DOE/UCAR Cooperative Agreement

Regional and Global Climate Modeling Program

Mechanisms of AMOC

Aixue Hu

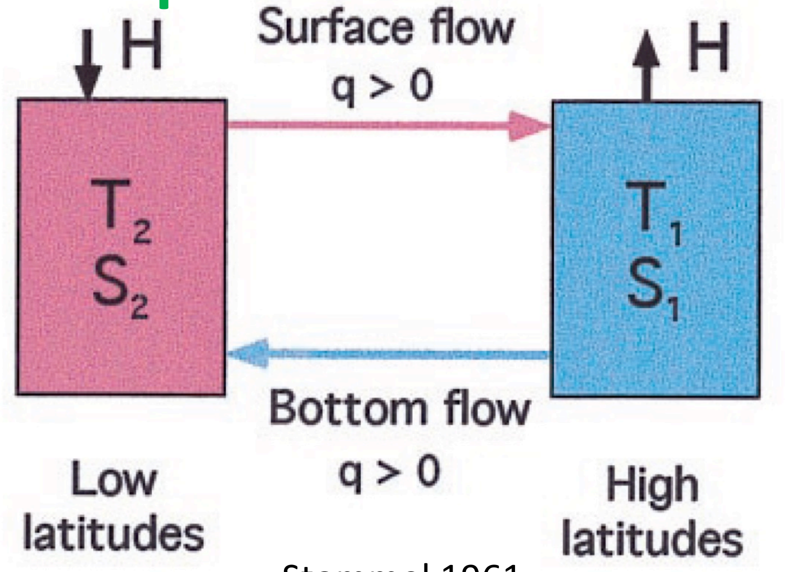
Thanks Prof. De-Zheng Sun for sharing his notes

TBI and AMV Summer School

Thermohaline Circulation or Atlantic Meridional Overturning Circulation

Theoretical models of AMOC (I)

Hemispheric two box model



Stommel 1961

Fig. 3. Stommel's conceptual model of the THC.

$$q = k[\rho_1 - \rho_2]$$
$$= k[\alpha(T_2 - T_1) - \beta(S_2 - S_1)]$$
$$= k[\alpha\Delta T - \beta\Delta S]$$

k is a hydraulic constant.

Linear sea water equation of state:

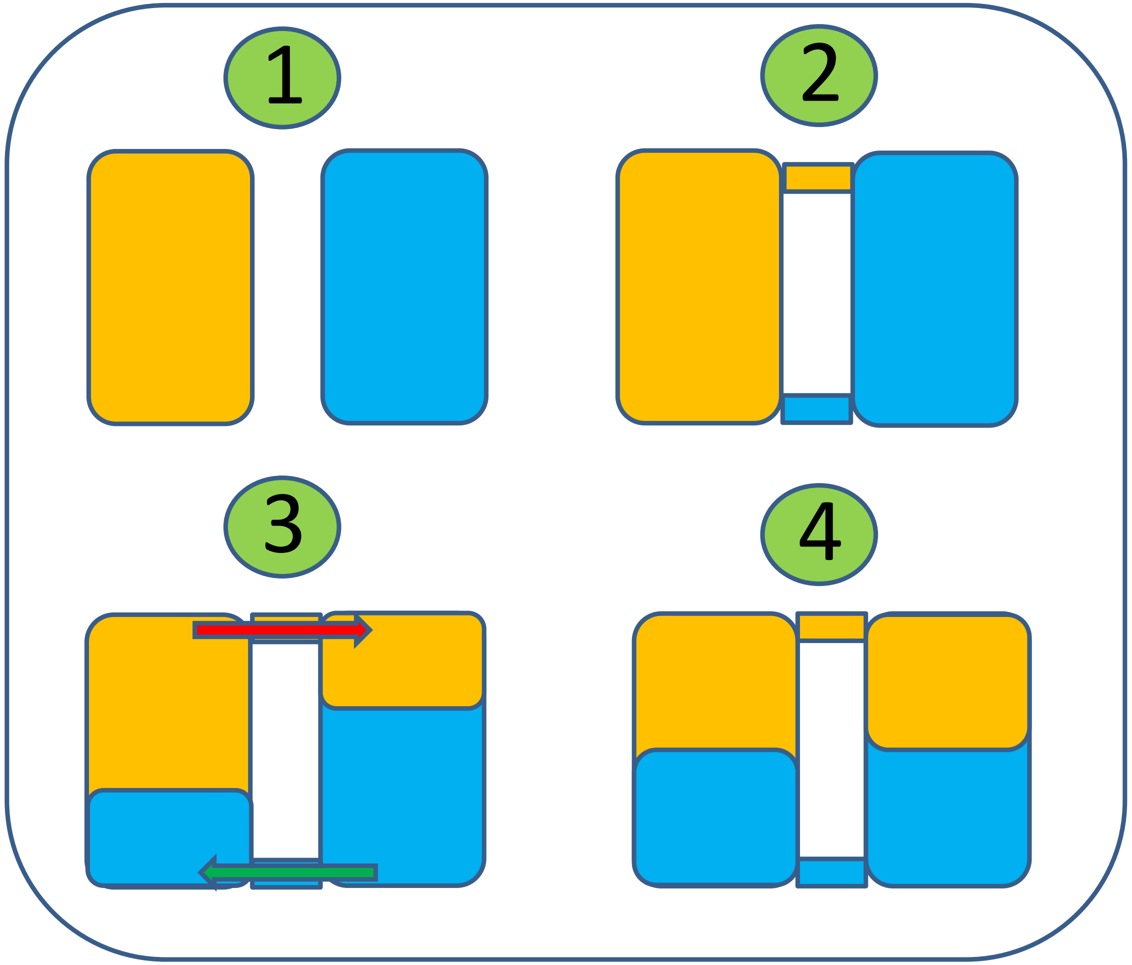
$$\rho = \rho_0 - \alpha(T - T_0) + \beta(S - S_0)$$

- AMOC strength determined by the thermal and haline related density differences between these two boxes.
- If thermal induced density contrast is larger than haline induced density contrast, AMOC flows clockwise; otherwise, AMOC flows counterclockwise.

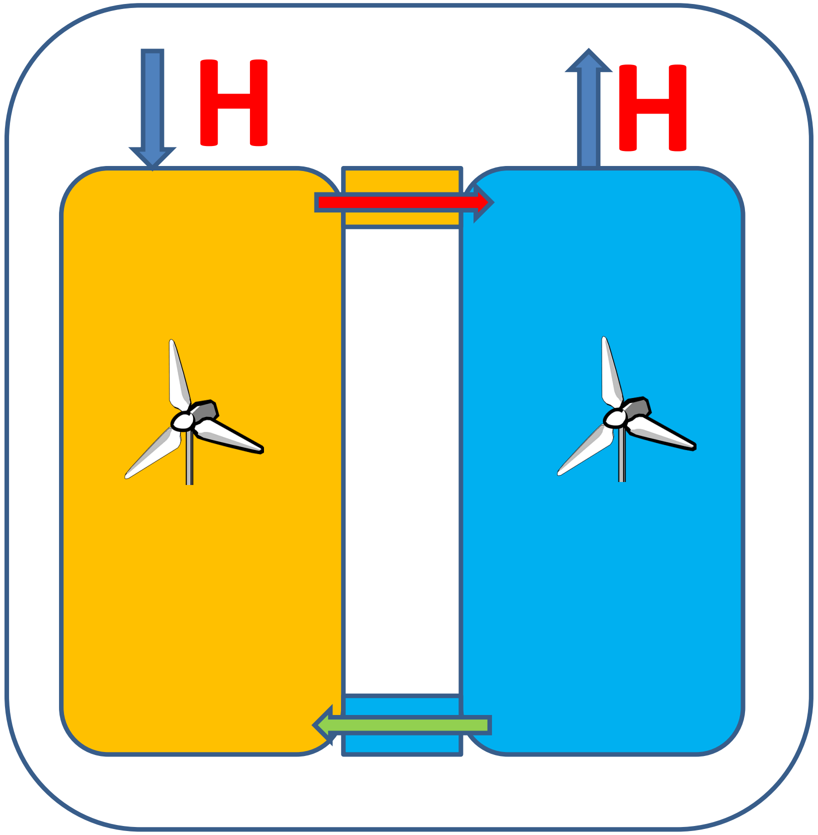
(Marotzke, PNAS, 2000)

Schematics of two box model (III)

Without external forcing



With external forcing



Two box AMOC model (2)

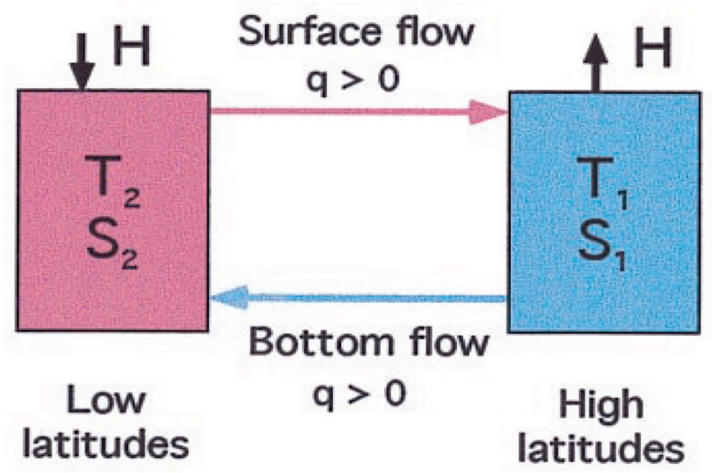


Fig. 3. Stommel's conceptual model of the THC.

$$q = k[\alpha\Delta T - \beta\Delta S]$$

$$\frac{dq}{dt} = -k\beta\frac{d\Delta S}{dt} \qquad (d\,\Delta T/dt = 0)$$

$$\frac{d\Delta S}{dt} = -2|q|\Delta S + 2H$$

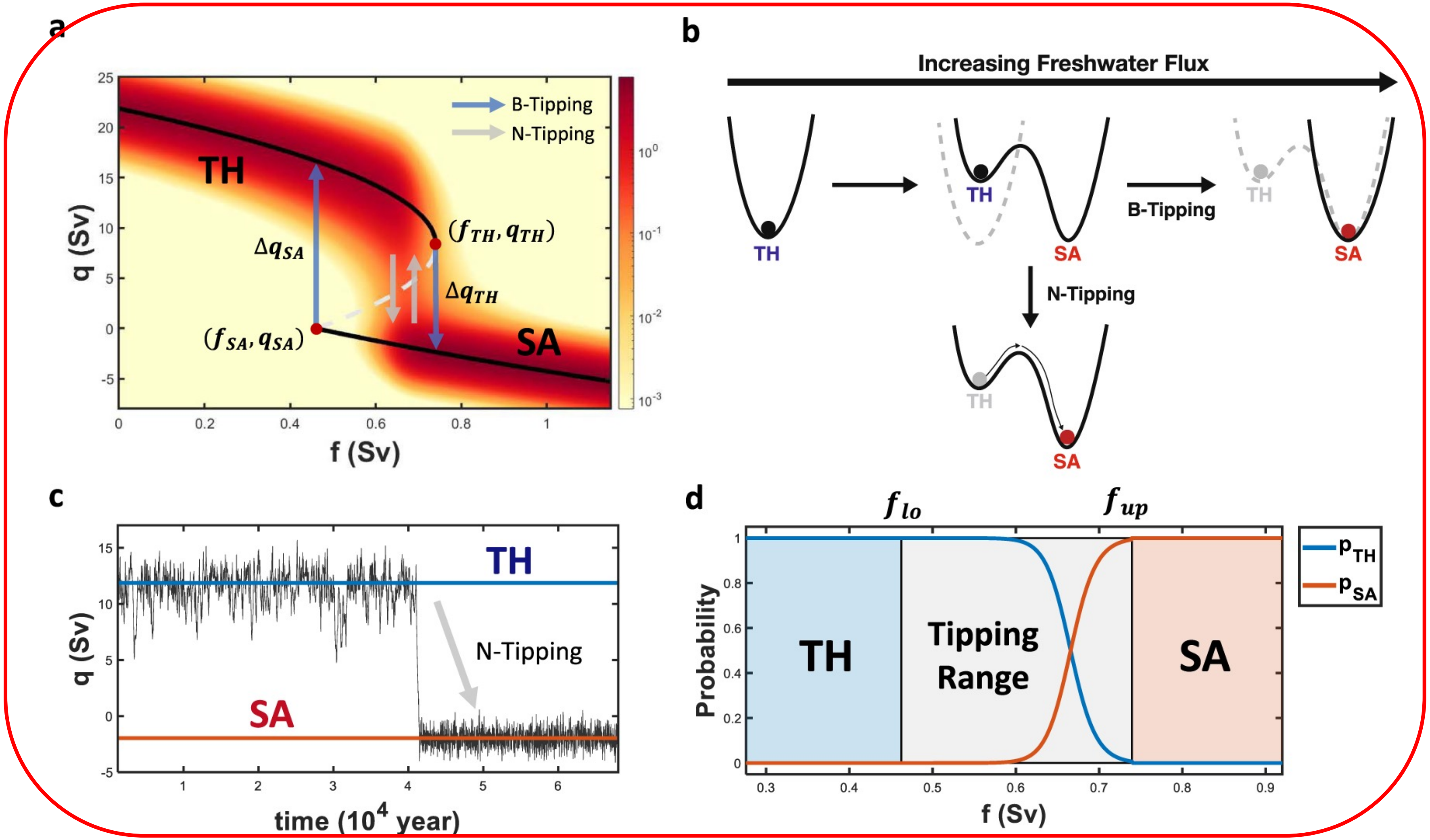
$$\Delta S = (-q + k\alpha\Delta\overline{T})/(k\beta)$$

$$\frac{dT_1}{dt} = -|q|\Delta T + \gamma(\overline{T_1} - T_1)$$
$$\frac{dT_2}{dt} = |q|\Delta T + \gamma(\overline{T_2} - T_2)$$
$$\frac{dS_1}{dt} = |q|\Delta S - H$$
$$\frac{dS_2}{dt} = -|q|\Delta S + H$$

$$\begin{cases} q_{1/2} = \frac{k\alpha\Delta\overline{T}}{2} \pm \sqrt{\left(\frac{k\alpha\Delta\overline{T}}{2}\right)^2 - Hk\beta} & q > 0 \\ q_{3/4} = \frac{k\alpha\Delta\overline{T}}{2} \pm \sqrt{\left(\frac{k\alpha\Delta\overline{T}}{2}\right)^2 + Hk\beta} & q < 0 \end{cases}$$

Referee:matteo.willeit@pik-potsdam.de

AMOC Tipping by Freshwater or Noise

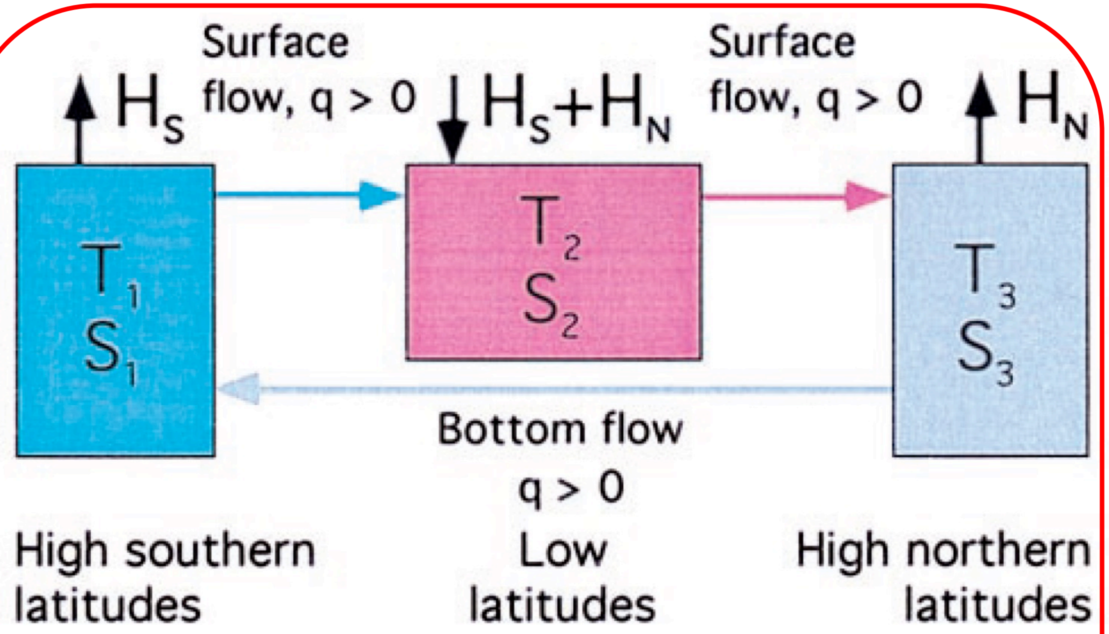


Kim, npj Climate and Atmospheric Sciences, 2022

Questions to think

- What if ΔT is not constant, but changes slowly with the varying AMOC strength? Is this slowly varying ΔT going to affect the AMOC stability?
- What if H is not only a salt flux, but also includes heat flux? Is this combined thermohaline forcing going to affect the AMOC stability?

Across Hemisphere three box model



Rooth, 1982

Fig. 4. Rooth's interhemispheric conceptual model of the THC.

$$\frac{dS_1}{dt} = -H_S + q(S_3 - S_1),$$

$$\frac{dS_2}{dt} = H_S + H_N - q(S_2 - S_1),$$

$$\frac{dS_3}{dt} = -H_N + q(S_2 - S_3),$$

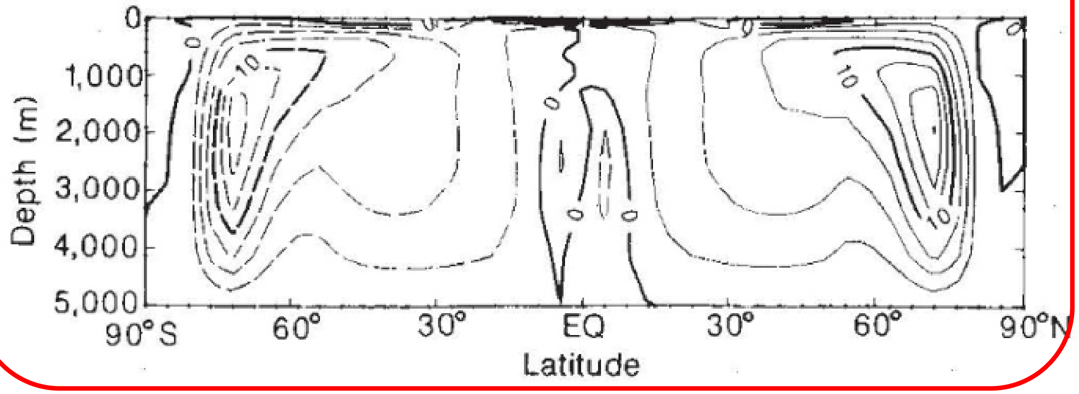
$$q = k'(\rho_3 - \rho_1) = k'\beta(S_3 - S_1),$$

$$\bar{q} = \sqrt{H_S/k'\beta}.$$

Marotzke, PNAS, 2000

Channel model in simulating AMOC stability I

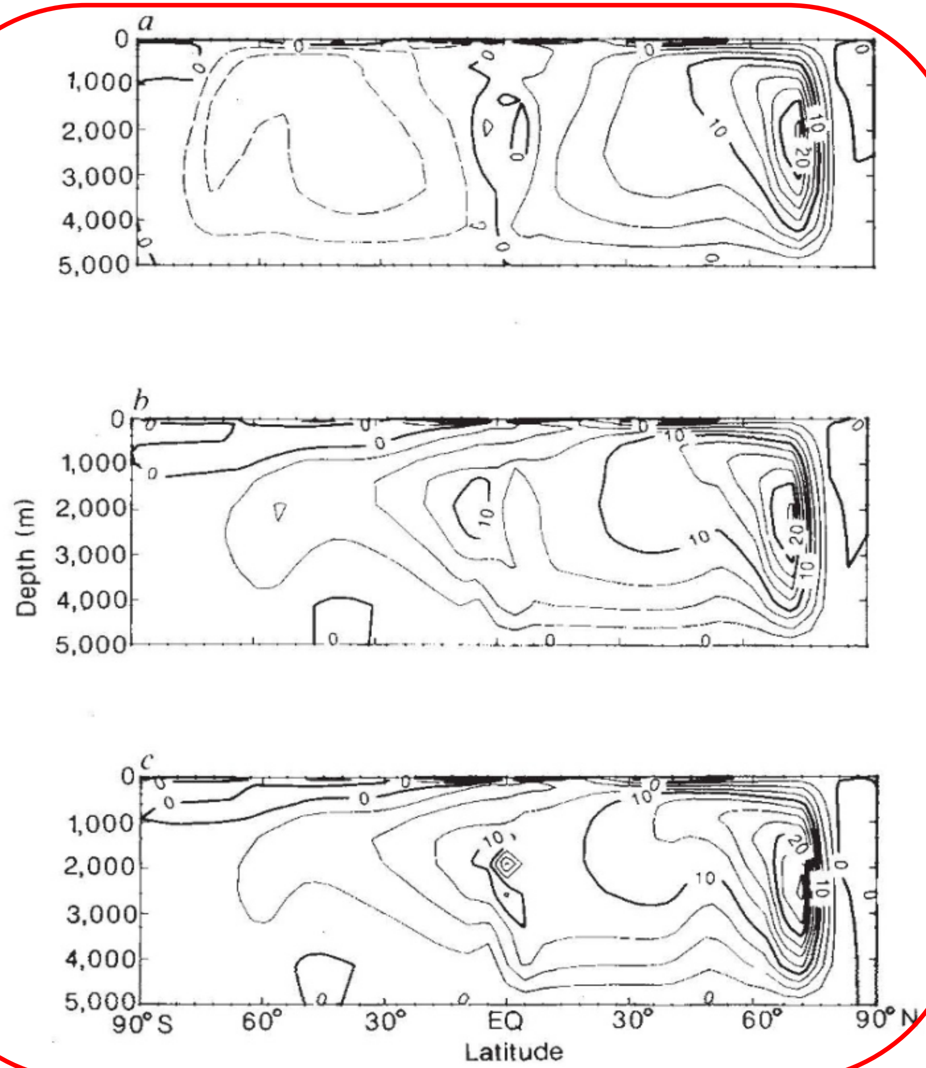
Symmetric circulation



Model:
3-d primitive equations of motion
60° wide pole to pole channel
Flat bottom with a depth of 5 km
3.75° longitude X 4.5° latitude
12 vertical levels

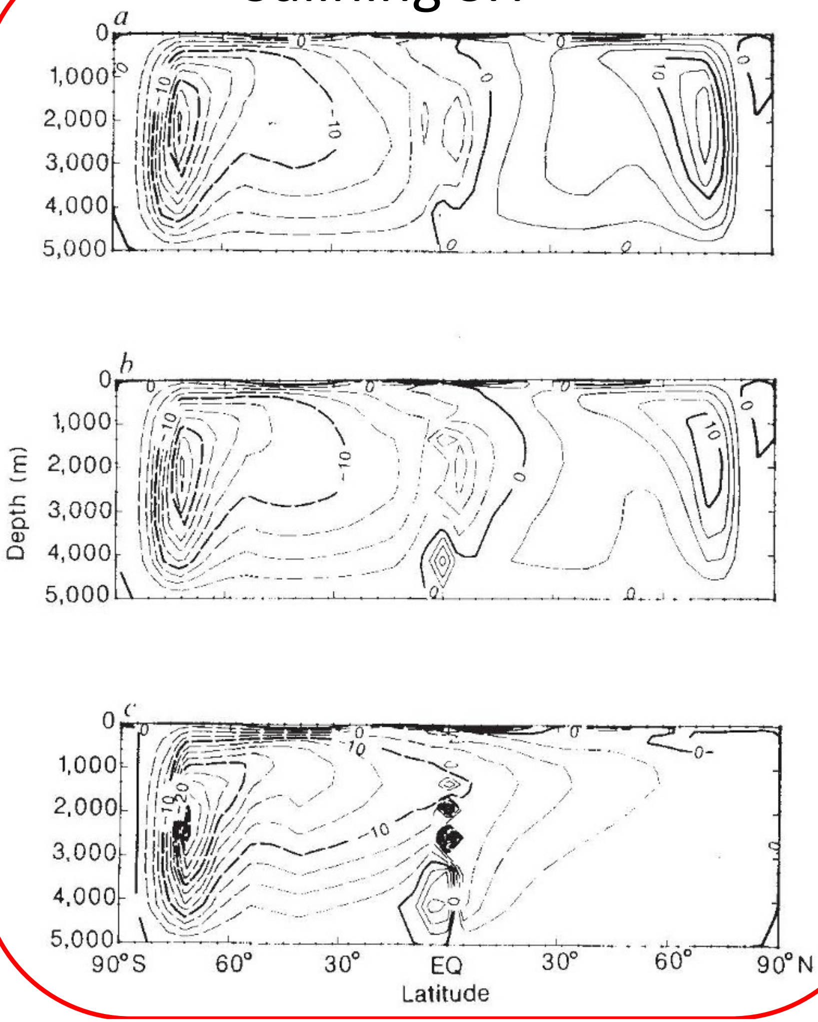
Bryan, Nature, 1986

Freshening SH



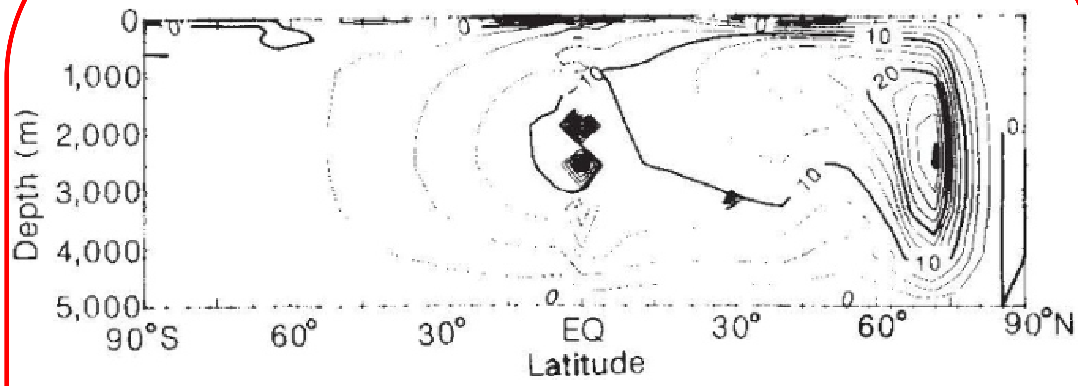
Channel model in simulating AMOC stability II

Salining SH

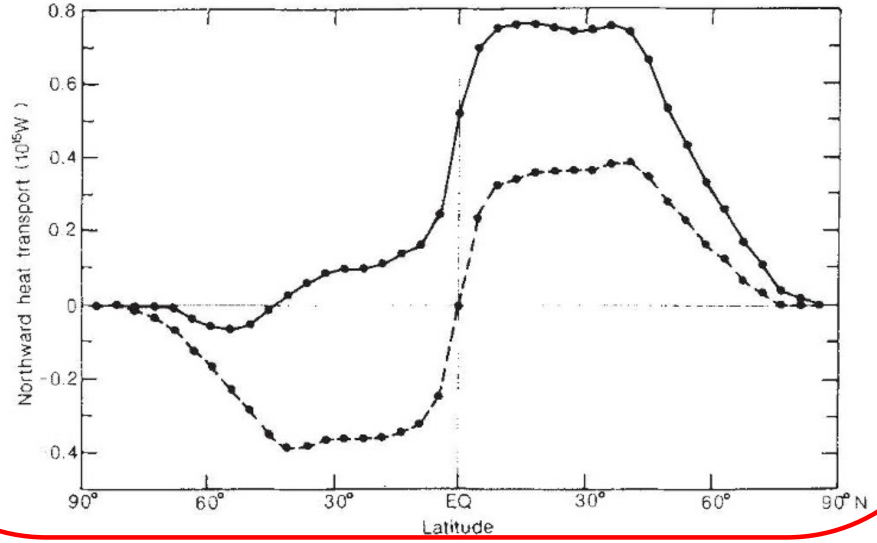


Bryan, Nature, 1986

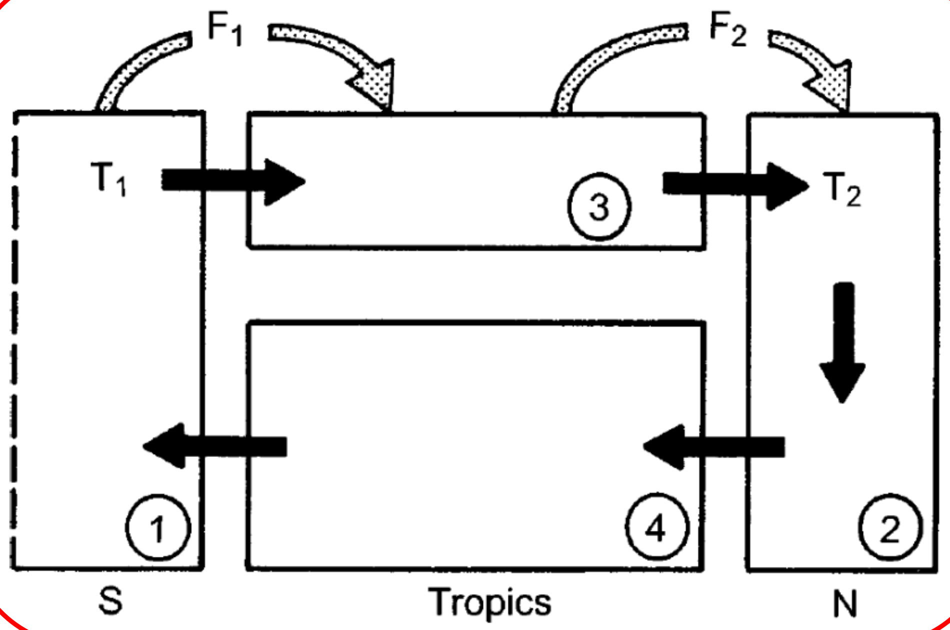
Salining NH



Meridional Heat Transport



4-Box mode for AMOC (I)



Rahmstorf, Climate Dynamics, 1996

$$m(S_2 - S_1) = -S_0 F_1$$

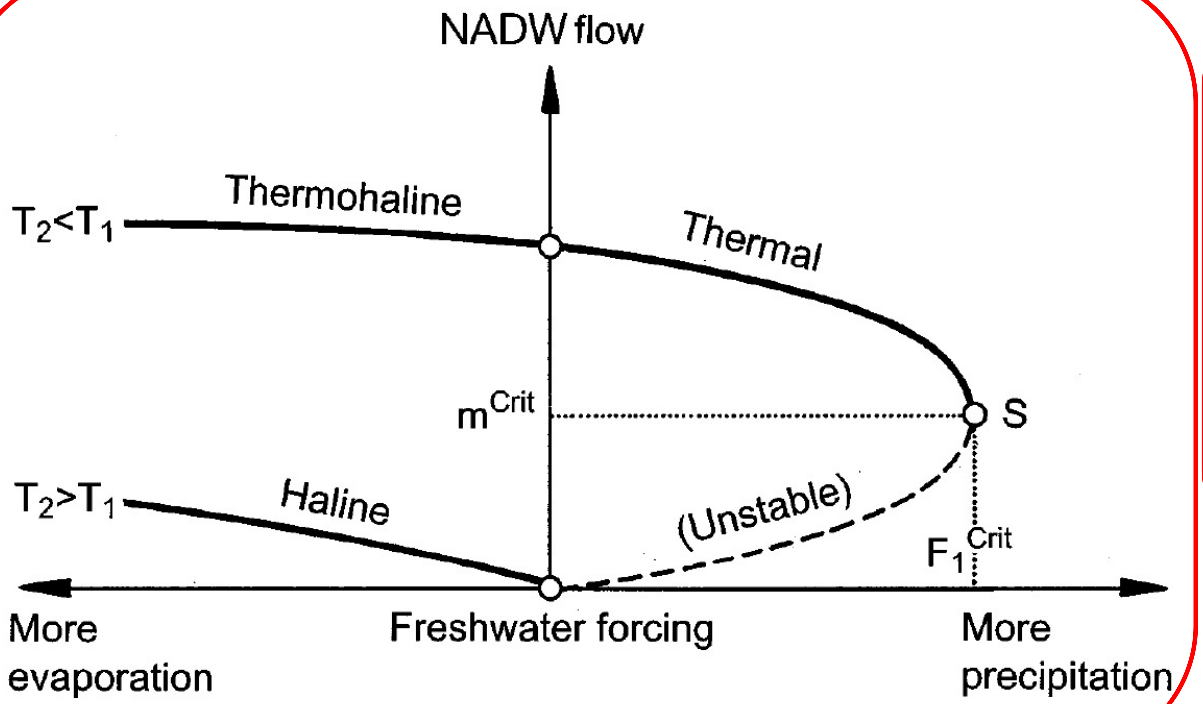
$$m(S_3 - S_2) = S_0 F_2$$

$$m = k(\rho_2 - \rho_1) = k[\beta(S_2 - S_1) - \alpha(T_2 - T_1)]$$

$$m^2 + k\alpha(T_2 - T_1)m + k\beta S_0 F_1 = 0$$

$$m = -\frac{1}{2}k\alpha(T_2 - T_1) \pm \sqrt{\frac{1}{4}[k\alpha(T_2 - T_1)]^2 - k\beta S_0 F_1}$$

4-Box mode for AMOC (II)



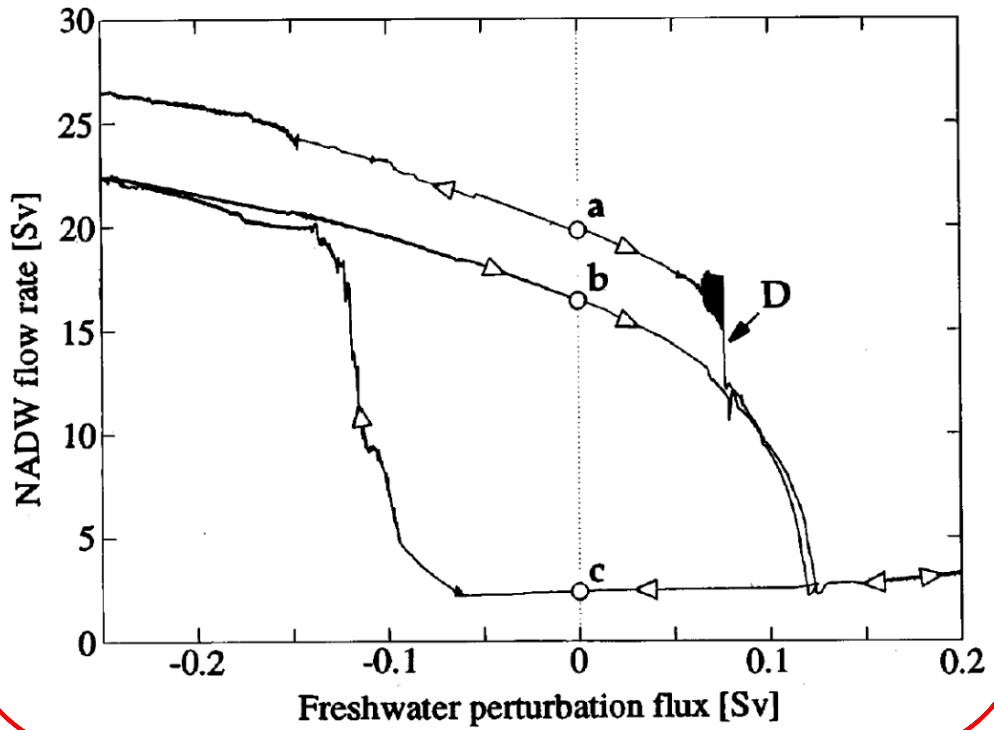
T2<T1, F1<0 Thermohaline driven
T2>T1, F1<0 purely haline driven
T2<T1, F1>0 Thermal driven, haline counter thermal effect
T2>T1, F1>0 m is unconditionally negative

Rahmstorf, Climate Dynamics, 1996

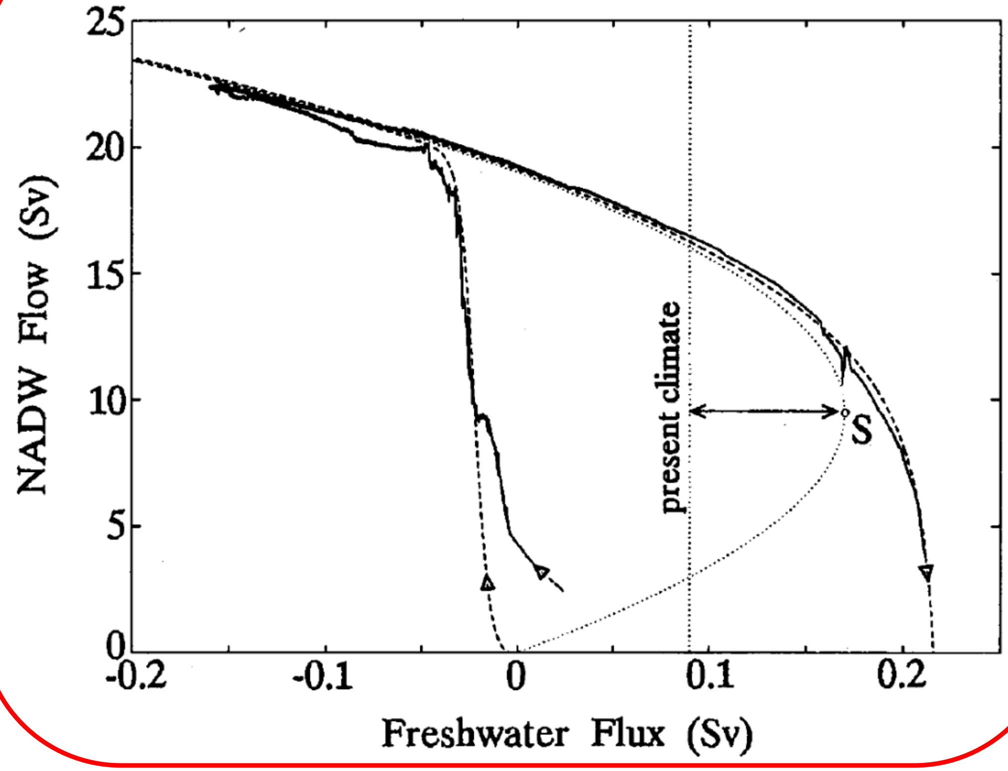
$$m = -\frac{1}{2}k\alpha(T_2 - T_1) \pm \sqrt{\frac{1}{4}[k\alpha(T_2 - T_1)]^2 - k\beta S_0 F_1}$$

Results of the 4-box model and EMIC model

Box model hysteresis loop



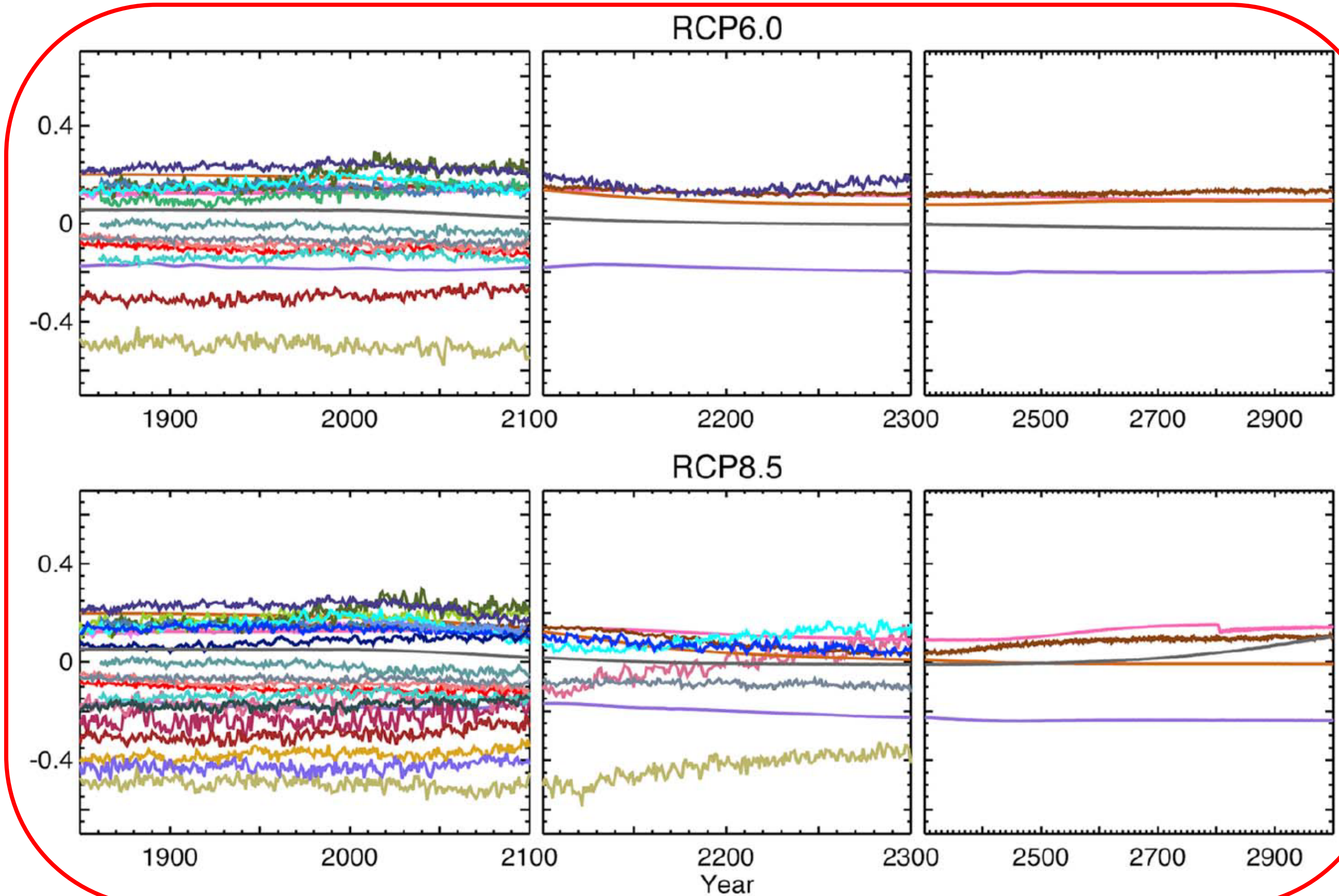
Box model and EMIC model hysteresis loop



F_{ov} is proposed as a diagnostic indicator of AMOC stability; $F_{ov} < 0$ AMOC multi-equilibrium states; $F_{ov} > 0$, AMOC monostable.

Rahmstorf, Climate Dynamics, 1996

AMOC stability and Fov in CMIP5 models

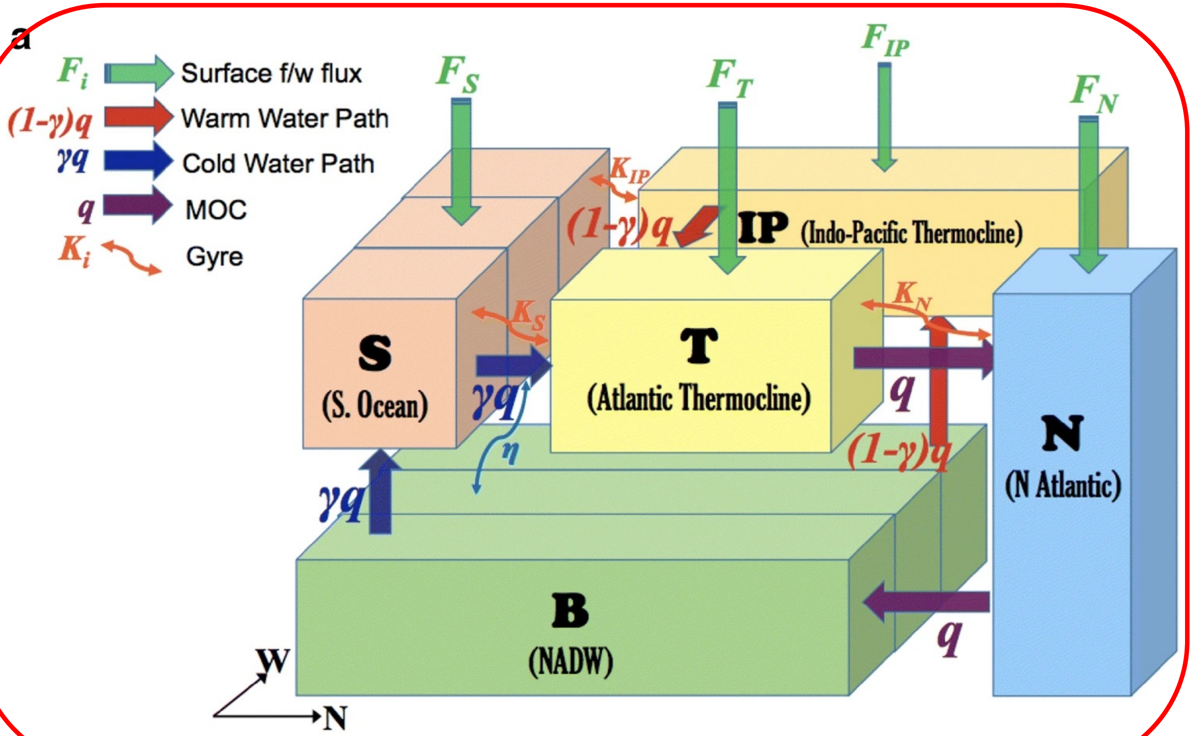


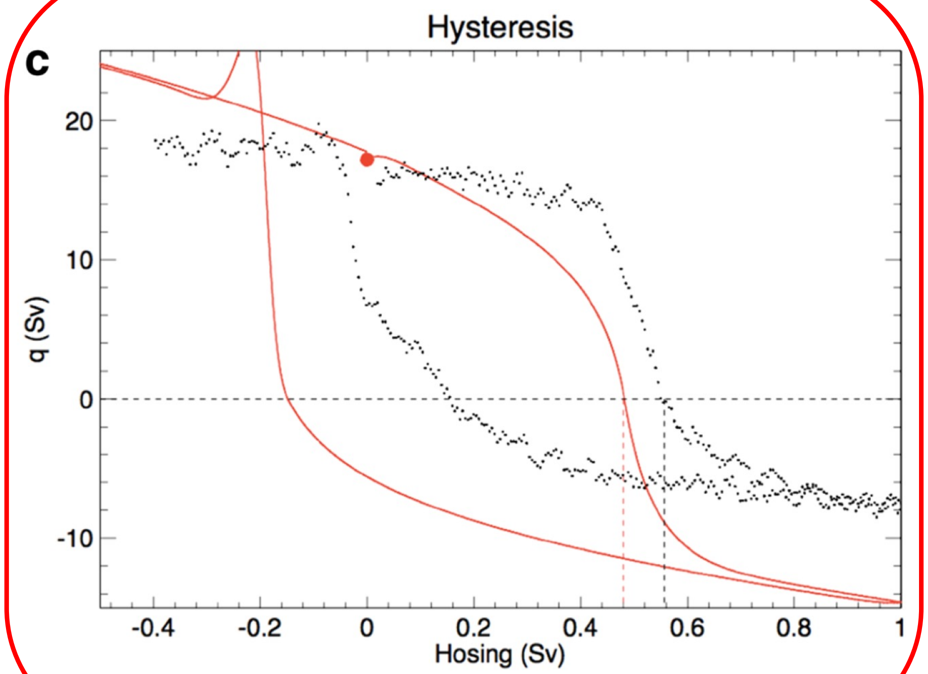
5 EMIC models and 25 CMIP5 models

11 models with $F_{ov} < 0$
15 models with $F_{ov} > 0$
4 models show that F_{ov} changes signs over time.

Weaver et al., GRL, 2013

5-box model for AMOC





$$q = \lambda \left[\alpha \left(T_S - T_N \right) + \beta \left(S_N - S_S \right) \right]$$

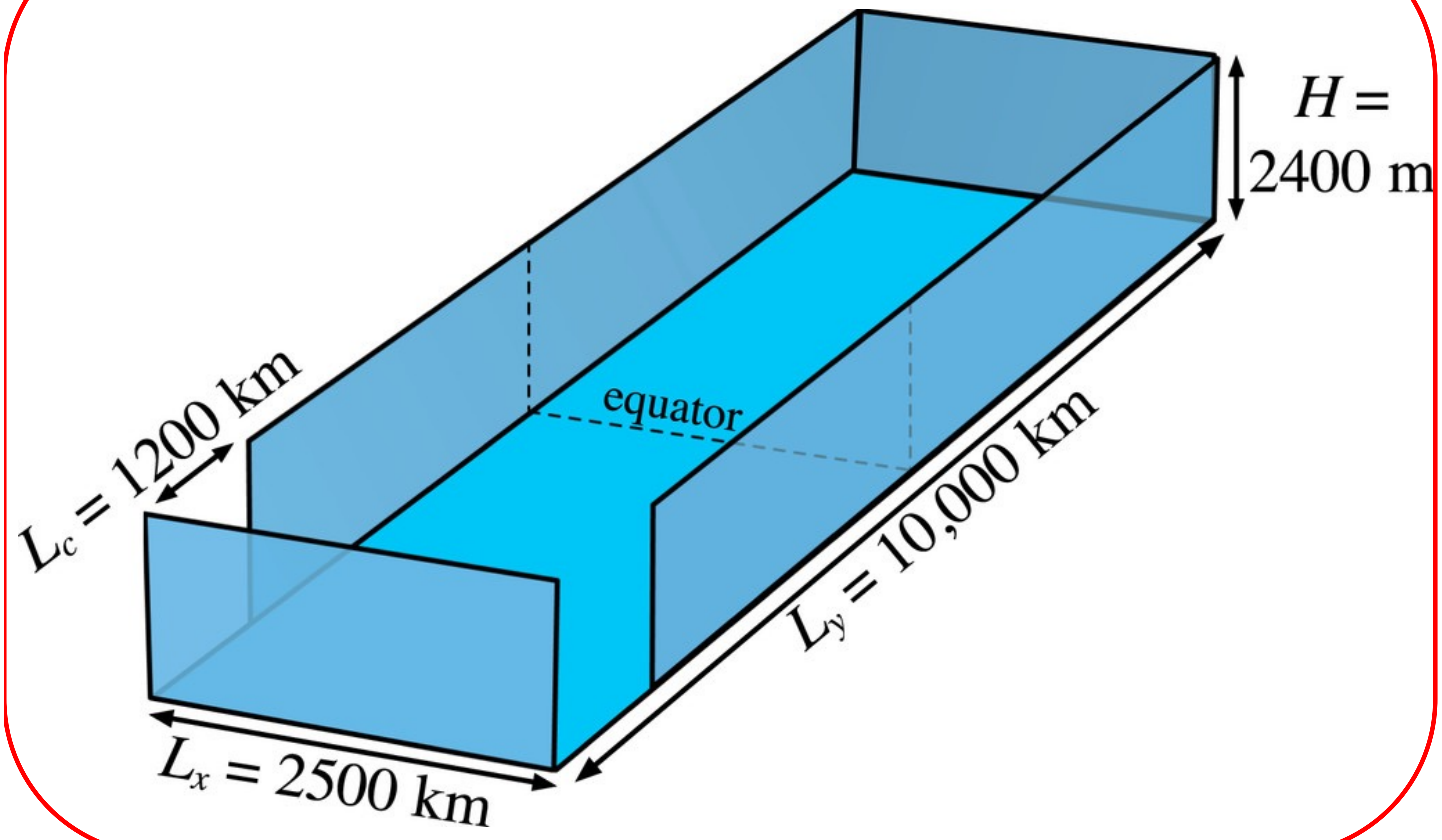
F_{ov} is not a good indicator of AMOC bi-stability
H_{crit} is about -0.2 Sv based on observations

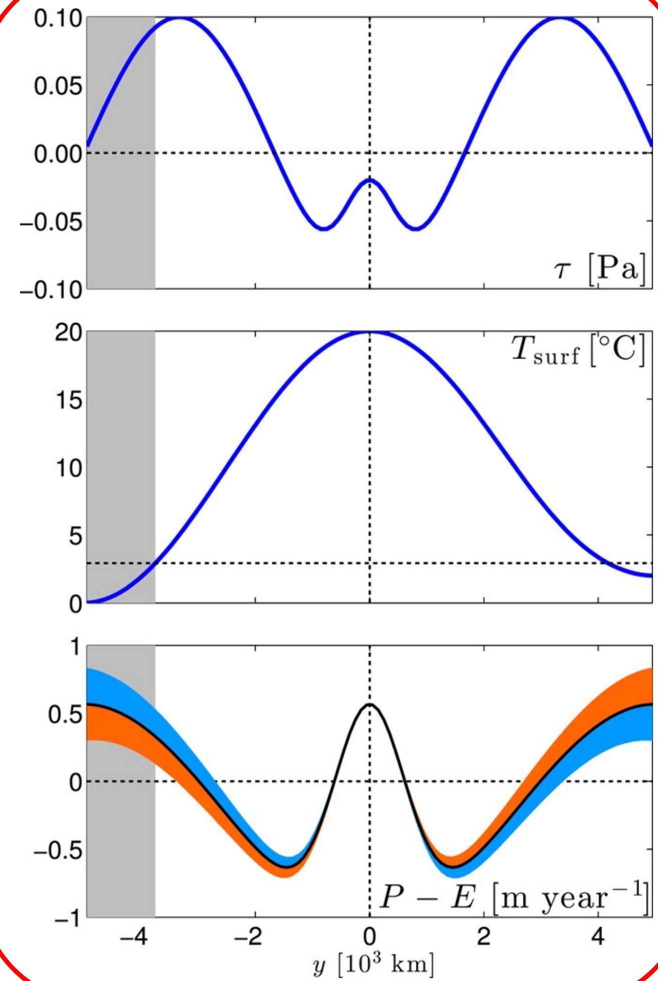
Wood et al.. Climate Dvnamics. 2019

5-box model for AMOC

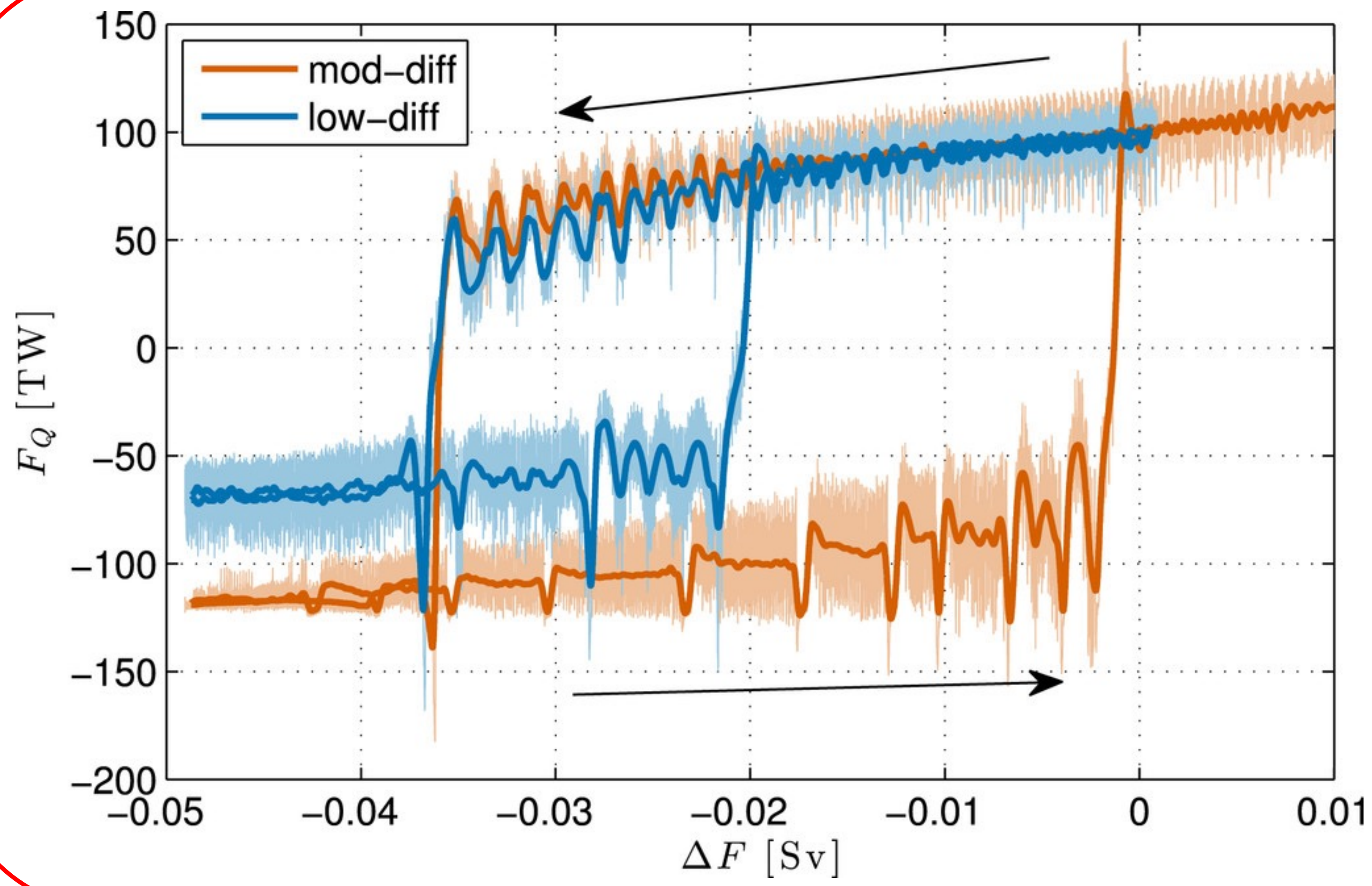
K_N: Higher values of K_N result in a larger H_{crit}.
K_S: Larger values of K_S result in a smaller H_{crit}.
KIP: Larger values of *KIP* result in a smaller H_{crit}.
λ: The sensitivity is weak because a change in λ does not directly change the North Atlantic freshening (hosing) needed to bring the N–S density difference to zero.
γ: Larger values of γ have smaller values of H_{crit}.
Fi: Here all the surface freshwater fluxes are scaled by a factor of 0.5 or 1.5, maintaining zero global mean flux in each case. A stronger mean hydrological cycle results in a larger initial salinity difference ($S_N - S_S$). Hence more hosing is needed to reverse the density gradient, and larger freshwater fluxes result in a larger H_{crit}.

Idealized channel model



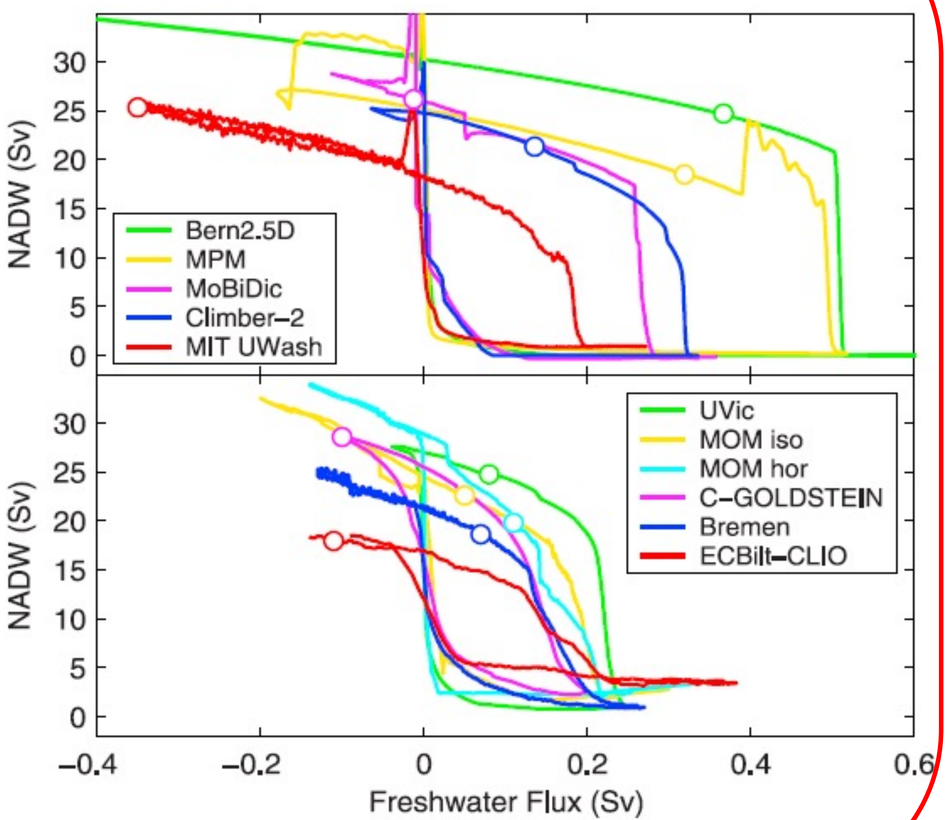


Idealized channel model



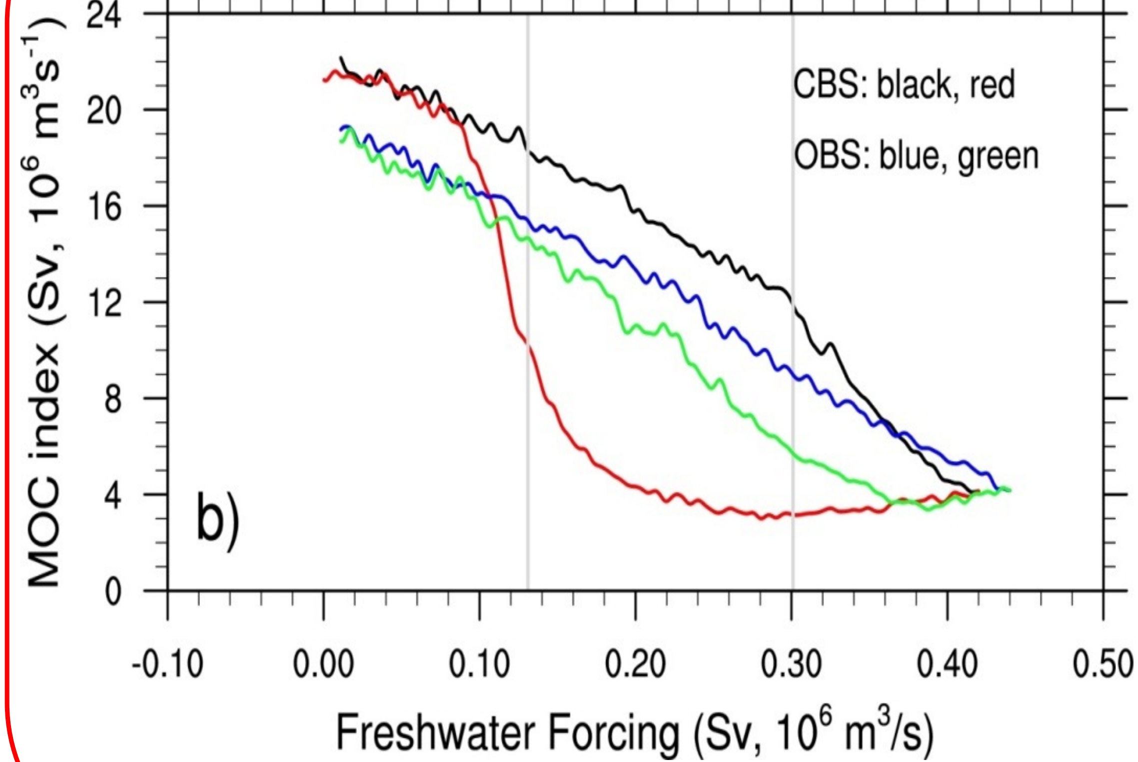
AMOC hysteresis in models

Earth model of intermediate complexity



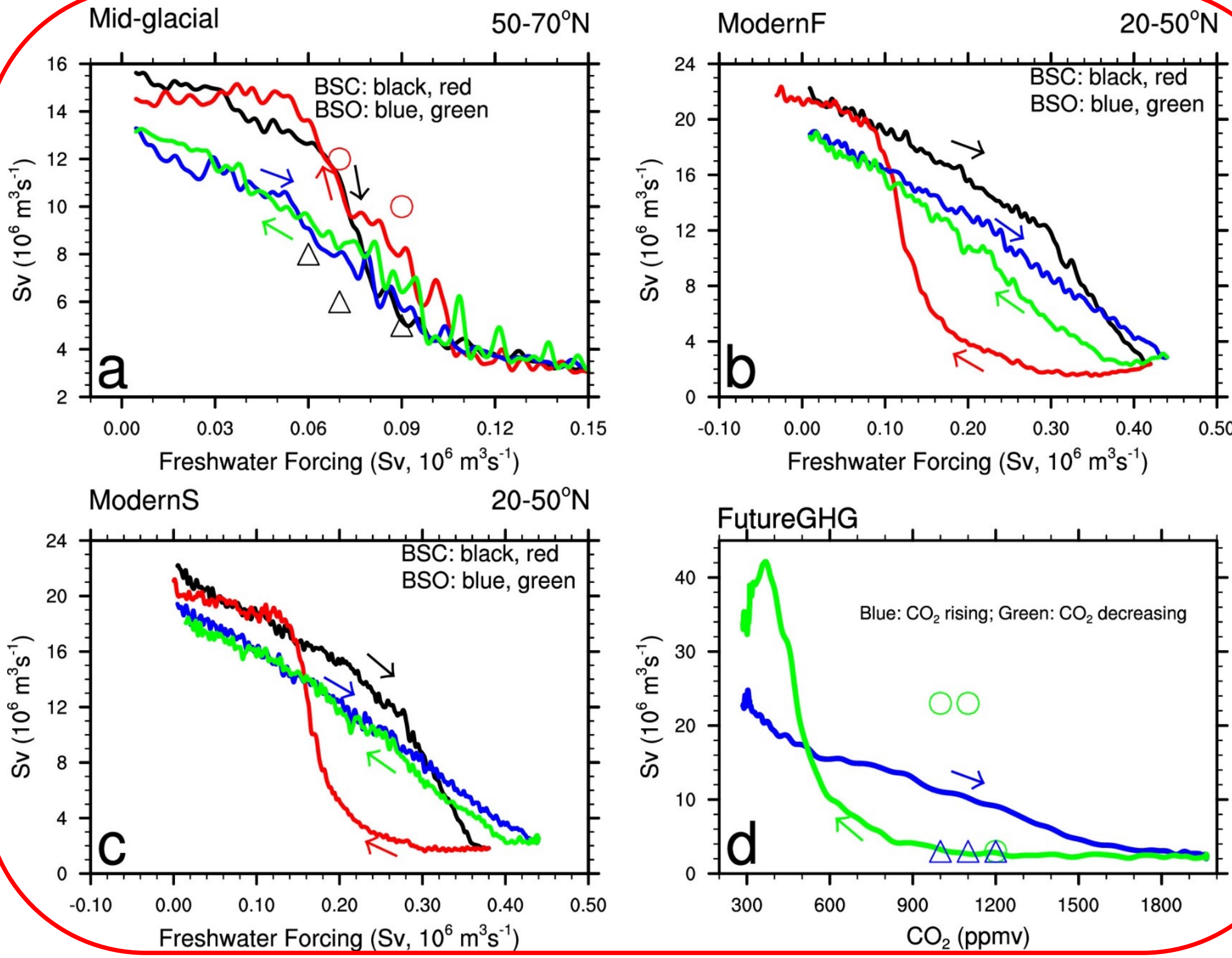
Rahmstorf et al., GRL, 2005

Fully Coupled Climate Model

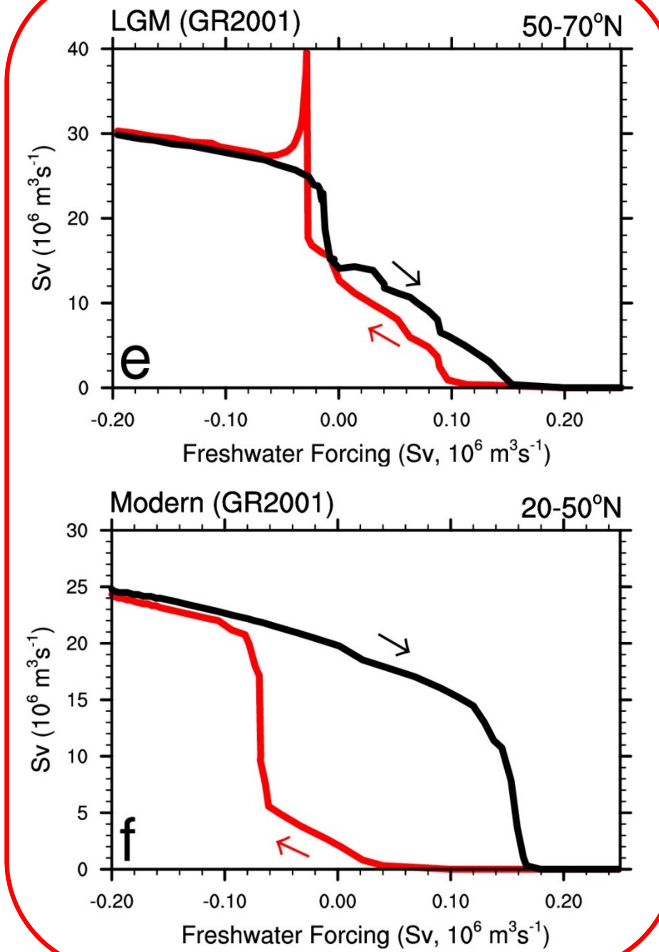


Hu et al., 2012

AMOC hysteresis under freshwater and Greenhouse gas forcing



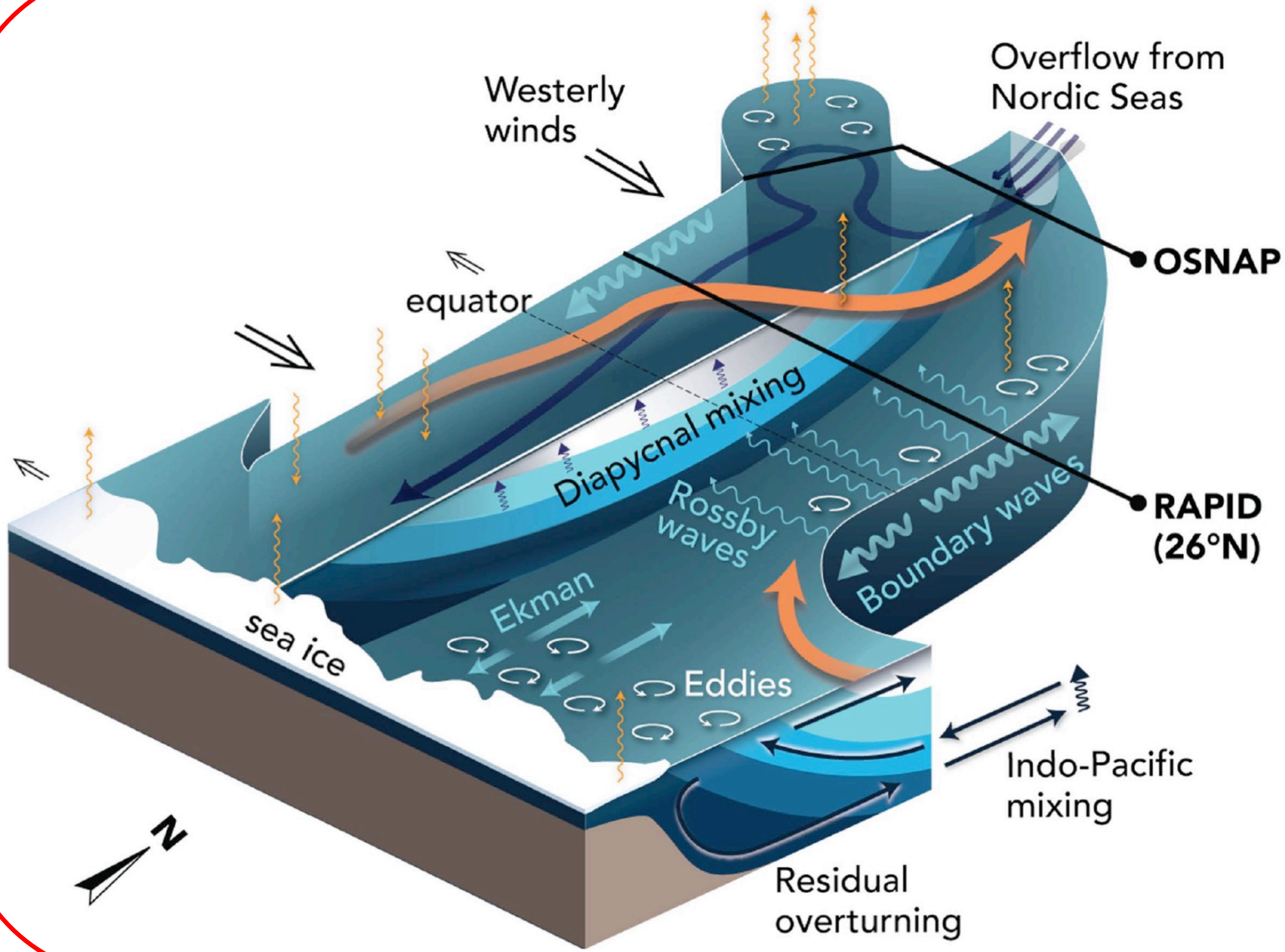
Hu et al., Communications Earth and Environment, 2013



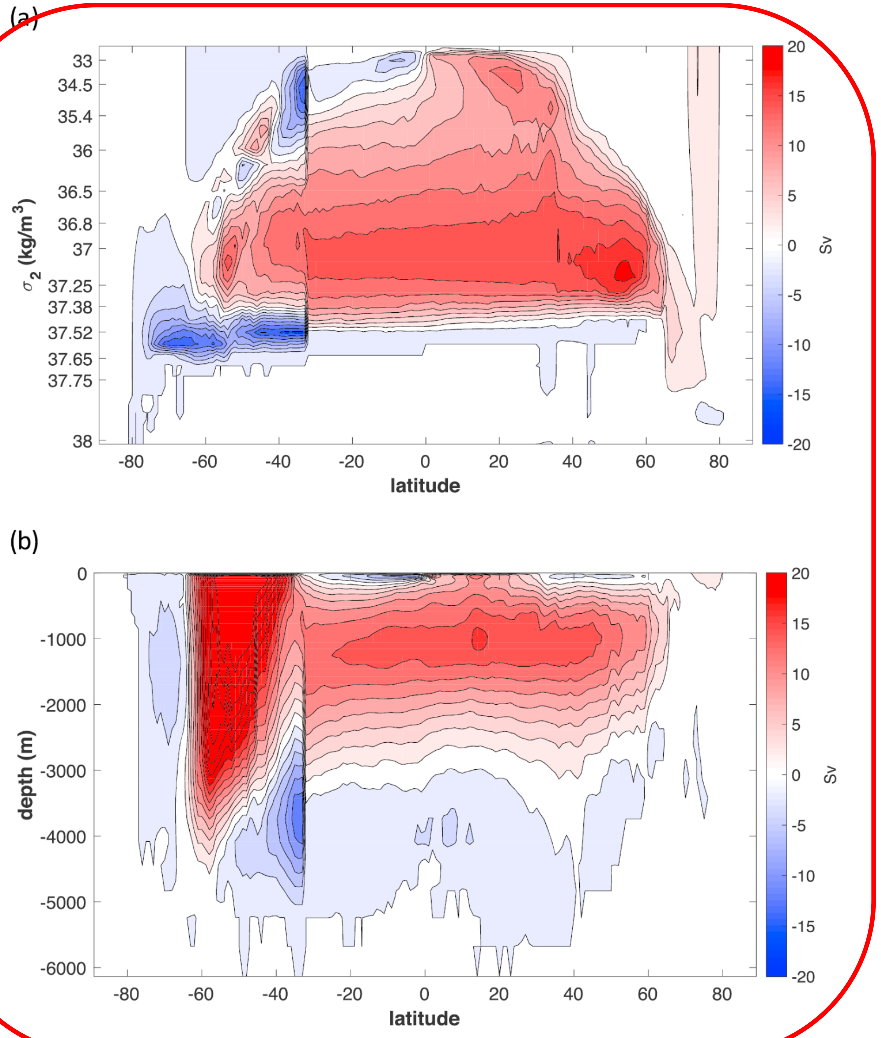
Ganopolski and Rahmstorf, Nature, 2001

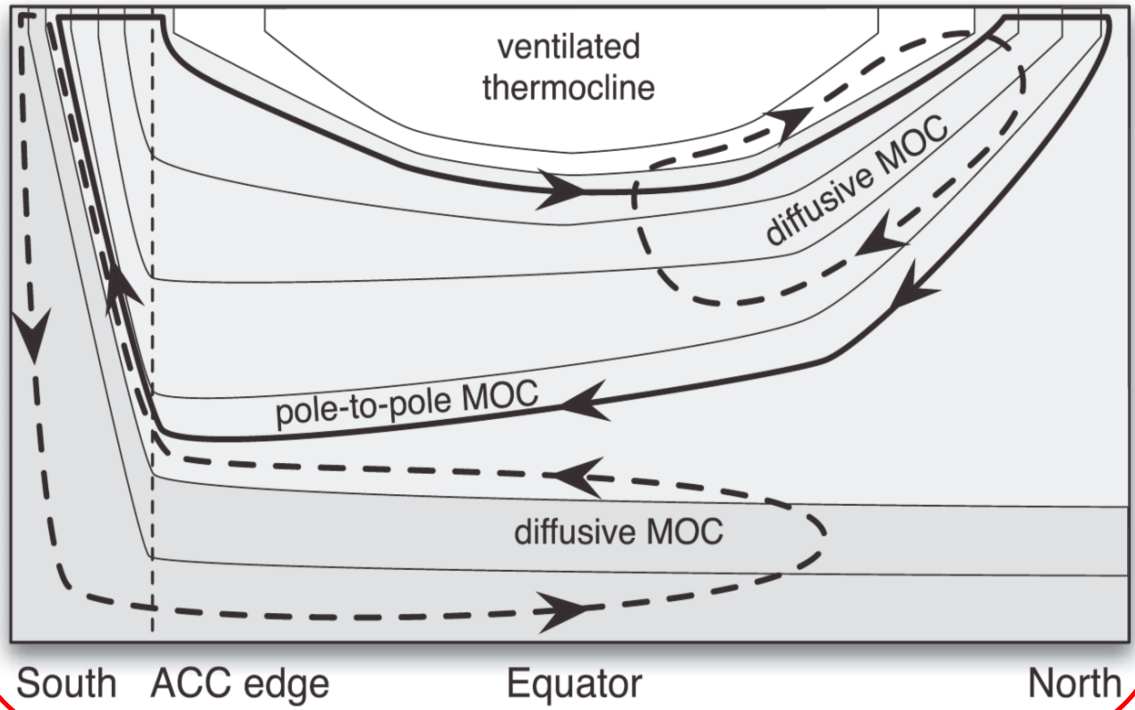
Key processes controlling the AMOC strength and variability

Johnson et al., JGR-Oceans, 2019



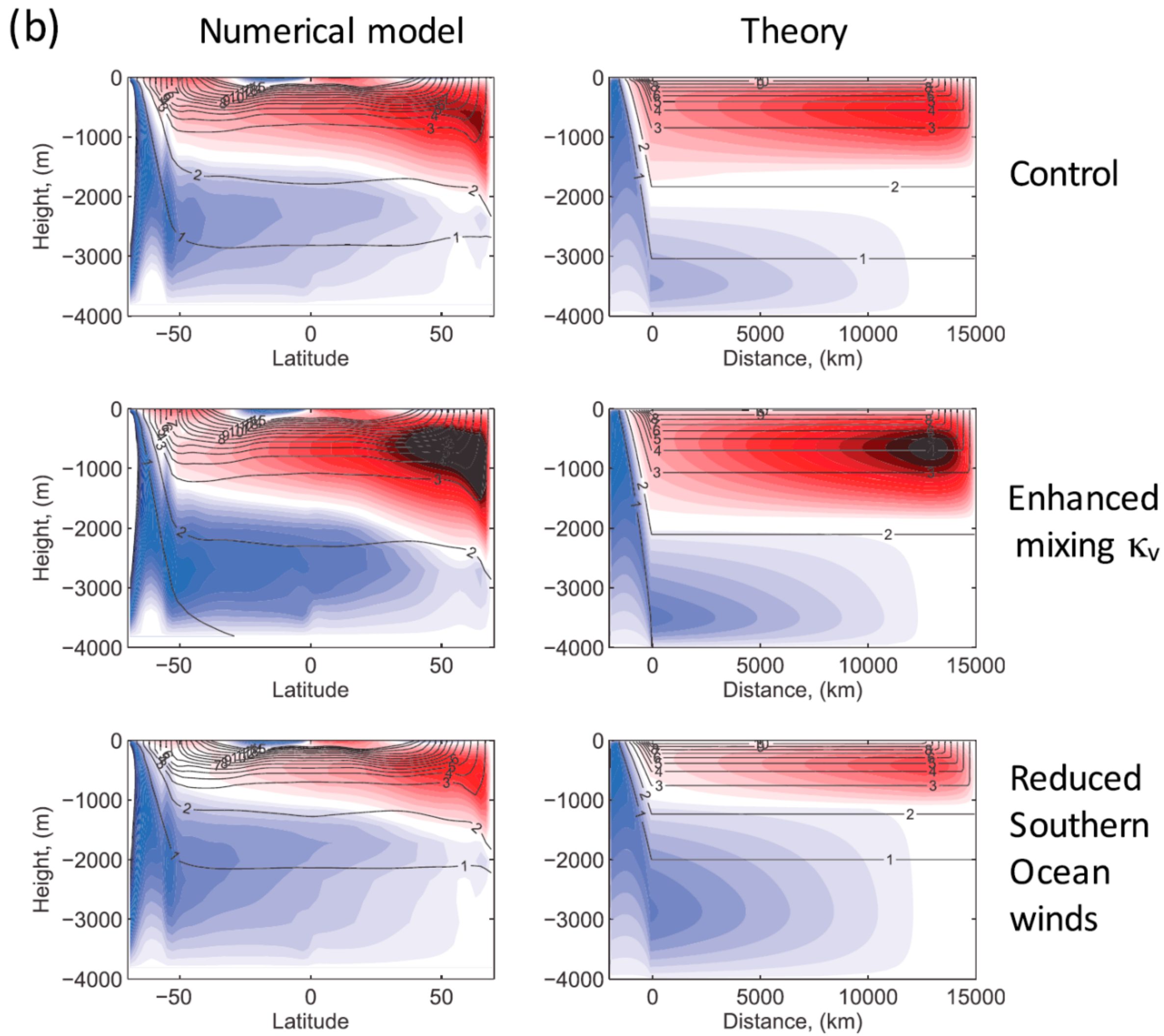
Atlantic Meridional Streamfunction





Johnson et al., JGR-Oceans, 2019

Sensitivity of AMOC to mixing and southern ocean winds



Summary

- It is now clear that the wind, in both hemispheres, plays a prominent role in setting the mean AMOC strength and determining its variability. This includes the interaction between the wind stress and the surface buoyancy distribution, because wind-driven upwelling, and the vertical flux of buoyancy associated with eddies and gyres, brings water to the surface where its density can be transformed by buoyancy fluxes.
- Simplified models are moving from two-dimensional zonally averaged representations to geometries that capture the fundamentally three-dimensional aspects of the circulation. This includes a distinction between the western boundary and the basin interior and a focus on the circulation between multiple basins.
- The degree to which water mass transformation in the ocean interior is important, or whether the circulation in the Atlantic sector is essentially adiabatic, remains an open question.
- Multiple lines of evidence suggest that eddies at high latitudes in both hemispheres are essential to the dynamics of the AMOC. Johnson et al., JGR-Oceans, 2019

DOE/UCAR Cooperative Agreement

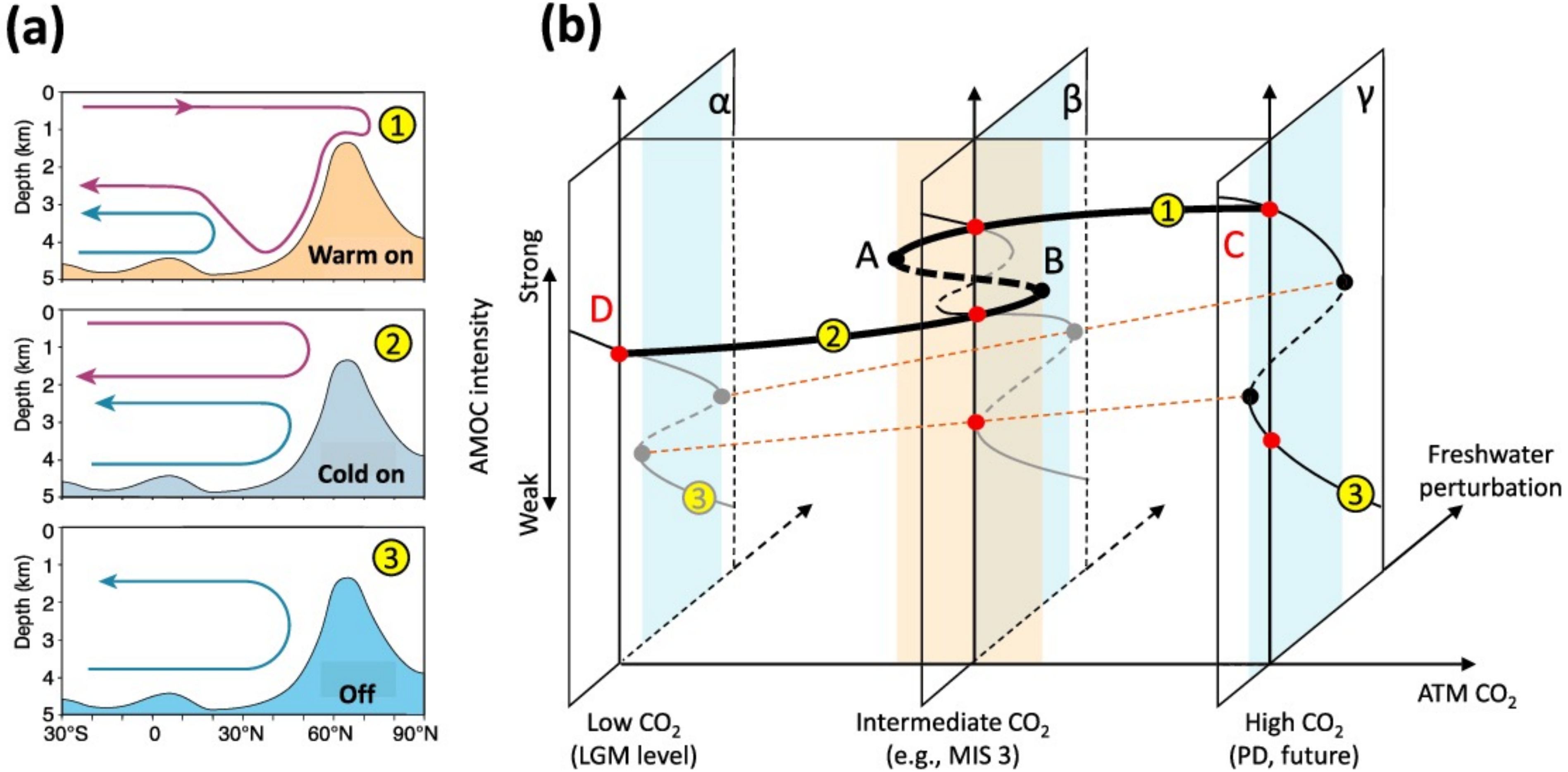
Regional and Global Climate Modeling Program

THANK YOU!

QUESTIONS?

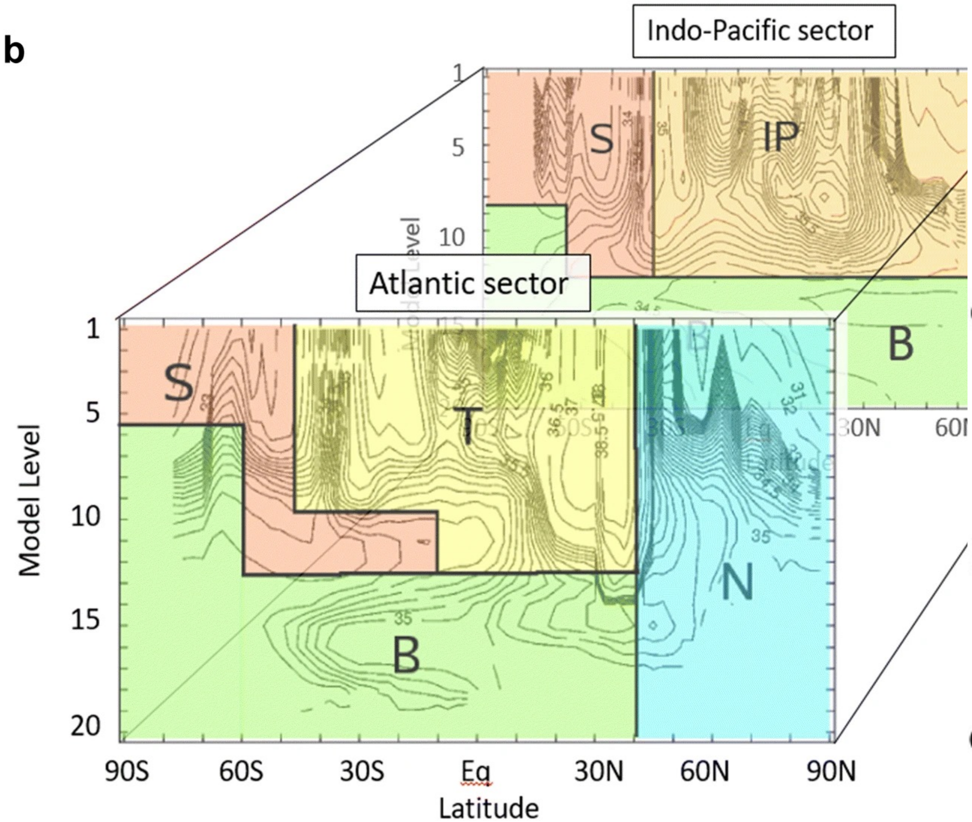
CONTACT: AHU@UCAR.EDU

AMOC hysteresis and abrupt climate change

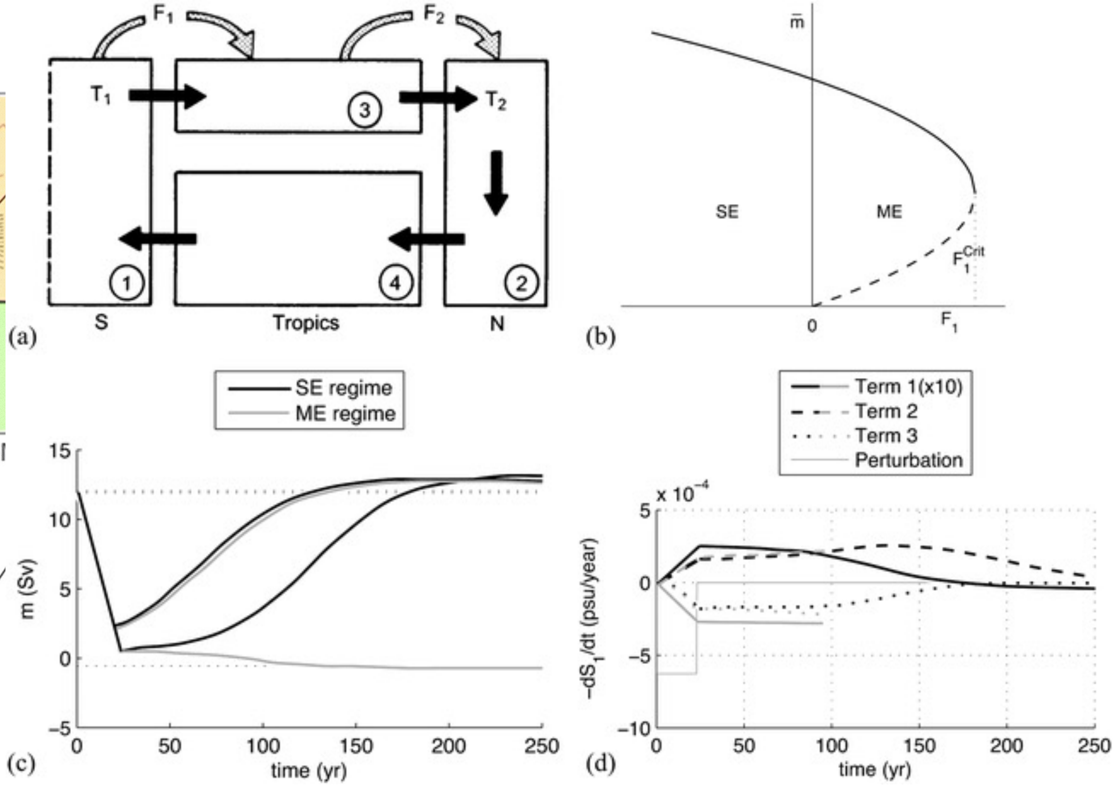


Weijer et al., JGR-ocean, 2019

b



Indo-Pacific sector
Atlantic sector
S
IP
B
T
N
Model Level
Latitude
90S
60S
30S
Eq
30N
60N
90N
1
5
10
15
20



(a)
F_1
F_2
T_1
T_2
1
2
3
4
S
Tropics
N
(b)
$\bar{m}$
SE
ME
F_1^{Crit}
0
F_1
(c)
SE regime
ME regime
m (Sv)
time (yr)
(d)
Term 1(x10)
Term 2
Term 3
Perturbation
x 10^{-4}
$-dS_1/dt$ (psu/year)
time (yr)