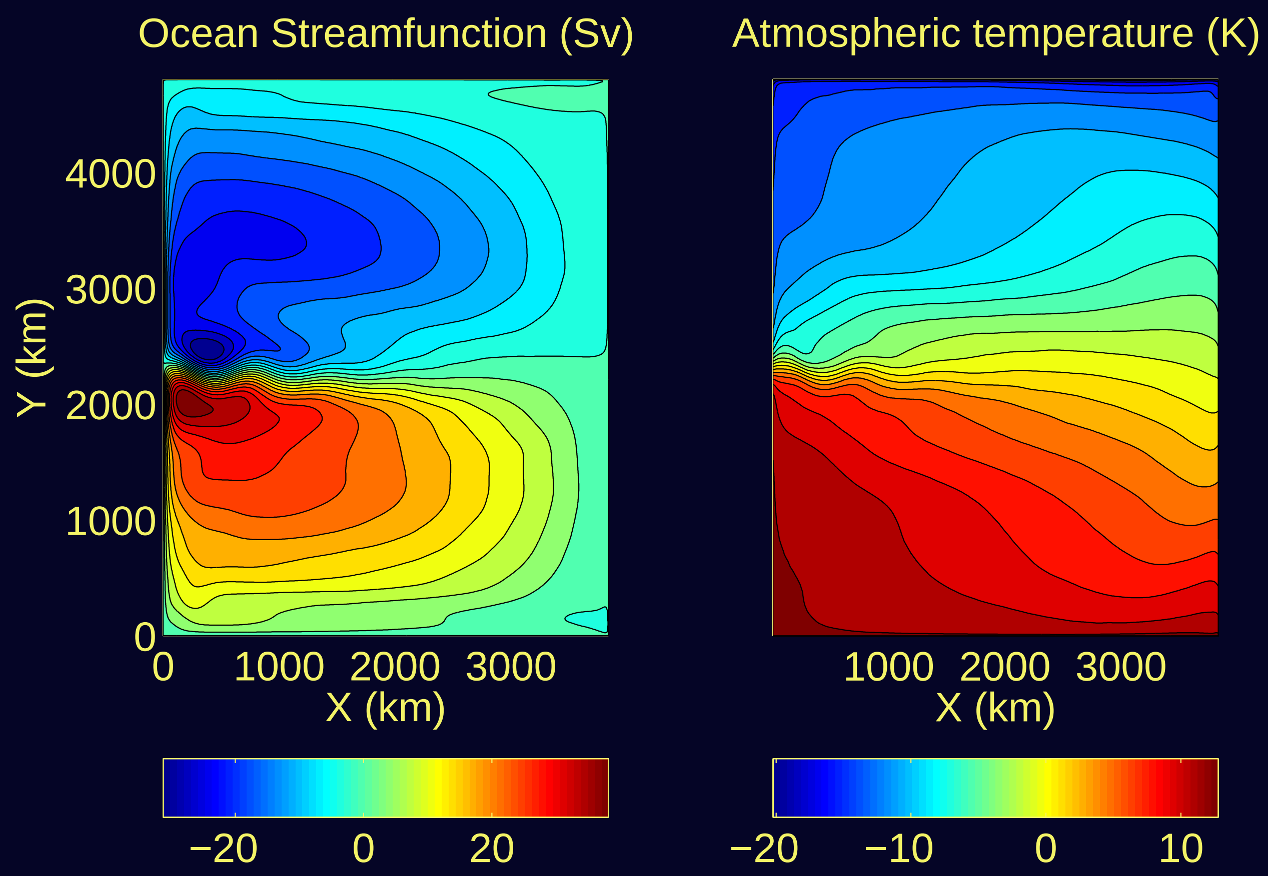
Spinup – energy balance in the ocean



Spinup – ocean mean state



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4. Results: Low Frequency Variability

OCEAN DATA

↓

Filter (2 year, low-pass Fourier filter)

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Empirical Orthogonal Function analysis

ATMOSPHERE DATA

↓

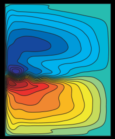
Filter (2 year, low-pass Fourier filter)

↓

Empirical Orthogonal Function analysis

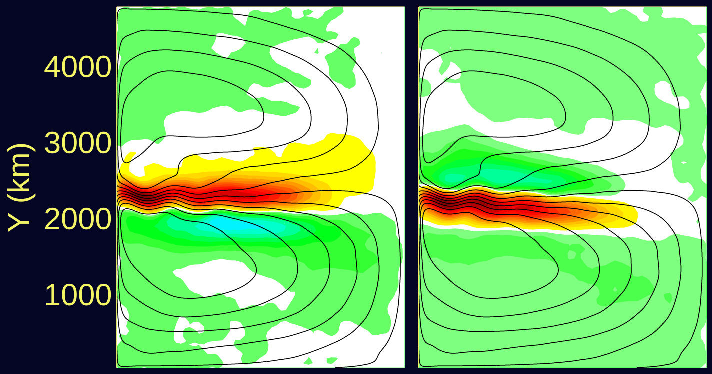
↘ ↙

Canonical Correlation Analysis

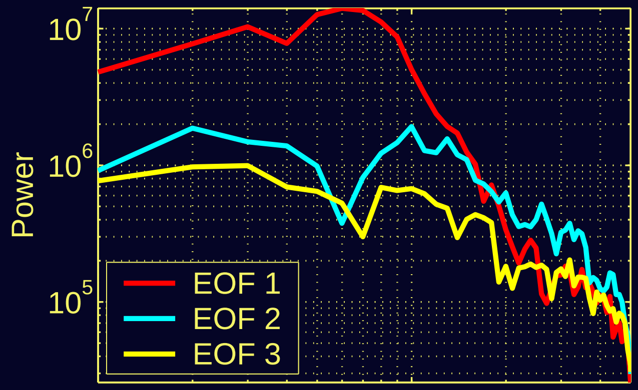


Ocean Low-Frequency Variability

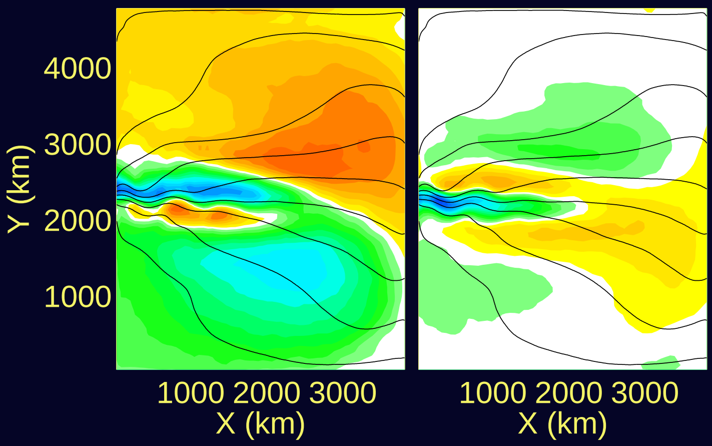
1st Hilbert EOF –– Ocean Interface Height –– 43 % of variance



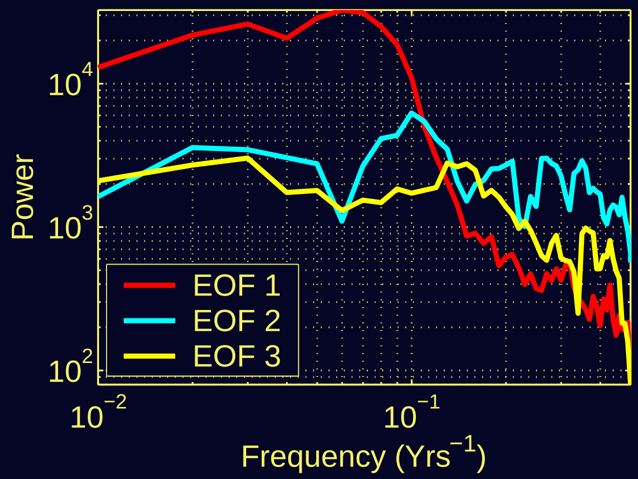
Spectra of Hilbert PCs



1st Hilbert EOF –– SST –– 37 % of variance

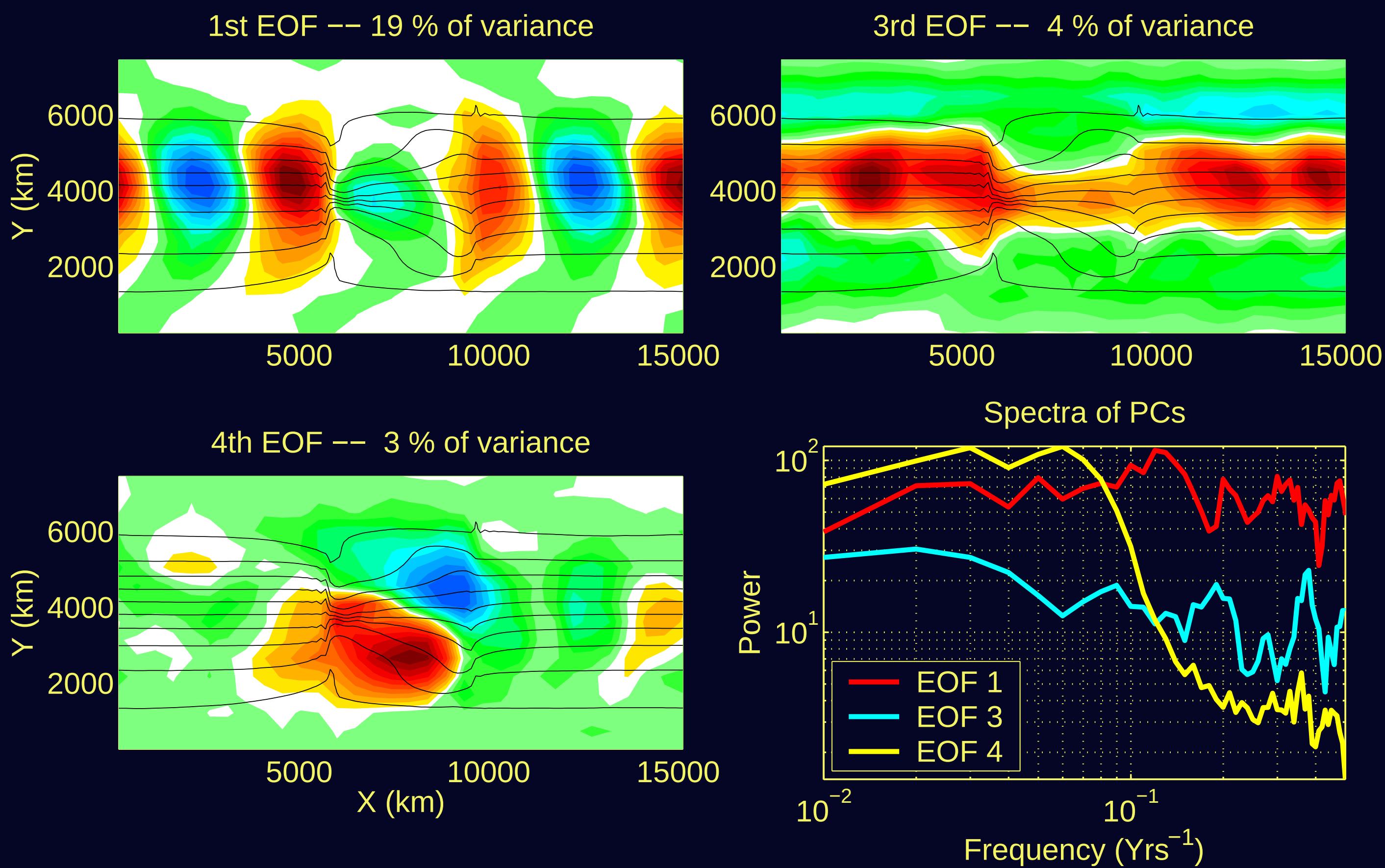


Spectra of Hilbert PCs



Atmosphere Low-Frequency Variability

Atmospheric Boundary Layer Temperature EOFs



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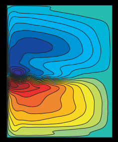
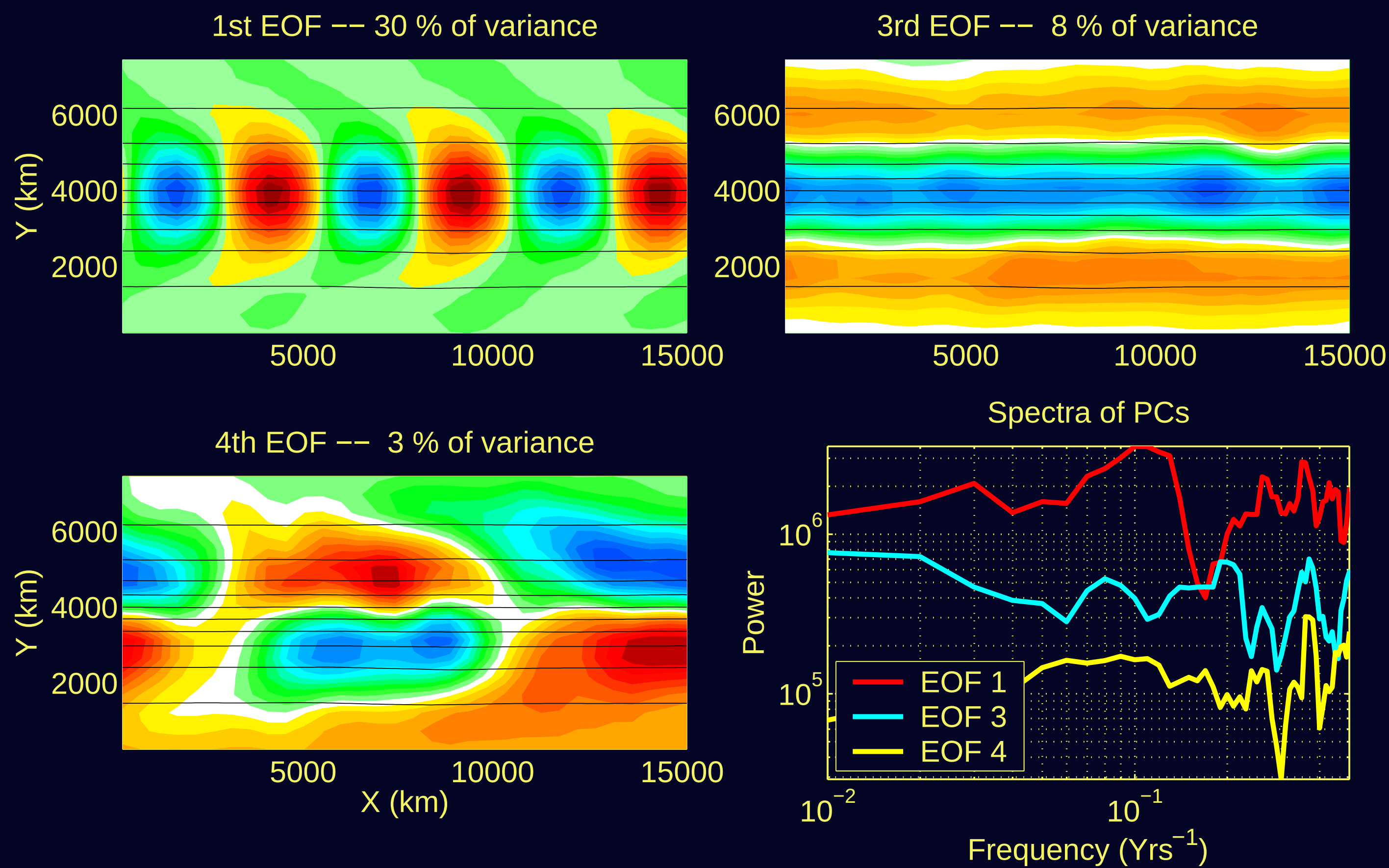
Atmosphere Low-Frequency Variability



Atmospheric Interface Height EOFs



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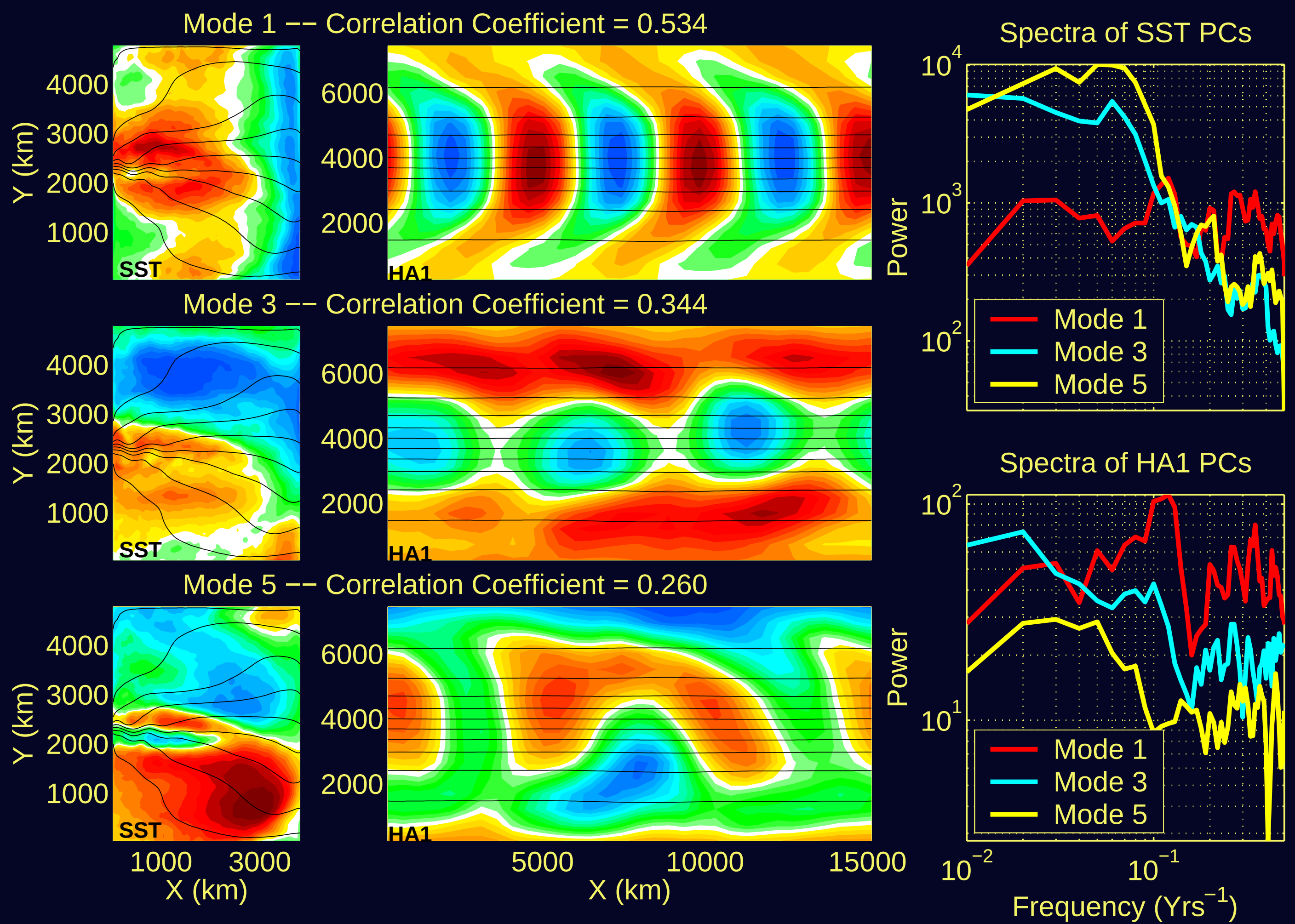
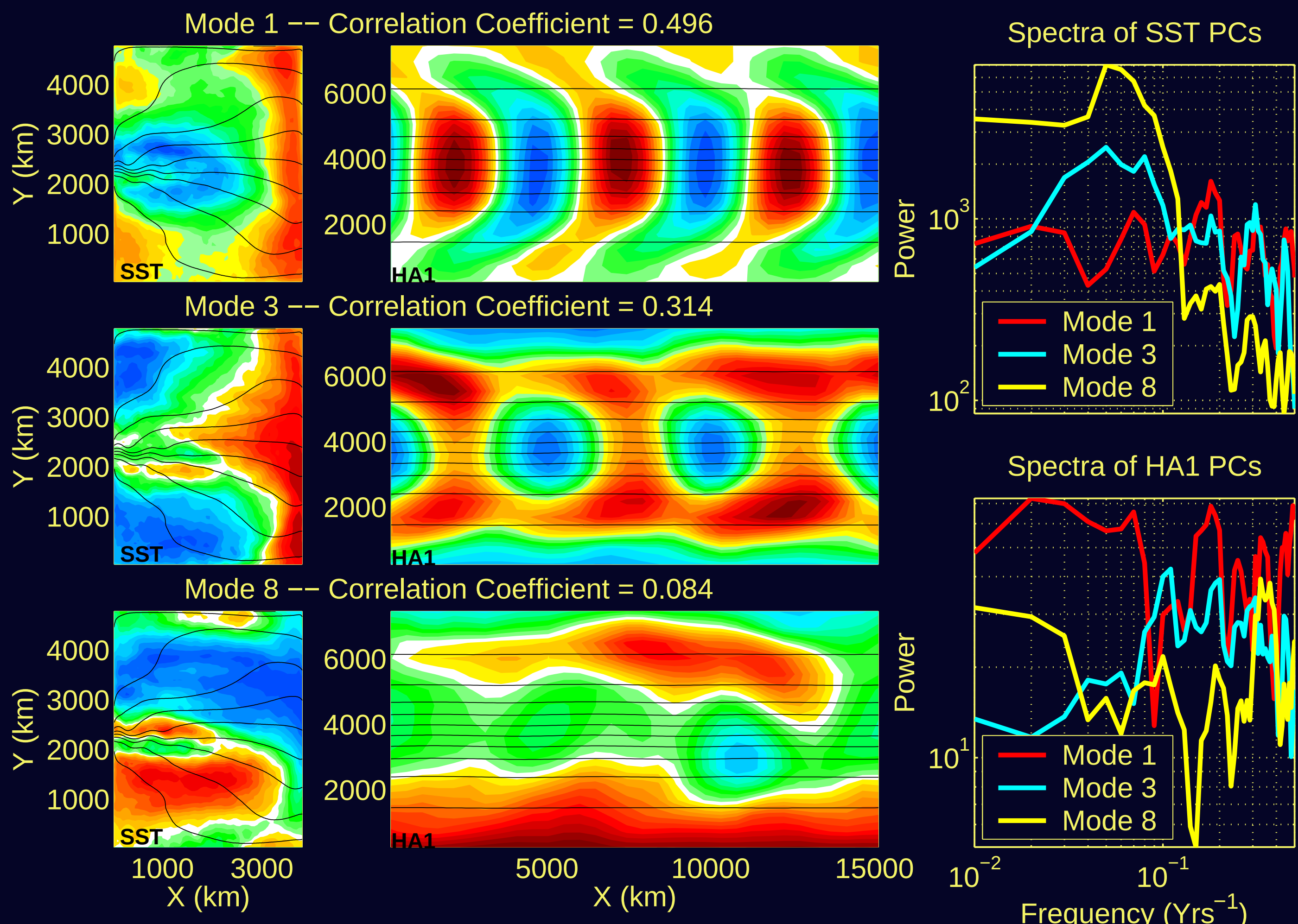
5. Results: Coupled Variability

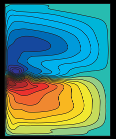
Fully coupled experiment



"Partially" coupled experiment



In this experiment, the atmosphere drives ocean circulation as usual, but there is no feedback from SST to atmosphere.



6. Conclusions

- When ocean eddies are resolved, intrinsic ocean low-frequency variability emerges;
- In our idealised model, the ocean can enhance atmospheric low-frequency variability. This result is not seen in comprehensive coupled climate models, which do not resolve ocean eddies;
- A coupled mode is generated by feedback between the atmospheric wavenumber-3 pattern and Rossby wave propagation in the ocean, producing a decadal peak in the primary atmospheric mode.
- In the future Q-GCM will be altered to simulate Southern Hemisphere midlatitude ocean-atmosphere dynamics.

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