A conceptual model for studying century-to-millennia timescale processes in the coupled ocean-atmosphere system

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The conceptual model couples a moist energy balance model to an advective-diffusive model of the ocean's global overturning circulation through a surface ocean mixed layer with sea ice. This enables us to explicitly represent heat and freshwater surface fluxes and evolve temperature and salinity in the ocean interior. A detailed description of the model can be found to the right.

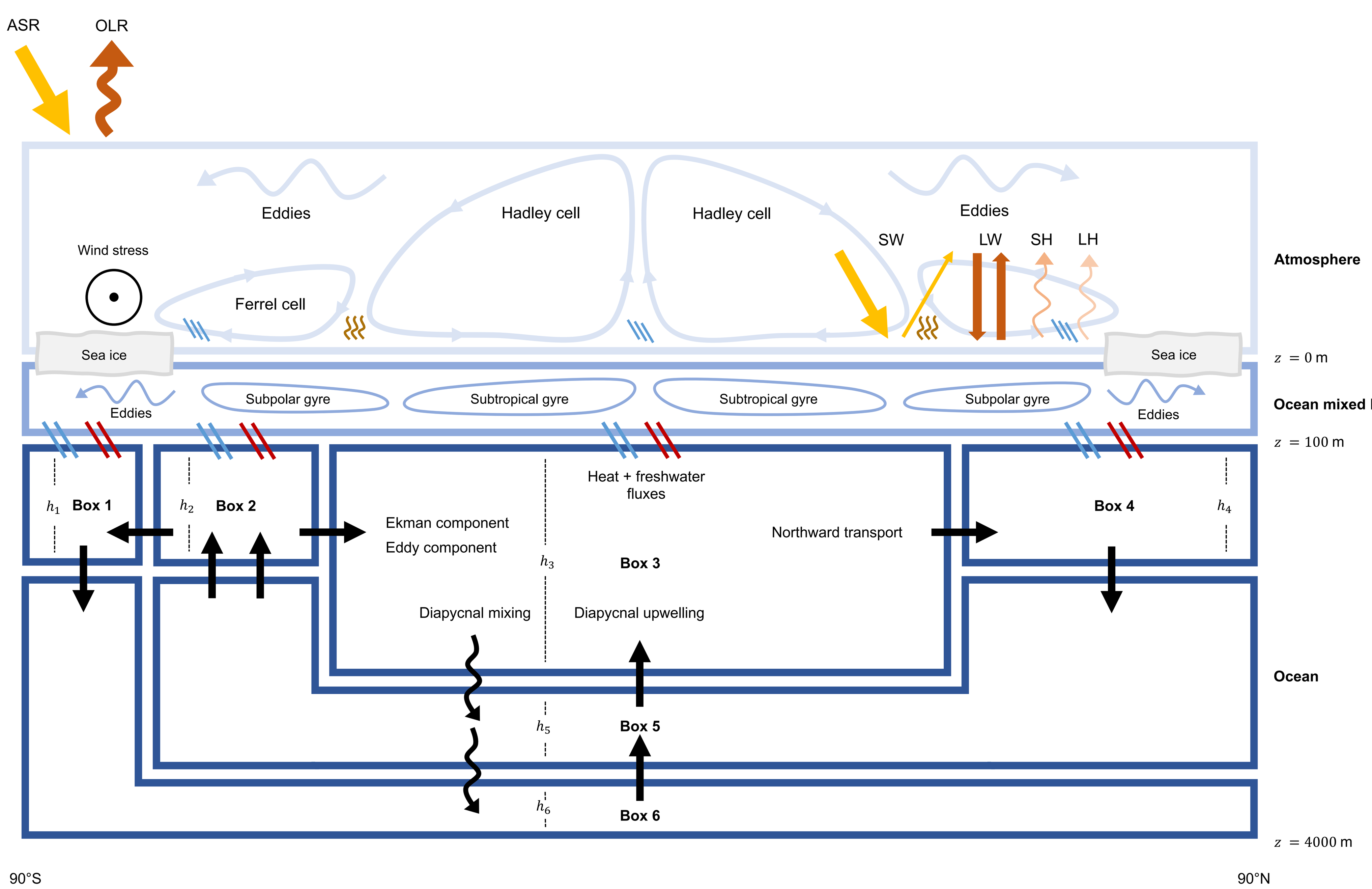

Figure 1. Schematic of the conceptual model. The model has a prescribed zonal-mean profile of incoming shortwave radiation that is either fixed or can vary in time. In the current version, we also prescribe a zonal-mean wind stress profile, but this can also be prognostic.

Model description

Atmosphere

The atmosphere is represented by a single layer with temperature $T_a(x,t)$, which evolves according to the net energy flux into the atmospheric column at each sine of latitude x:

$$C_a\frac{\partial T_a}{\partial t} = \eta S - F_{\mathrm{OLR}} + F_{\mathrm{UP}} - F_{\mathrm{DN}} + F_{\mathrm{SH}} + F_{\mathrm{LH}} - \nabla\cdot F_a$$

The top-of-atmosphere and surface radiative longwave fluxes are bulk representations and parameterized as either linear responses on T_a or using bulk aerodynamical formulas. The divergence of the northward column-integrated atmospheric energy transport is represented as Fickian diffusion of near-surface moist-static energy:

$$\nabla\cdot F_a = -\frac{p_s}{a^2 g}D_a\frac{\partial}{\partial x}\left[(1-x^2)\frac{\partial h}{\partial x}\right] \longrightarrow h = c_pT_a + L_v q_a$$

Parameterization of the Hadley and Ferrell cells

To correctly simulate the partitioning between latent and dry-static energy transport, we use weighting function that partitions atmospheric heat transport into a component due to the mean meridional circulation (F_{MMC}) and a component due to eddies (F_{EDDY}):

$$F_{\mathrm{MMC}} = wF_a \quad \text{and} \quad F_{\mathrm{EDDY}} = (1-w)\,F_a$$

The strength of the moisture flux by the mean meridional circulations is determined by ψ which is found by approximating gross moist stability H as

$$H \approx \varsigma\cdot h(0) - h \longrightarrow F_{\mathrm{MMC}} = \psi H$$

The moisture flux by the Hadley and Ferrel cell, and the moisture flux by eddies is:

$$F_{L,\mathrm{MMC}} = -\psi L_v q_a \qquad F_{L,\mathrm{EDDY}} = \frac{2\pi p_s}{g}(1-w)\,L_v D_a\left(1-x^2\right)\frac{\partial q_a}{\partial x}$$

$P-E$ can be found by taking the divergence of the sum of the two components, which gives:

$$P - E = -\nabla\cdot F_L(x) = -\frac{1}{2\pi a^2}\frac{dF_L}{dx}$$

Ocean mixed layer and sea ice

The ocean mixed layer is also represented by a single layer with temperature $T_m(x,t)$ and salinity $S_m(x,t)$. The evolution equations take the form:

$$\rho_o c_o h_m\frac{\partial T_m}{\partial t} = \left(1-\eta\right)(1-\alpha)\,S - F_{\mathrm{UP}} + F_{\mathrm{DN}} - F_{\mathrm{SH}} - F_{\mathrm{LH}} + \gamma\left(T_m - \tilde{T}_o\right) - \nabla\cdot F_{\mathrm{GYRE}} - \nabla\cdot F_{\mathrm{EDDY}}$$

$$\rho_o h_m\frac{\partial S_m}{\partial t} = S_0\cdot\left(P - E - \rho_i\frac{\partial H_i}{\partial t}\right) + \gamma\left(S_m - \tilde{S}_o\right) - \nabla\cdot S_{\mathrm{GYRE}} - \nabla\cdot S_{\mathrm{EDDY}}$$

The mixed layer depends on the surface heat or freshwater fluxes and heat or freshwater transport by eddies or ocean gyres. The mixed layer also interacts with the ocean overturning circulation. Sea ice is represented by a simple thermodynamic layer.

When T_m is below the freezing point (271.15 K), sea ice grows and when T_m is above the freezing point, sea ice melts. The sea ice evolution equation takes the form:

$$-L_f\frac{\partial H_i}{\partial t} = \left(1-\eta\right)(1-\alpha)\,S + F_{\mathrm{UP}} + F_{\mathrm{DN}} - F_{\mathrm{SH}} - F_{\mathrm{LH}} + \gamma\left(T_m - \tilde{T}_o\right)$$

There is an additional ice flux that alters the surface temperature T_s when the region is ice covered.

Ocean overturning

The ocean overturning model is represented by a series of boxes that exchange temperature and salinity, and can adjust their volume according to closures for the overturning circulation:

$$A_3\frac{\partial h_3}{\partial t} = q_{23} + q_{53} - q_{34} \qquad A_3\frac{\partial h_5}{\partial t} = q_{65} + q_{12} - q_{23} - q_{53} + q_{34} \qquad A_3\frac{\partial h_6}{\partial t} = -q_{12} - q_{65}$$

Southern Ocean transport is represented as the balance between Ekman and eddy transport:

$$q_{12} = q_{23} = q_{\mathrm{EKMAN}} - q_{\mathrm{EDDY}} \qquad q_{\mathrm{EKMAN}} = \frac{\tau_{\mathrm{SO}}L_x}{\rho_0 f_{\mathrm{SO}}} \qquad q_{\mathrm{EDDY}} = \kappa_{\mathrm{GM}}L_x\frac{\partial\rho/\partial z}{\partial\rho/\partial y}$$

Diapycnal upwelling and mixing are represented with a spatially varying profile of κ_v and depends on the background stratification:

$$q_{53} = q_{65} = A_3\left(\frac{\partial\kappa_v}{\partial z} + \kappa_v\frac{\partial^2 b/\partial z^2}{\partial b/\partial z}\right)$$

Northward transport depends on the meridional density gradient and the depth of the mid-depth pycnocline:

$$q_{34} = \frac{h_3^2}{g2f_N}\left(\rho_4 - \rho_3\right)$$

The temperature and salinity evolution equations take the form:

Box 1

$$\frac{\partial}{\partial t}\left(T_1V_1\right) = Q_1 - q_{12}\left(T_2 - T_1\right)$$

Box 2

$$\frac{\partial}{\partial t}\left(T_2V_2\right) = Q_2 + q_{23}\left(T_5 - T_2\right) + q_{12}\left(T_2 - T_5\right)$$

Box 3

$$\frac{\partial}{\partial t}\left(T_3V_3\right) = Q_3 + q_{23}T_2 - q_{34}T_3 + q_{53}T_5 + q_{\mathrm{MIX}}\left(T_5 - T_3\right)$$

Box 4

$$\frac{\partial}{\partial t}\left(T_4V_4\right) = Q_4 + q_{34}\left(T_3 - T_4\right)$$

Box 5

$$\frac{\partial}{\partial t}\left(T_5V_5\right) = q_{12}T_5 - q_{23}T_5 + q_{34}T_4 - q_{53}T_5 + q_{65}T_6 + q_{\mathrm{MIX}}\left(T_3 - T_5\right) + q_{\mathrm{MIX}}\left(T_6 - T_5\right)$$

Box 6

$$\frac{\partial}{\partial t}\left(T_6V_6\right) = -q_{12}T_1 - q_{65}T_6 + q_{\mathrm{MIX}}\left(T_5 - T_6\right)$$

The salinity evolution equations take the same form.

Reference state

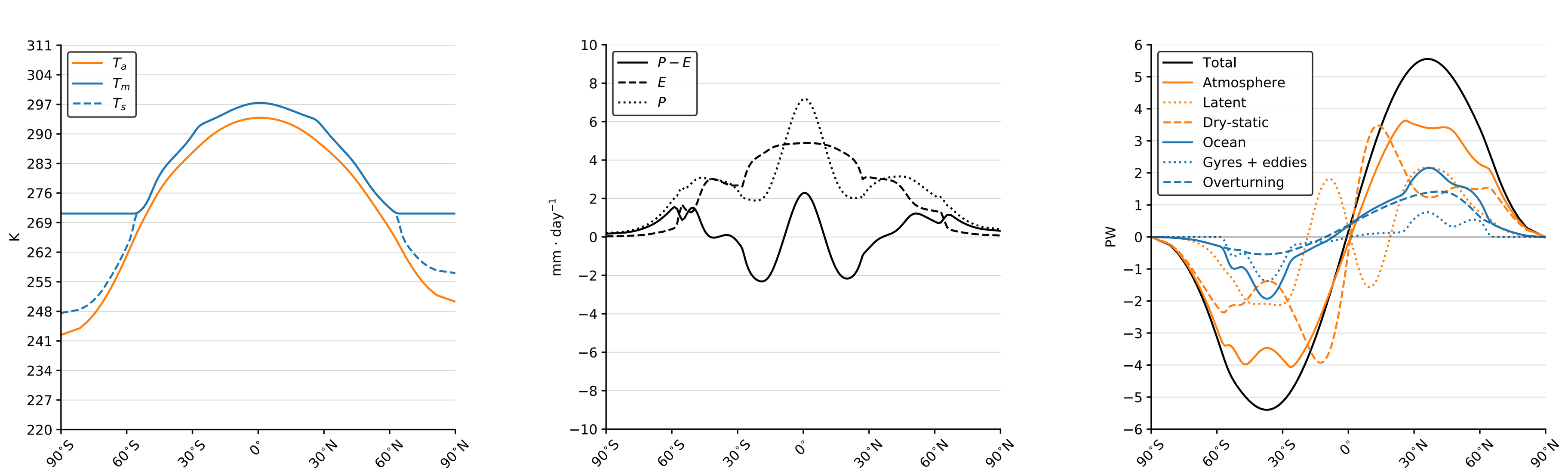

Figure 2. Profiles of (left) temperature, (middle) precipitation, and (right) meridional heat transports for the conceptual model run to equilibrium in default parameter space. The meridional heat transport is separated into the atmosphere and ocean, as well as dry-static energy, latent energy, ocean eddies and gyres, and ocean overturning.

- Equator-to-pole temperature gradient of ~ 50°C.
- Good representation of the ITCZ and subtropical regions, with a precipitation rate of 7 mm day in the deep tropics.
- Meridional heat transport by the atmosphere accounts for ~3 to 4 PW, while Meridional heat transport by the ocean accounts for ~1-2 PW. Ocean is net northward on the equator.

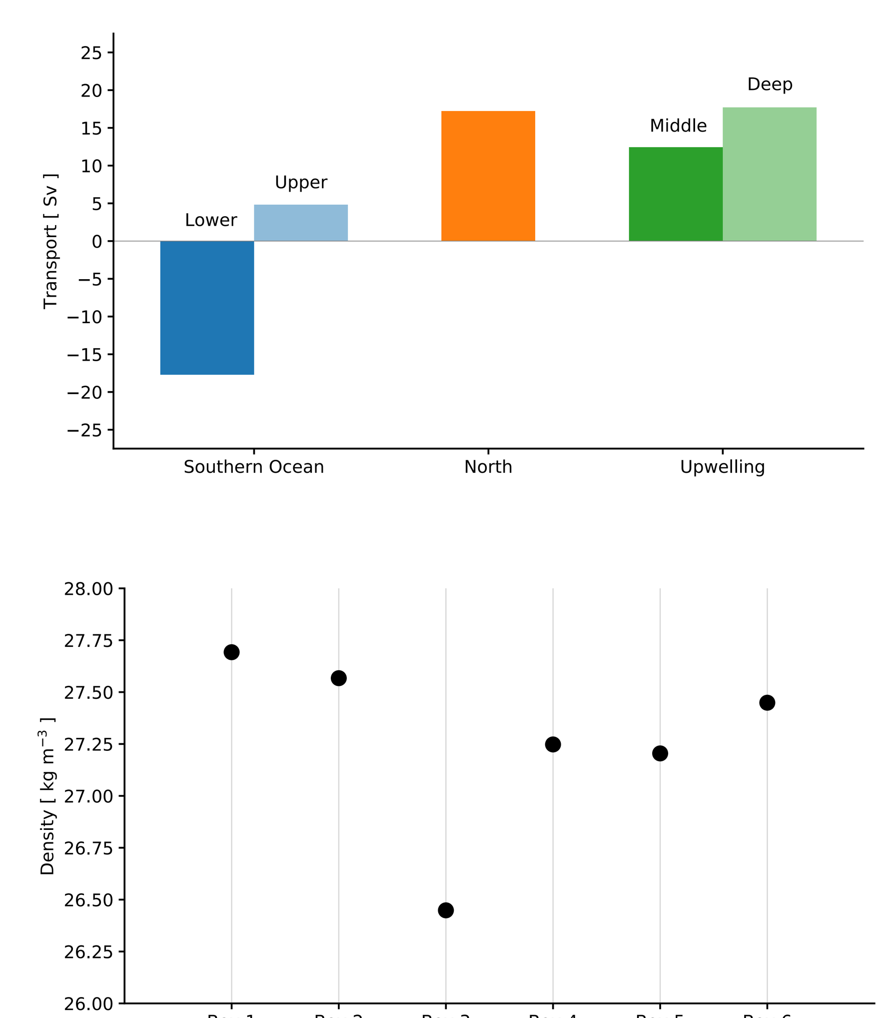

Figure 3. Plots showing the (top) strength of the ocean circulation and (bottom) stratification of the ocean in each region.

- Northward transport (e.g., AMOC) is approximately 18 Sv.
- Upper cell Southern Ocean transport is approximately 5 Sv, while lower cell Southern Ocean transport is approximately -17 Sv.
- Model captures bottom-enhanced upwelling.
- Densest water is in the Southern Ocean, with a strong meridional density gradient between the tropics and north.

Ongoing work (still in progress!)

Mechanisms of ocean heat uptake under warming

- Examining the effect of various ocean stratifications and ocean overturning circulation states on transient ocean heat uptake.

Impact of ocean-atmosphere coupling on the ocean's overturning circulation

- Evaluating the role of surface coupling on sensitivity of the ocean's overturning circulation to changes in ocean parameters using a set of novel experiments (uncoupled / coupled ocean components).

Causes of intermodel variations in ocean heat transport across GCMs

- Examining the role of ocean dynamics and coupled ocean-atmosphere processes in setting the structure of meridional ocean heat transport across CMIP GCMs.

Response of the ocean's overturning circulation to orbital forcing

- Investigating how Milankovitch cycles impact the strength and structure of the ocean's overturning circulation. Currently forcing the conceptual model with LGM and present-day like insolation.