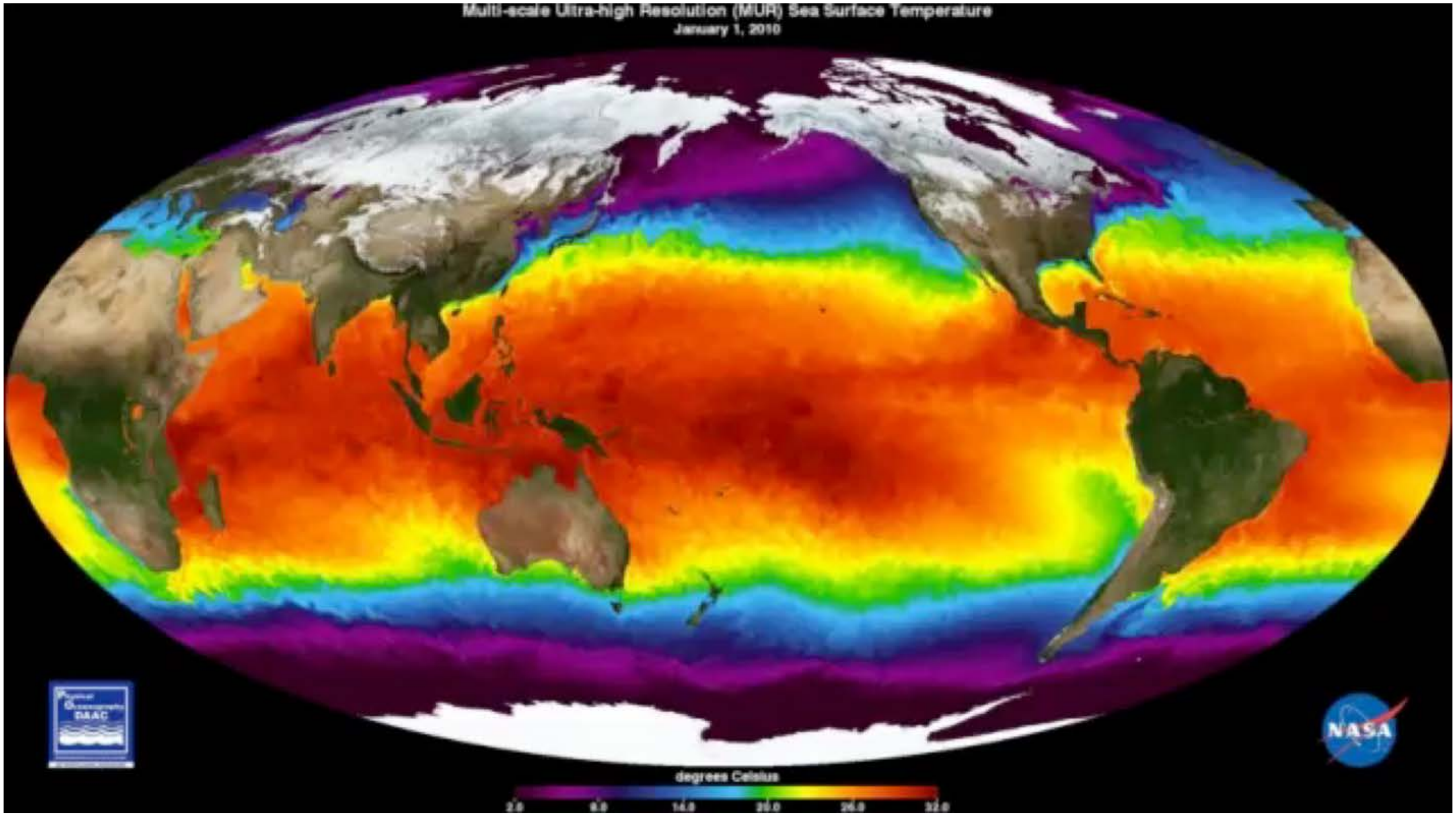
Week 8: Seasonality, air-sea interaction and ventilation

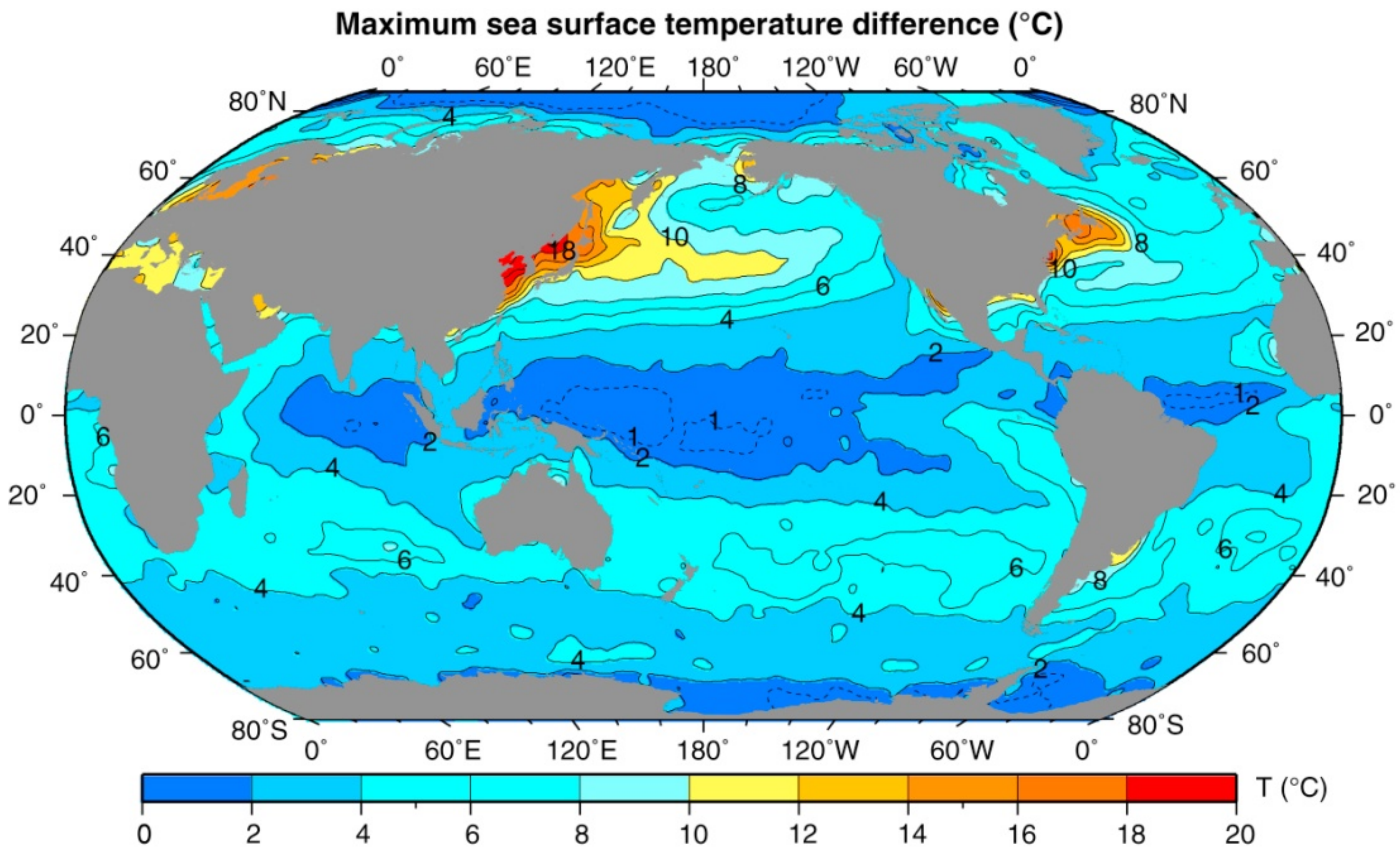


Satellite SST: NASA JPL http://podaac.jpl.nasa.gov/AnimationsImages/Animations

What we cover this week

- Atmosphere-Ocean interaction on seasonal timescale
  - Surface heating / cooling
  - Buoyancy flux
  - Mixed layer convection
- Vertical circulations in the ocean
  - Ventilation
  - Formation of thermocline and deep waters
  - Thermohaline circulation

Seasonality of SST



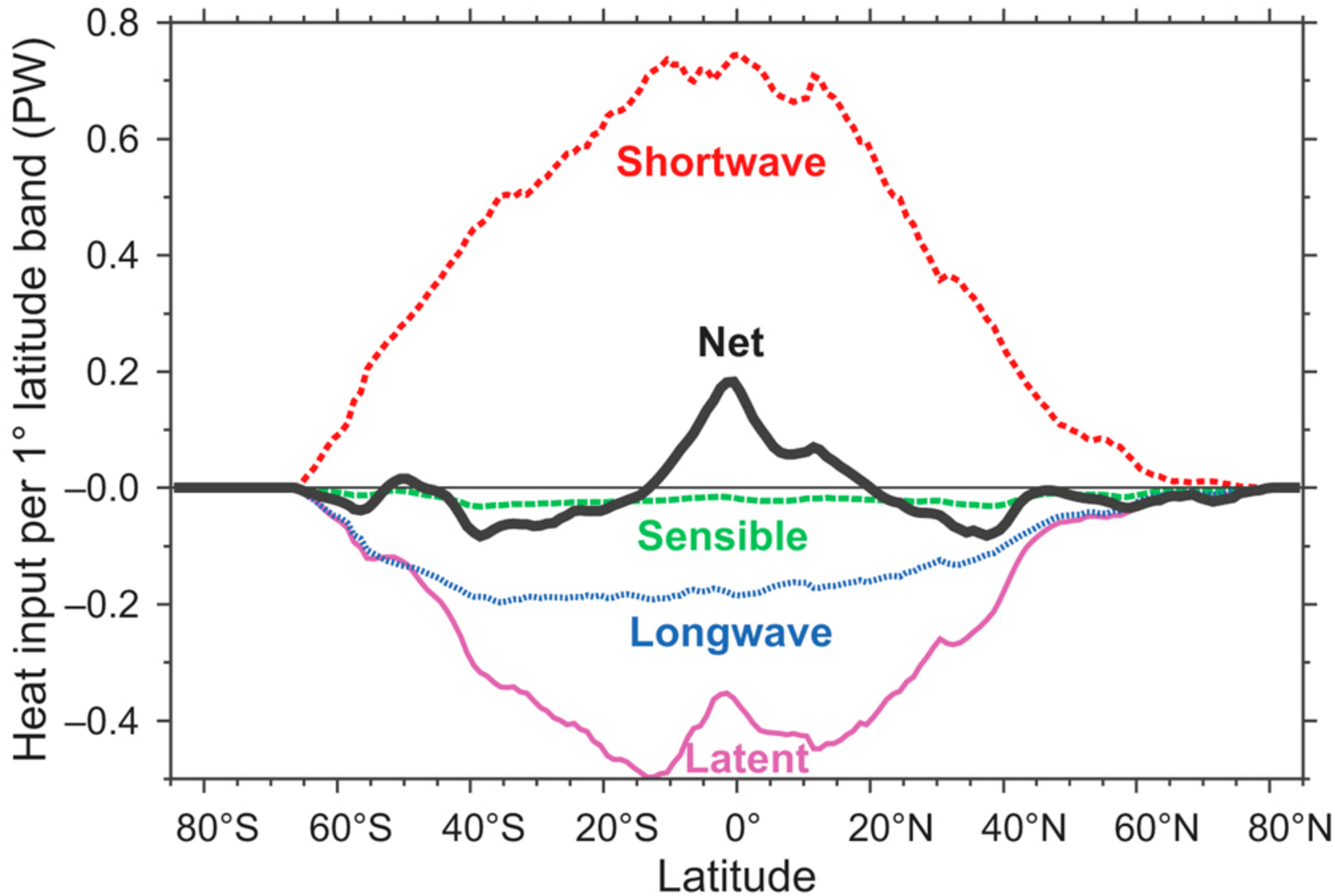
Annual range of sea surface temperature (°C), based on monthly climatological temperatures from the World Ocean Atlas (WOA05) (NODC, 2005a, 2009).

Air-sea exchange of heat

Heat flux components: [W/m^2]

- Short-wave radiation
- Long-wave radiation
- Sensible heat flux
- Latent heat flux

Annual mean, zonally averaged surface heat fluxes



Seasonality of net heat flux

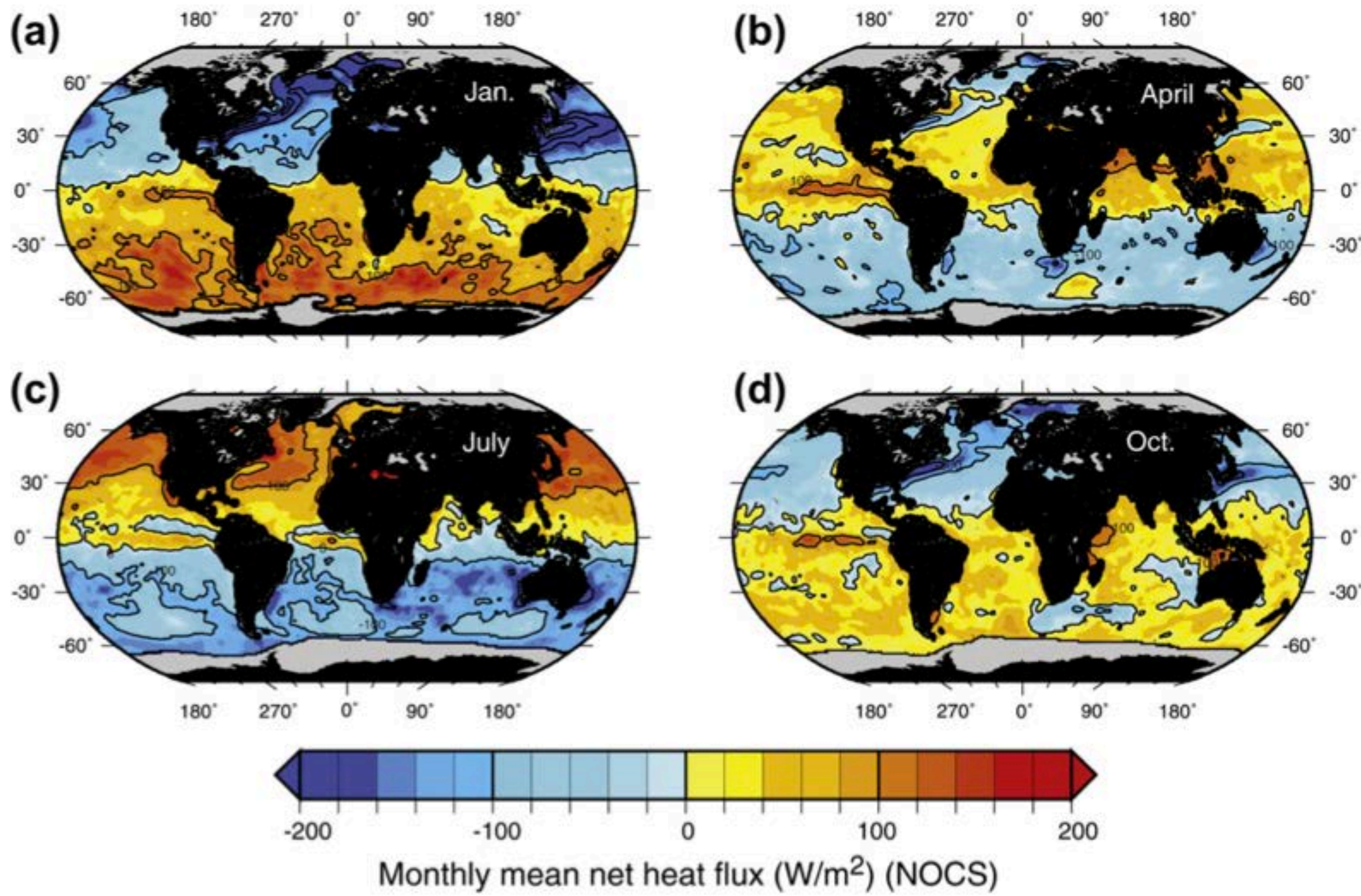


FIGURE S5.7 Monthly mean net heat flux (W/m^2) for (a) January, (b) April, (c) July, and (d) October. Data are from the NOCS product of Grist and Josey (2003).

Air-sea heat flux

- **Short wave radiation** is the dominant source of heat among the four components of air-sea heat fluxes. The strongest heating occurs in the summer. It is also a function of cloud cover.

- **Latent heat flux** is the largest heat sink for the surface water. It is proportional to the rate of evaporation. The strongest cooling occurs in the winter.
  - Evaporation rate depends on the difference between SST and air temperature directly above the sea surface, wind speed, and relative humidity.

- **The net heat flux** shows a strong seasonality and its amplitude generally increases with latitude.

Thermodynamic equation

- Heat balance
  - $\mathcal{H}$: heat flux (positive into the ocean)

$$\rho_0 c_P \left(\frac{\partial T}{\partial t} + u \cdot \nabla T \right) = \frac{\partial \mathcal{H}}{\partial z}$$

$$\frac{\partial T}{\partial t} + u \cdot \nabla T = \frac{1}{\rho_0 c_P} \frac{\partial \mathcal{H}}{\partial z}$$

ρ_0 : reference density ~ 1025 kgm^{-3}
c_P : specific heat of seawater ~ 3900 $Jkg^{-1}K^{-1}$

Salinity

- Seasonal cycle of S reflects the imbalance evaporation and precipitation
- Consider conservation of salt in the mixed layer
  - For unit area:

$$Mass : M = \rho_0\, h \qquad \delta M = \rho_0 (P - E)$$

$$Salt : I = S\, M \qquad \delta I = 0$$

- Continuity equation for salinity

$$\frac{\partial S}{\partial t} + u \cdot \nabla S = \frac{S}{h}(E - P)$$

Air-sea exchange of buoyancy

- T and S together control the density of s.w.
- What is buoyancy?

$$b = -\frac{g\rho}{\rho_0}$$

- Seasonal cycle of T and S leads to density changes → affects dynamics: mixing and circulation

Air-sea buoyancy flux

- Combine temperature and salinity equations to write down the equation for buoyancy

$$\delta b = -\frac{g}{\rho_0}\delta\rho = g\left(\alpha\delta T - \beta\delta S\right)$$

$$\frac{\partial b}{\partial t} + u \cdot \nabla b = \frac{\partial \mathcal{B}}{\partial z}$$

$$\mathcal{B} = \frac{\alpha g}{\rho_0 c_P}\mathcal{H} + \beta g S(P - E)$$

T, S and buoyancy flux (annual mean)

b

T

S

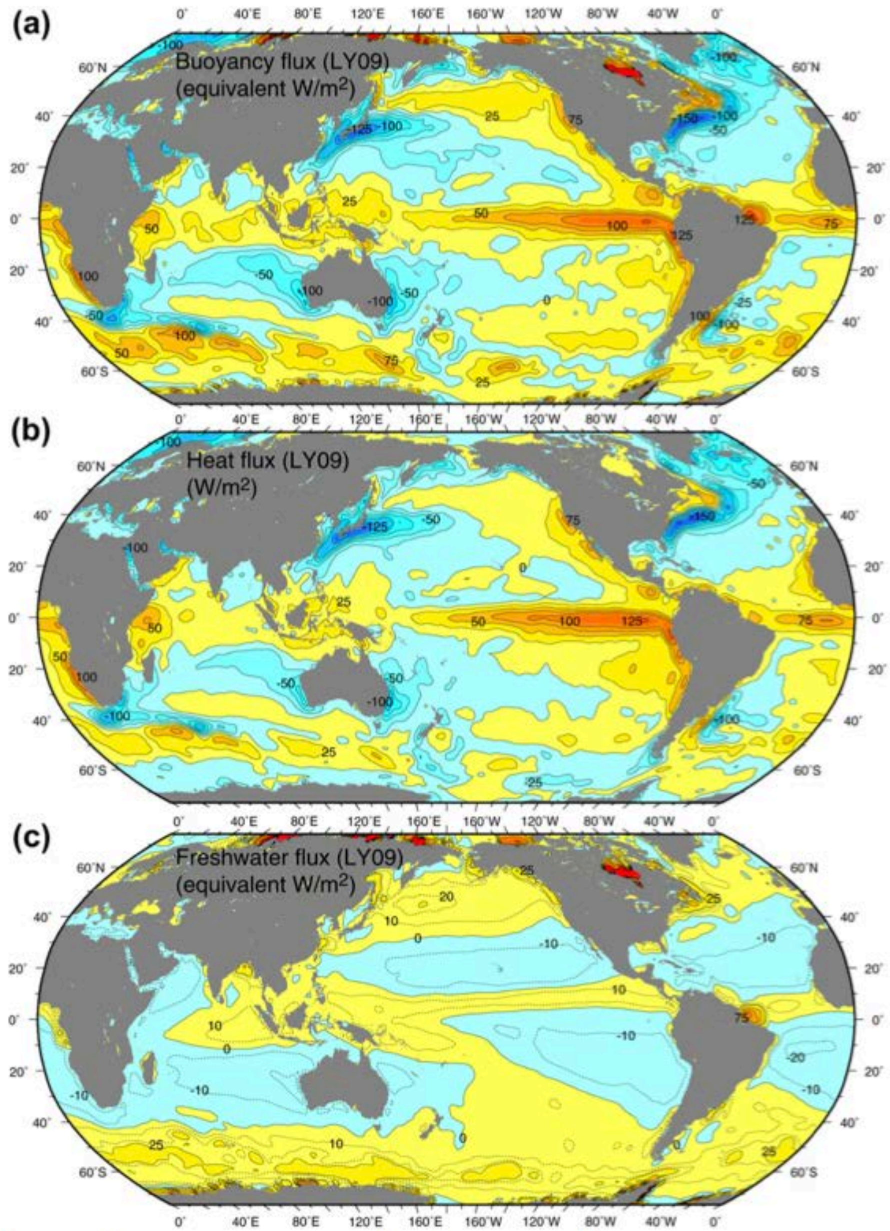


FIGURE S5.8 Annual mean air–sea (a) buoyancy flux, (b) heat flux, and (c) freshwater flux (precipitation, evaporation, and runoff) with the buoyancy and freshwater fluxes converted to equivalent heat fluxes (W/m^2), based on Large and Yeager (2009) air–sea fluxes. Positive values (yellows-reds) indicate that the ocean is becoming less dense, warmer, or fresher in the respective maps. Contour interval is 25 W/m^2; in (c) dotted contours are 10 and 20 W/m^2.

Air-sea exchange of momentum

- Wind stress is what drives the ocean (waves, currents).
- Wind speed is measured above the sea surface. In practice measurements are at different heights. For consistency, adjust all measurements to a 10 meter height.

- Actual stress on the ocean: Units of wind stress: N/m^2

$$\tau = c_D \rho u_{10}^2$$

where u_{10} is the wind speed at 10 meters, ρ is the air density 1.3 kg/m^3, and c_D is the (dimensionless) drag coefficient, which is determined empirically.

Drag coefficient

Again, wind stress is $\tau = \rho c_D u_{10}^2$ in units of N/m^2

where u_{10} = wind speed at 10 m height
ρ is air density 1.3 kg/m^3
c_D is a "drag coefficient" (dimensionless)

At low wind speeds, $c_D \approx 1.1 \times 10^{-3}$

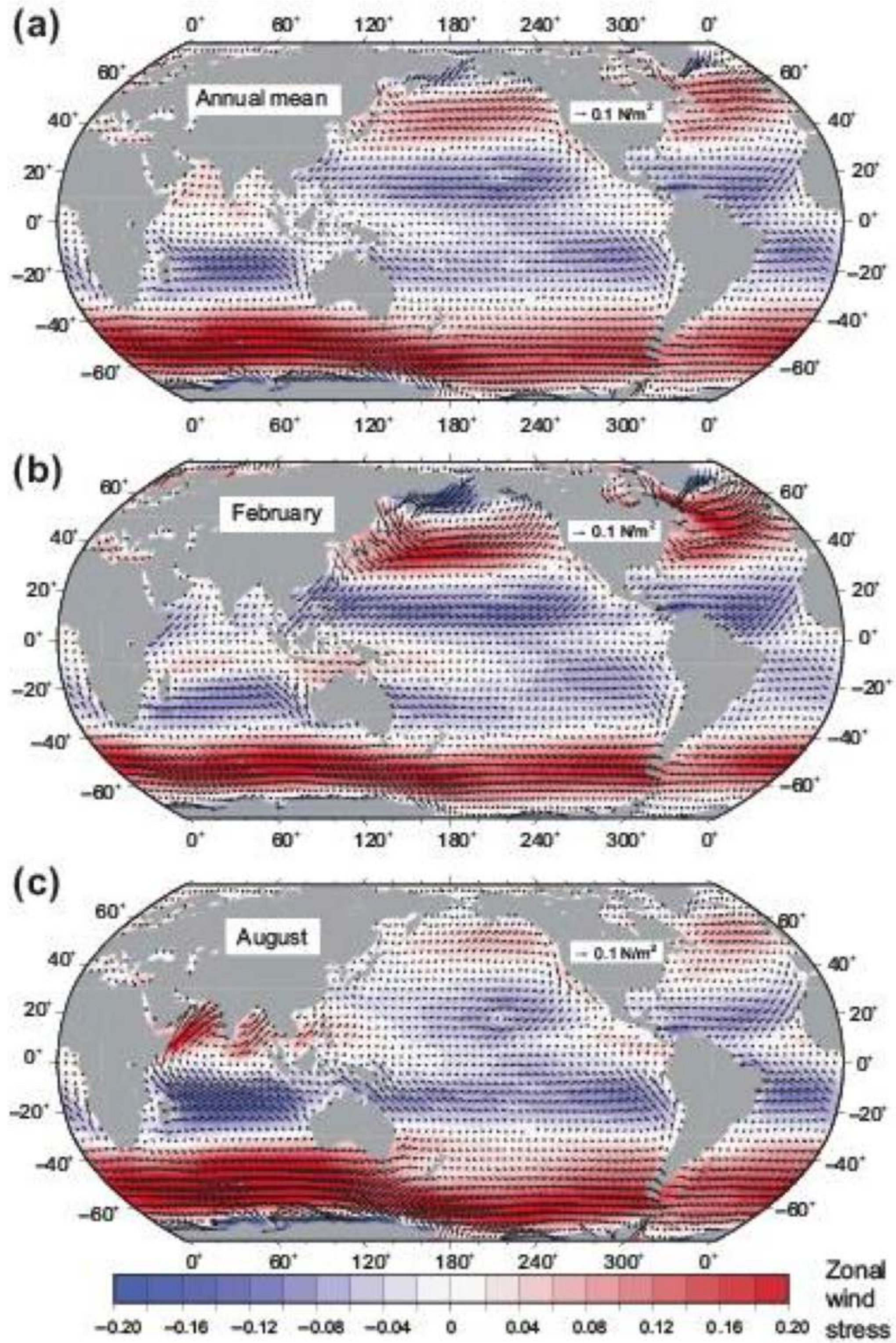
At higher wind speeds, $c_D \approx (0.61 + 0.063\ u) \times 10^{-3}$

Fairall et al. (2003, J. Climate)

Annual mean

NH winter

NH summer



What regulates the seasonality in the oceans?

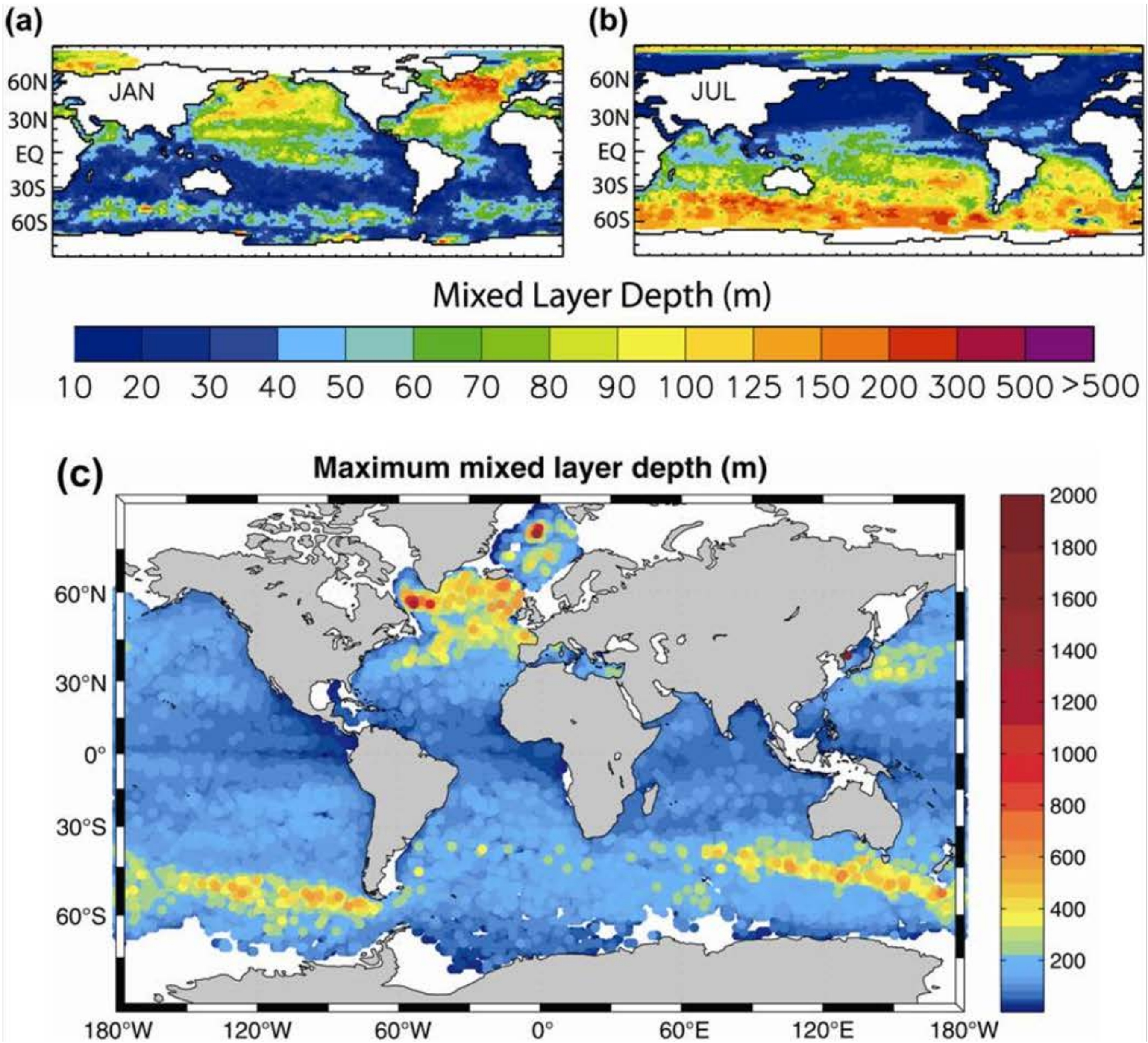
- Heat flux
  - SW radiation in the summer and latent heat loss in the winter
  - Temperature and salinity controls the density (buoyancy) flux, which is primarily temperature-driven

- Wind stress
  - The wind stress is generally stronger in the winter seasons
  - Monsoon wind is driven by the seasonal reversal of land-ocean temperature gradient

Seasonality of the surface mixed layer

NH winter

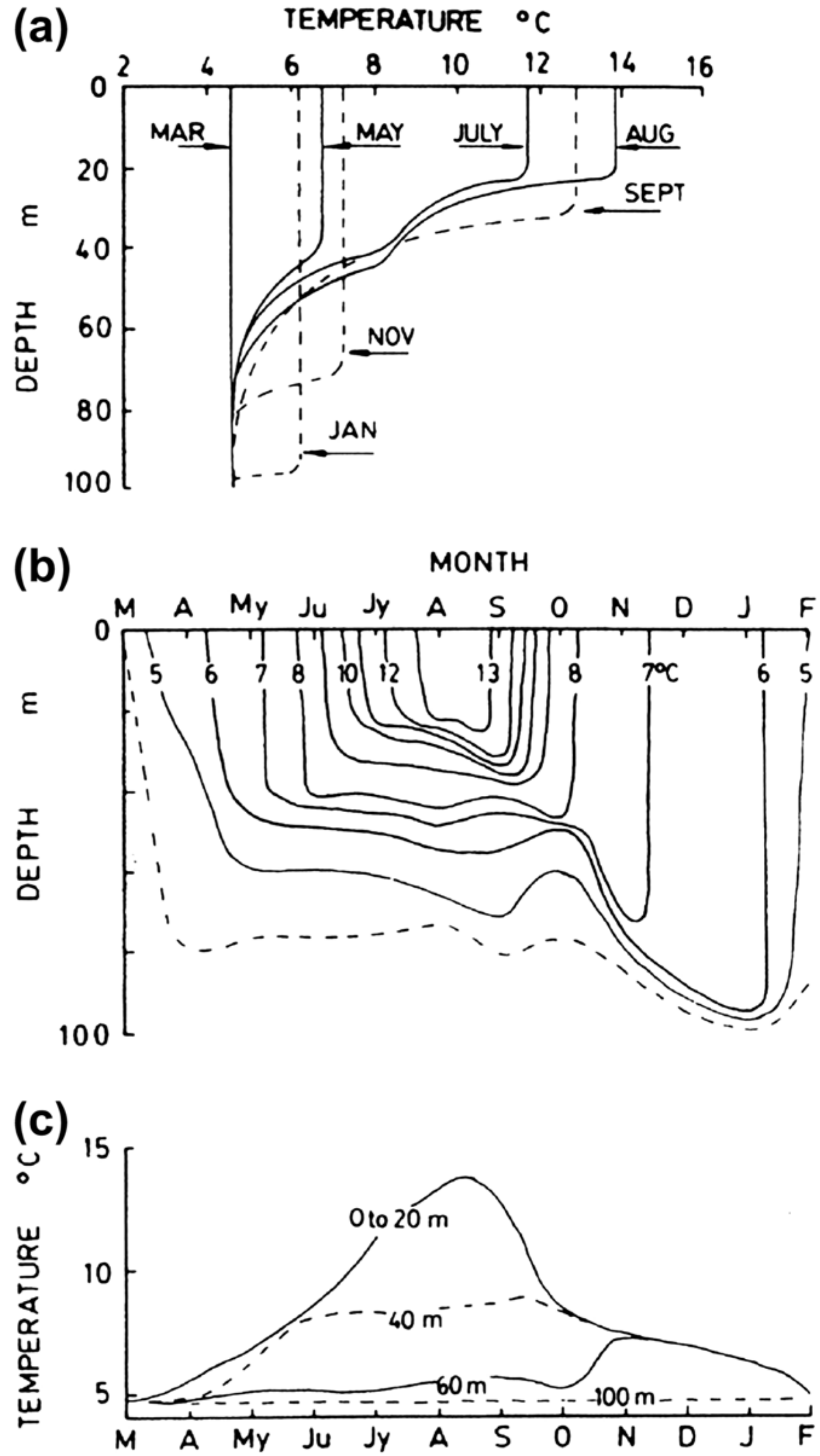
NH summer



Vertical structure of seasonal T change

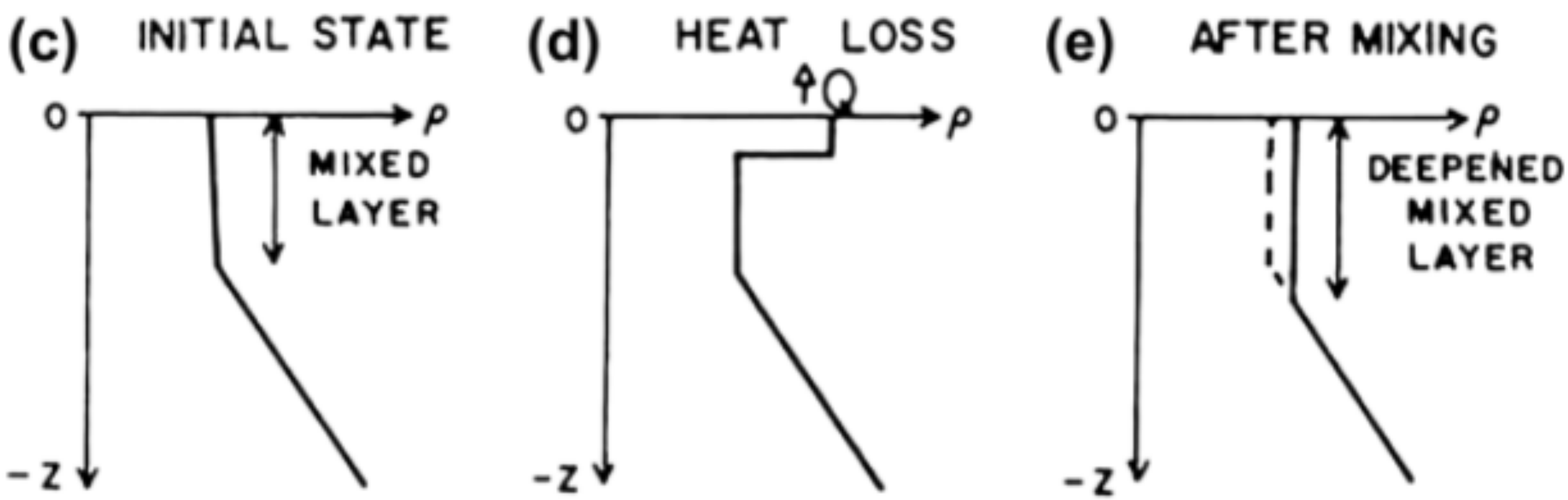
Observed in the subpolar North Pacific

(a) Vertical temperature profiles
(b) Time series of isothermal contours, and
(c) Time series of temperatures at depths shown.



What controls the mixed layer depth?

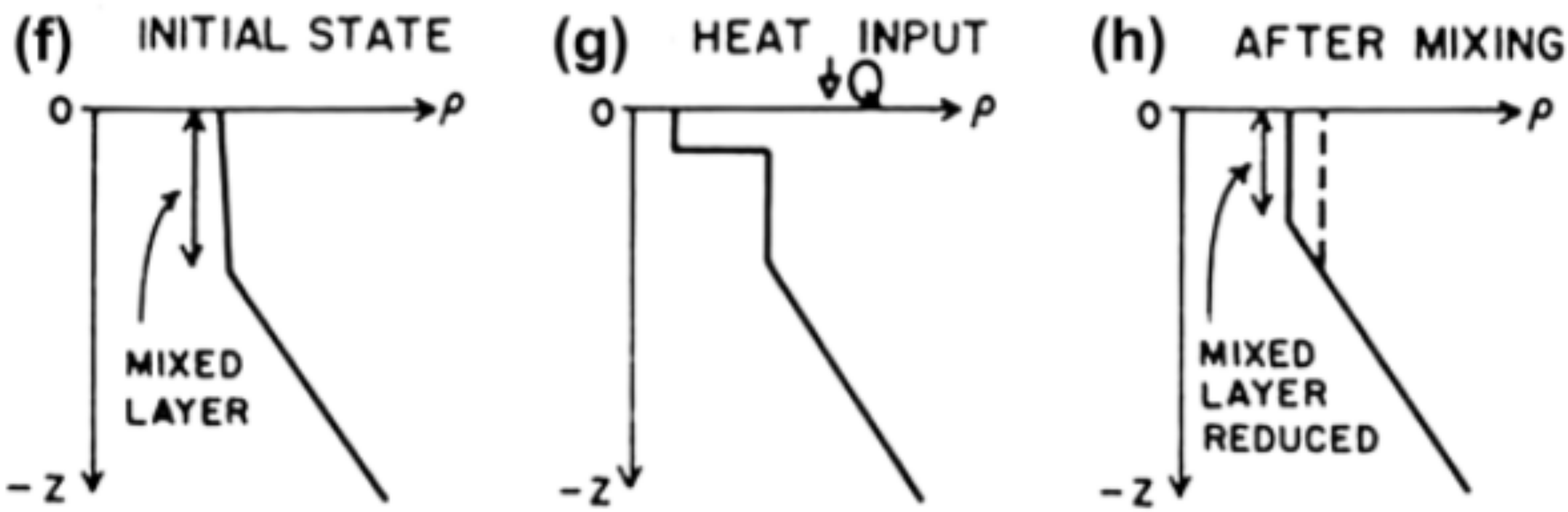
- Cooling → Makes surface water dense → Dense water sinks/mixes with deeper layer → Deepens the mixed layer



The vertically integrated density increase (buoyancy decrease) balances the net heat loss at the surface

What controls the mixed layer depth?

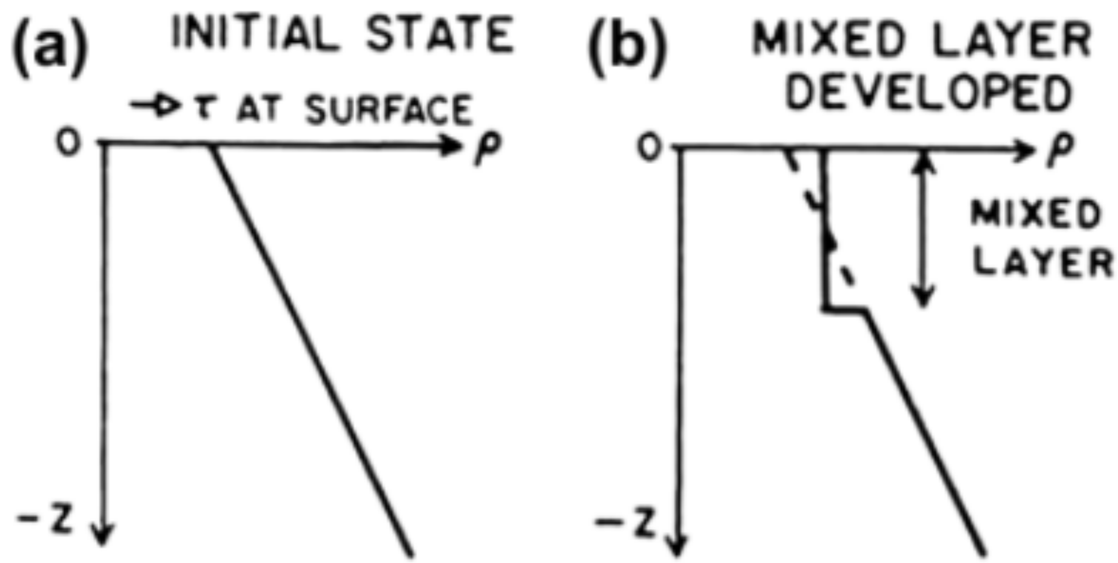
- Heating → Makes surface water light → Shoals the mixed layer



The vertically integrated density decrease (buoyancy increase) balances the net heat gain at the surface

Wind driven mixing

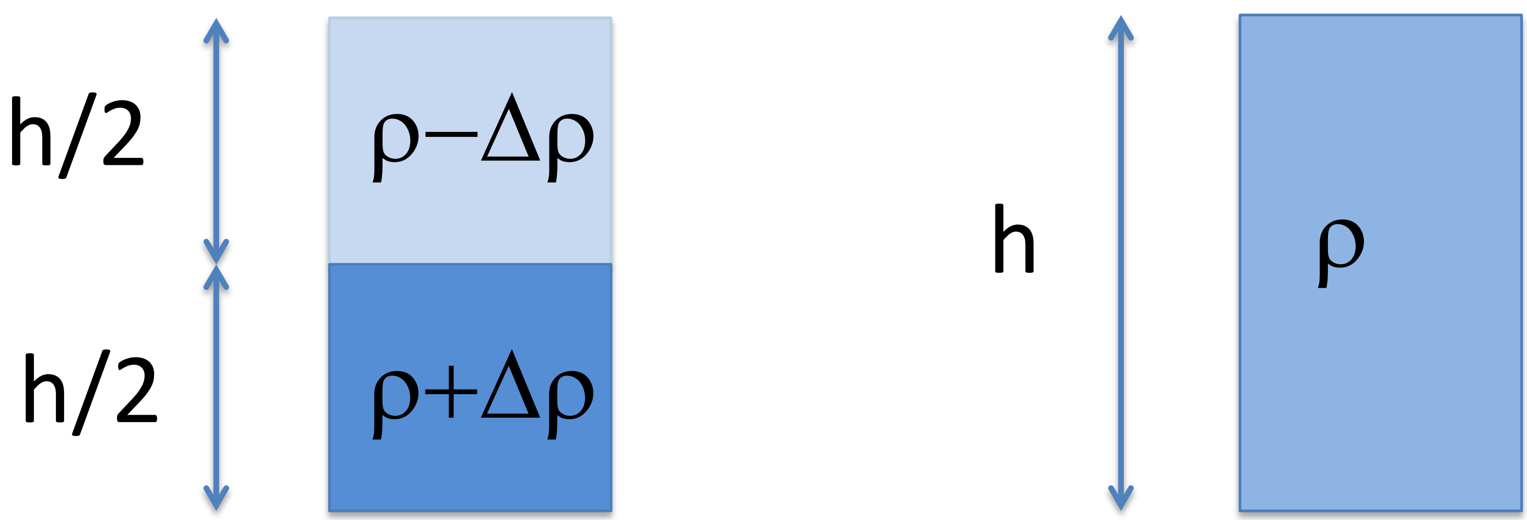
- Wind stress → Intensify the surface currents → Intensify the near-surface turbulence → Deepens the mixed layer



Vertically integrated buoyancy is conserved in this scenario

Energetics of mixed layer convection

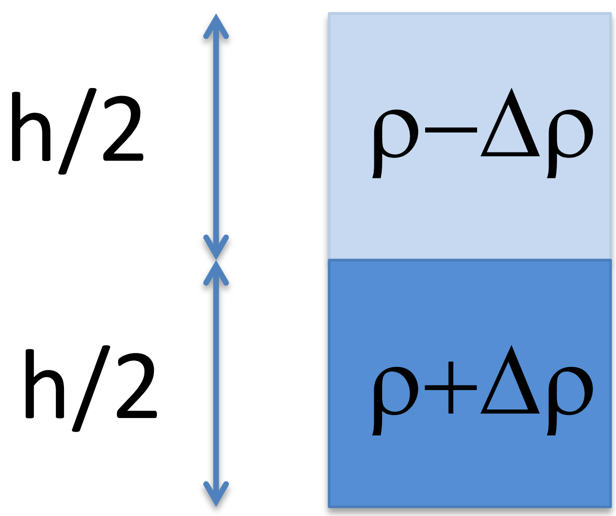
- Potential energy (per unit volume) = $\rho g z$
- Two cases:
  - 1. Well-mixed with thickness h
  - 2. Stratified ($\Delta\rho$) with equal thickness h/2



Energetics of mixed layer convection

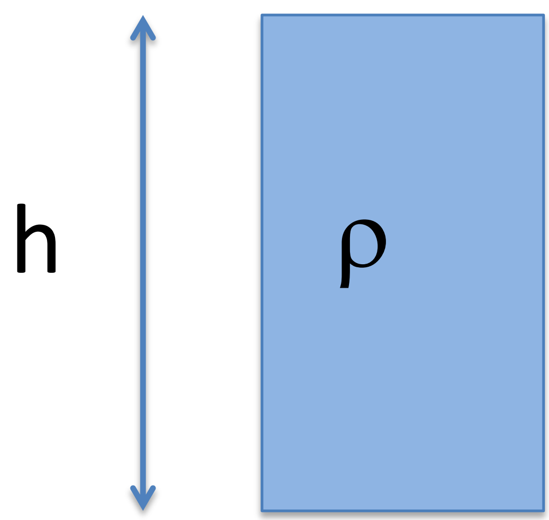
(2) Stratified state

$$PE_2 = \int_0^h \rho(z) g z\, dz$$
$$= \frac{-1}{2}\rho g h^2 - \frac{1}{4}\Delta\rho g h^2$$



(1) Well-mixed state

$$PE_1 = \int_0^h \rho g z\, dz$$
$$= \frac{-1}{2}\rho g h^2$$



Well-mixed state has higher potential energy. Transition from (2) to (1) requires energy input.

Energetics of mixed layer convection

- **Mixing the stratified fluid requires energy input**
  - This compensates the potential energy increase
  - Wind stress (kinetic energy source ➔ conversion to the potential energy)
  - Cooling (making surface layer heavy ➔ potential energy gain)
  - Heating (making surface layer light ➔ potential energy loss)

$$\Lambda \Delta b h \frac{\partial h}{\partial t} = 2 m u_*^3 - \mathcal{B} h$$

$$\Lambda = \begin{cases} 0 & \text{if } \partial h / \partial t < 0 \\ 1 & \text{if } \partial h / \partial t > 0 \end{cases}$$

(Potential energy gain due to deepening of ML)

(Wind kinetic energy input)

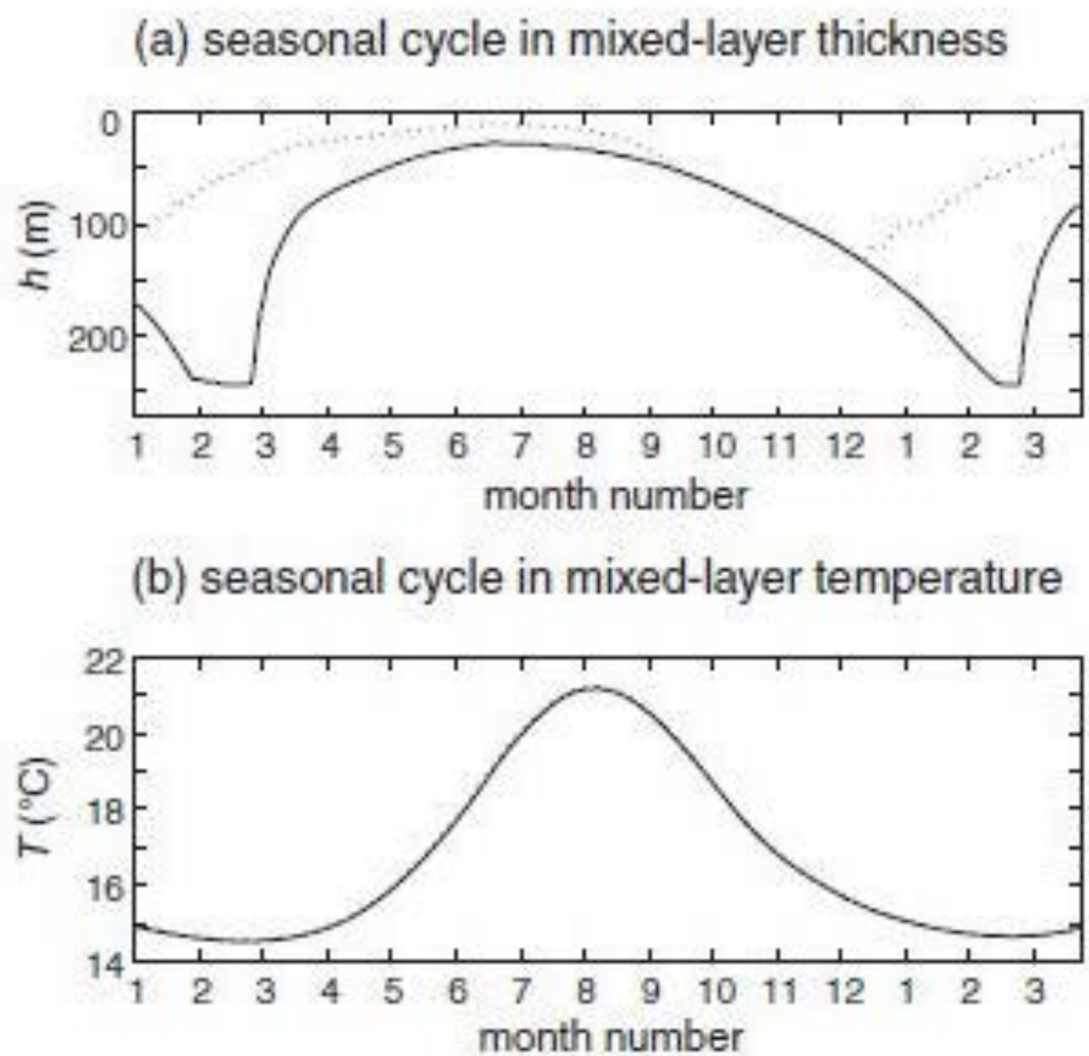
(Potential energy input due to buoyancy loss)

A bulk mixed layer model

- Klaus and Turner (1967)

$$\Lambda \Delta b h \frac{\partial h}{\partial t} = 2 m u_*^3 - \mathcal{B} h$$

Where



Seasonal asymmetry

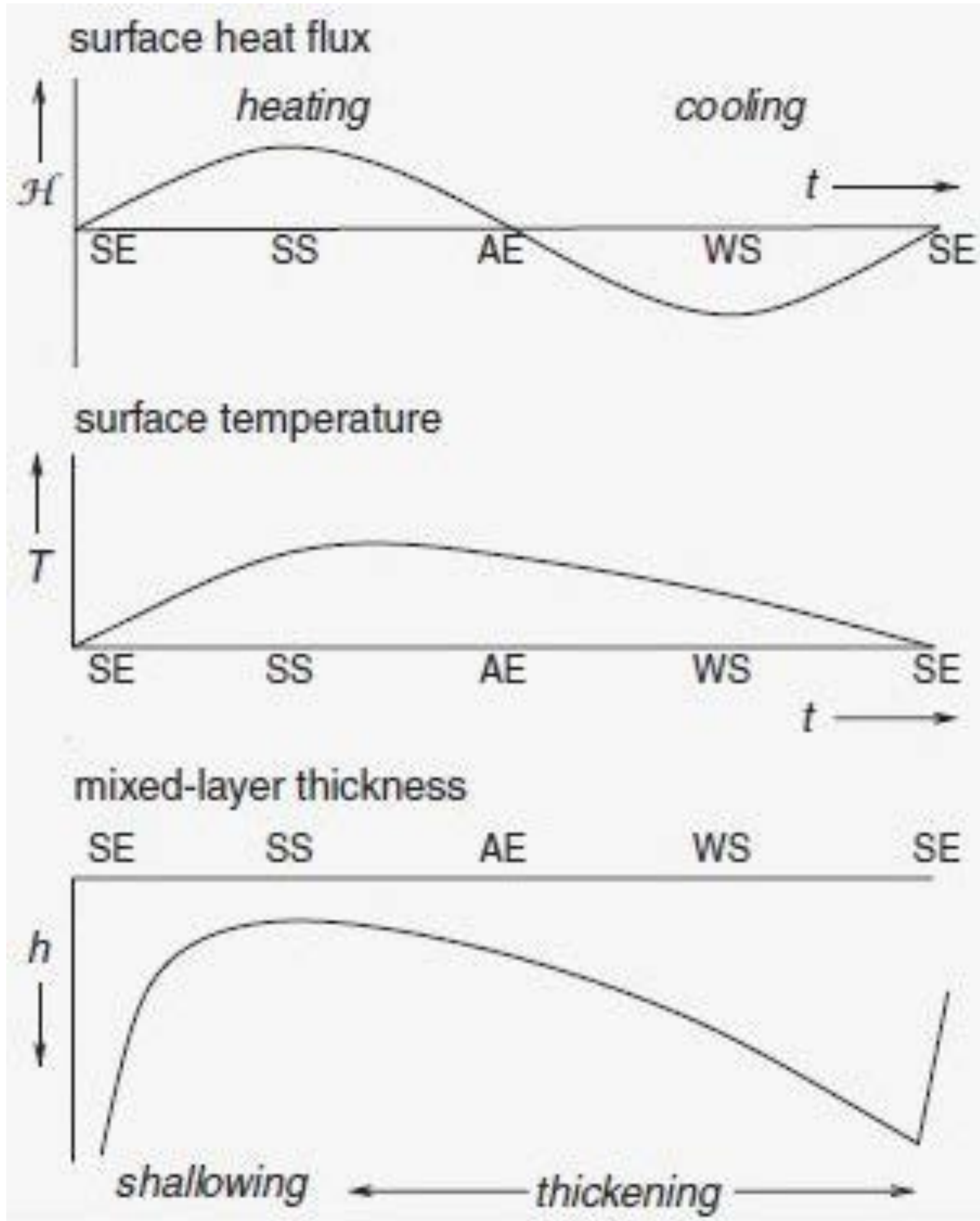
- SST integrates the heat flux
  - Max SST lags behind the maximum heating
  - Min SST also lags behind the max cooling
- MLD
  - Rapid shoaling in spring
    - Heating from the top during spring and summer
    - Heating occurs only the top thin layer
  - Gradual deepening
    - Mixing/Entrainment of subsurface water from the deeper layer

Heat flux, SST and MLD

Heat flux

SST

MLD



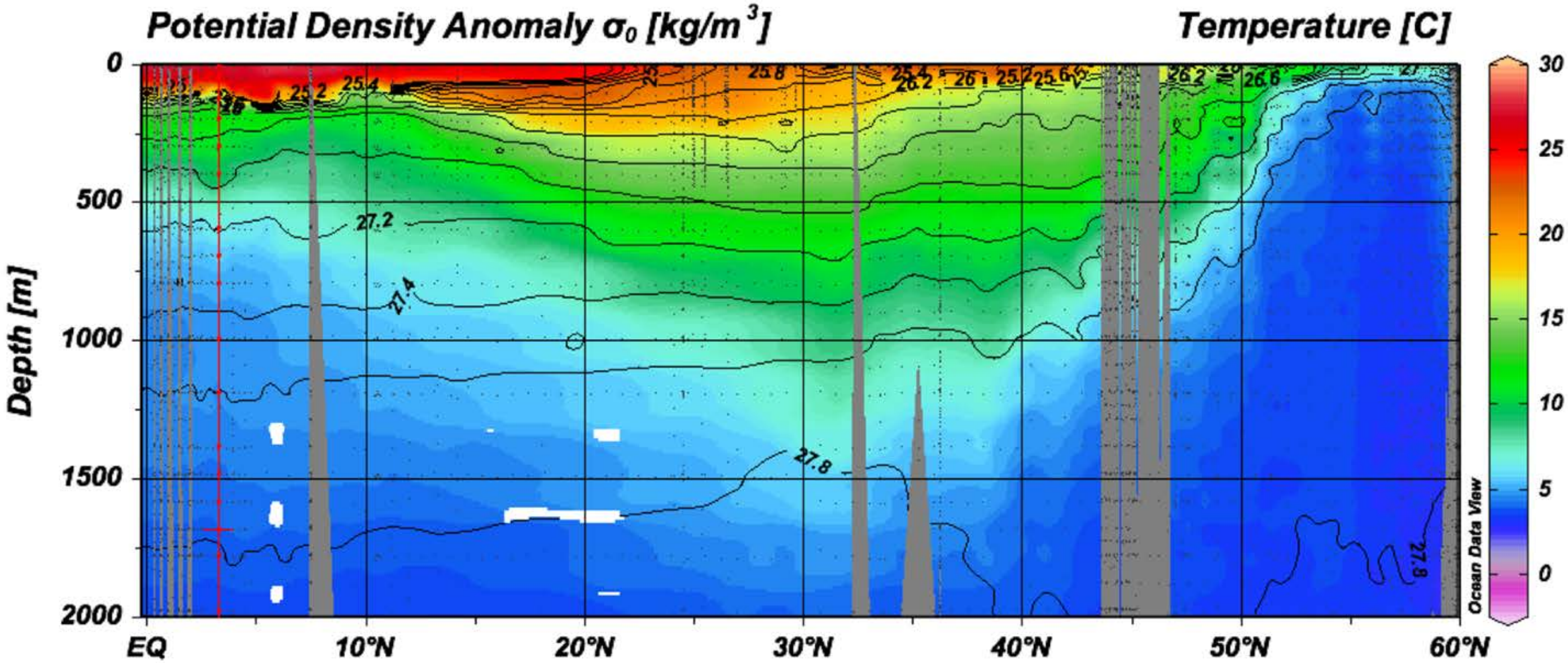
“Ventilation” and “Subduction”

“Ventilation” means that the surface mixed layer water sinks into the thermocline or deep ocean and becomes a subsurface water. **“Subduction”** essentially means the same with emphasis on the advective transfer of mass from the surface mixed layer to the interior ocean.

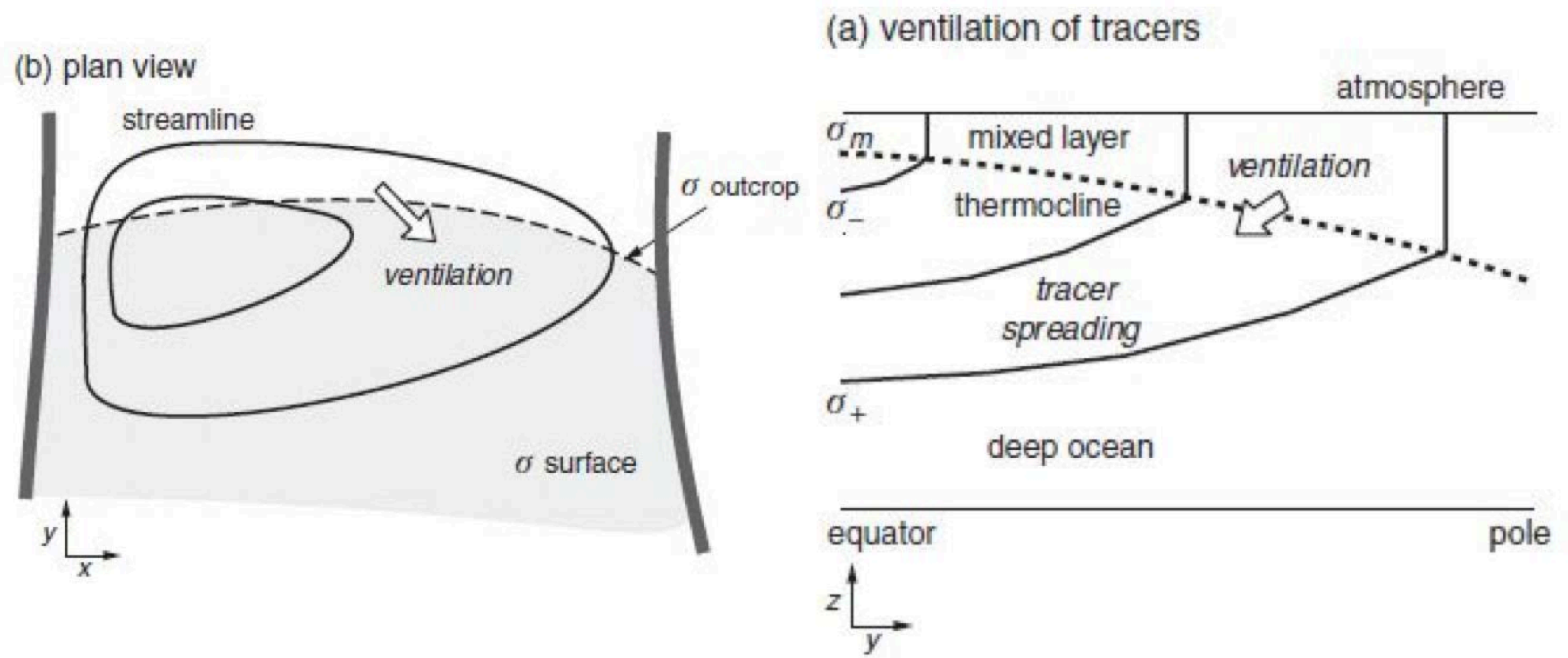
Ventilation of thermocline (mode water) happens in the mid-latitudes through subduction.

Ventilation of deep water happens in the polar oceans through deep convection and/or overflows.

Atlantic transect



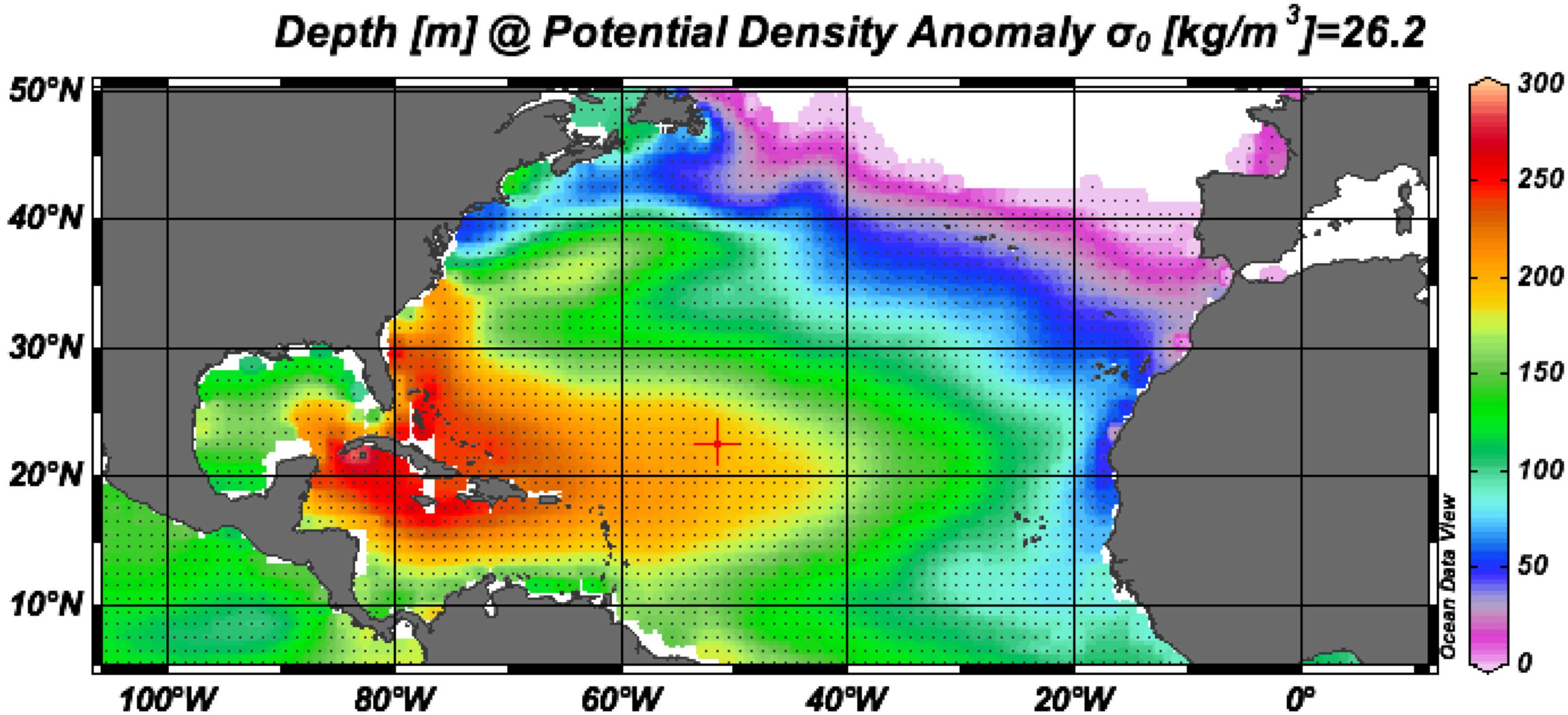
Mid-latitude thermocline ventilation



Isopycnals intersect the base of the winter-time mixed layer at the **outcrop** (ventilation happens here).

Subtropical Mode Water

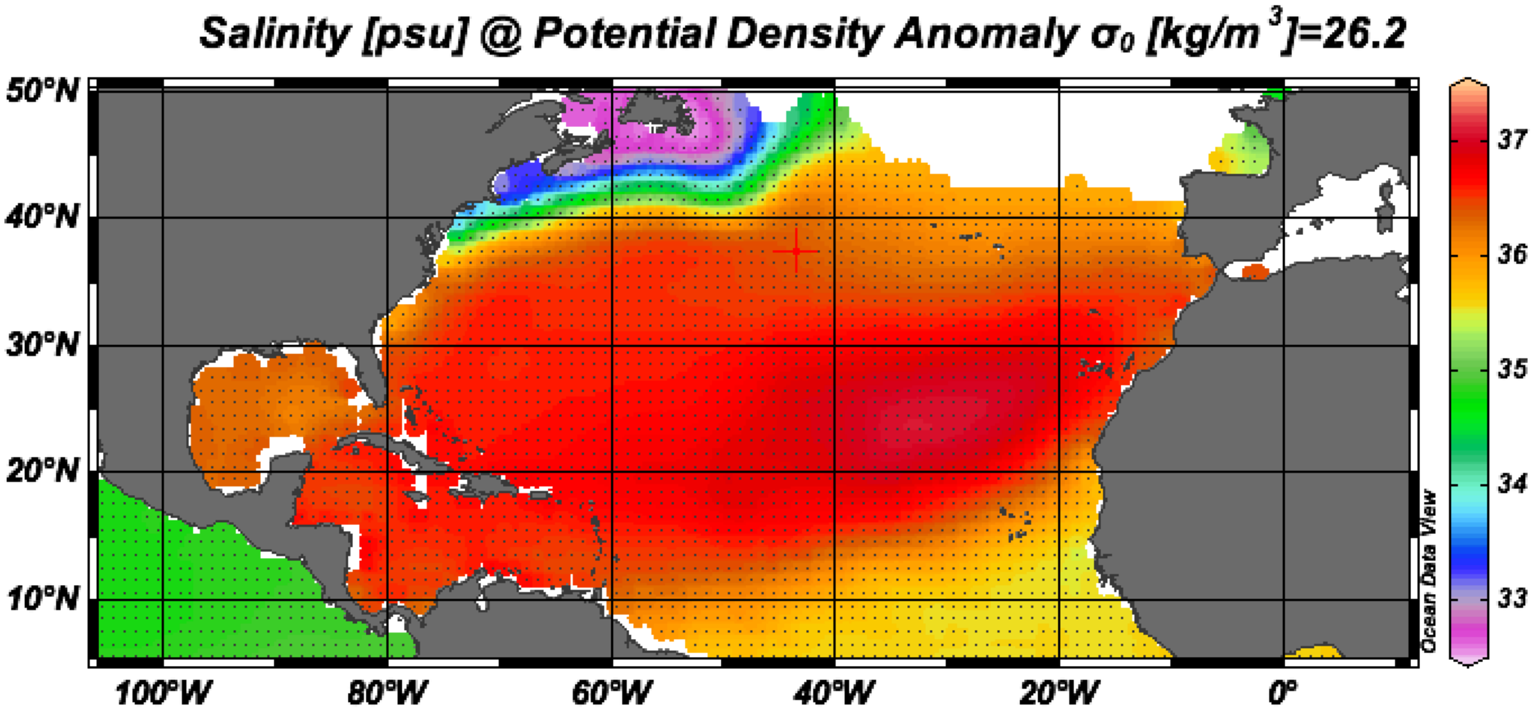
- Sigma-0 26.2: upper thermocline water



Eastward shoaling of the thermocline depth in the subtropical gyre

Subtropical Mode Water

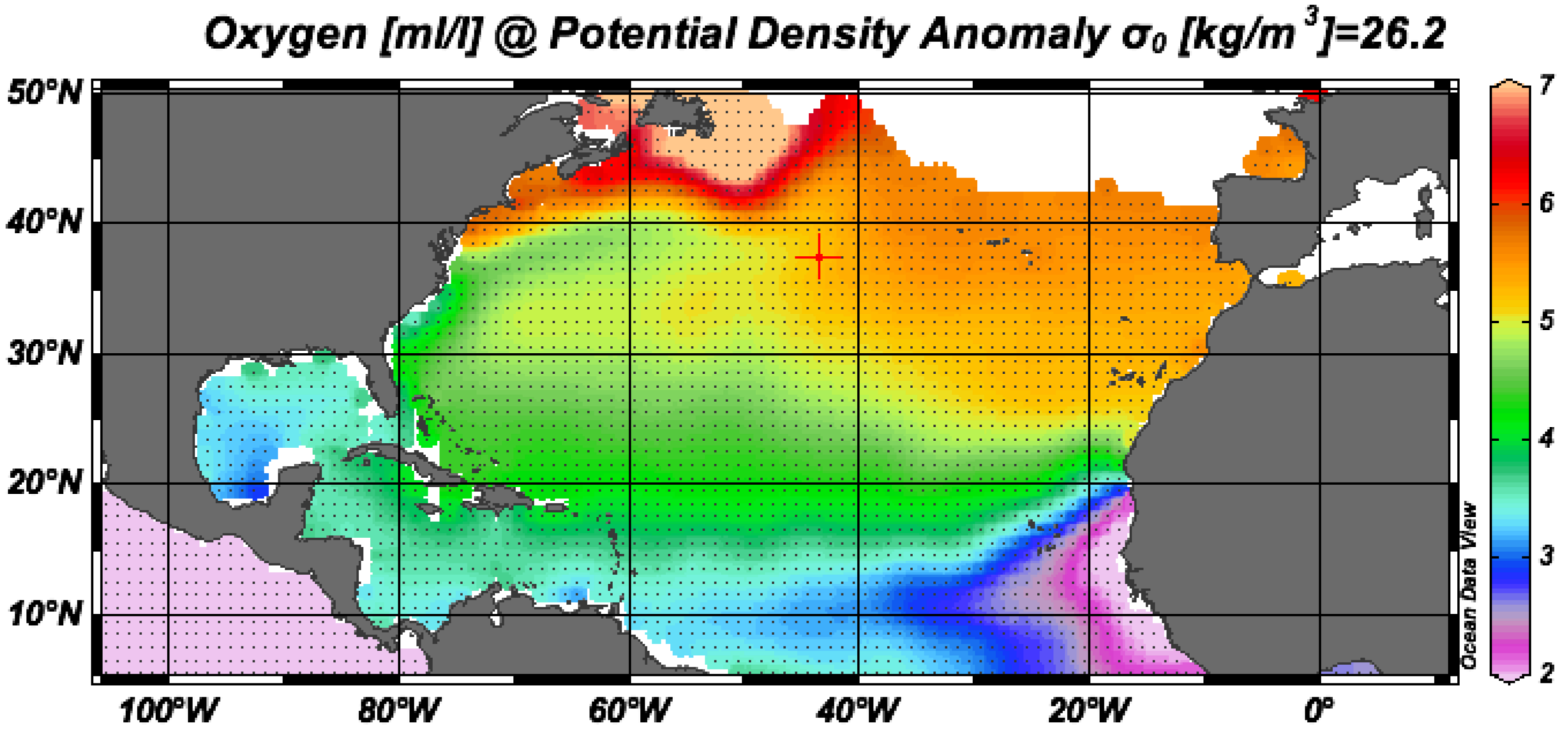
- Sigma-0 26.2: upper thermocline water



Following the density layer, (T,S) properties are nearly constant

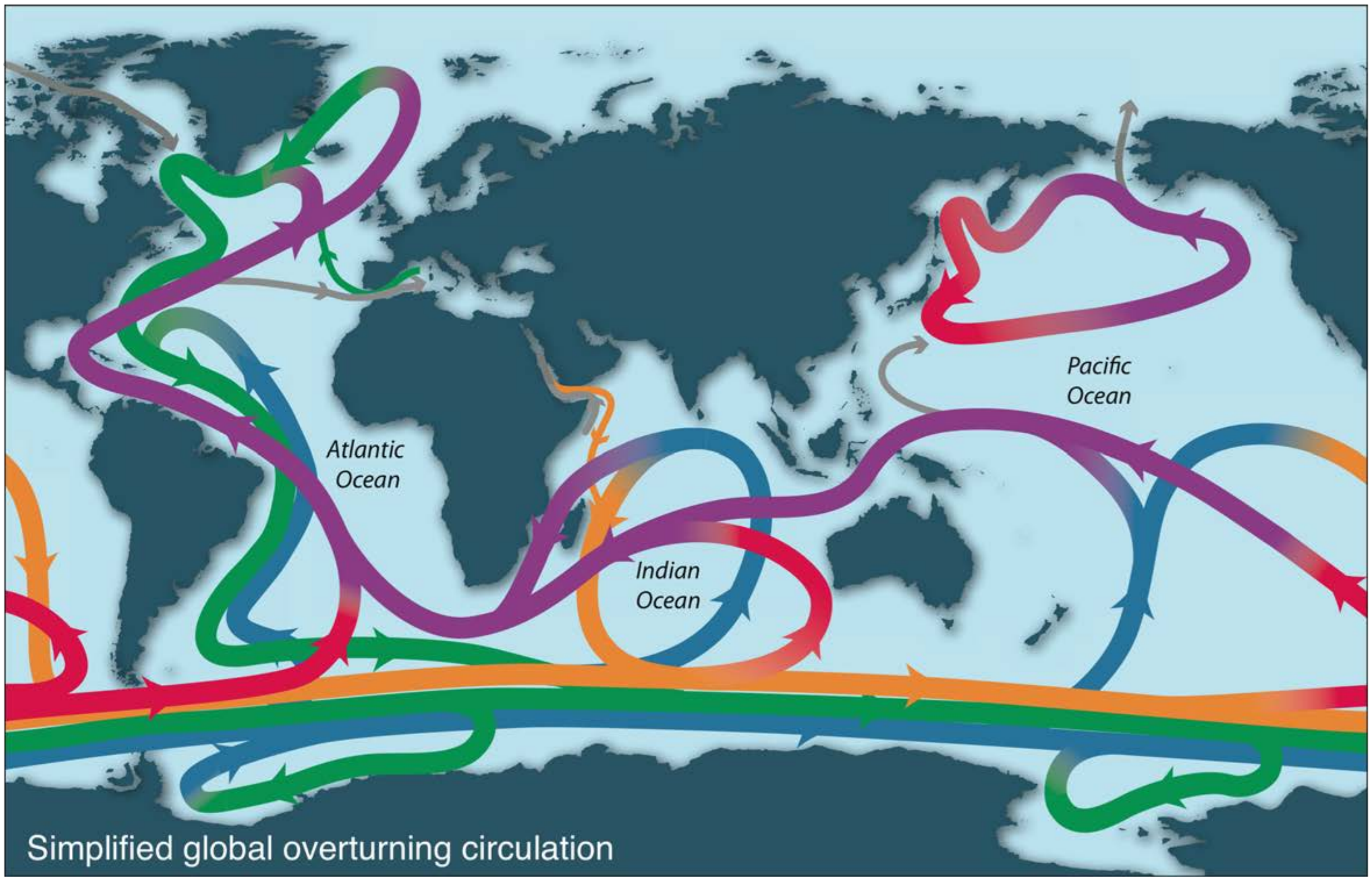
Subtropical Mode Water

- Sigma-0 26.2: upper thermocline water



Oxygen decreases away from the isopycnal outcrop

Subpolar circulation & meridional overturning circulation



Simplified global overturning circulation

Some terminologies

- **Thermohaline circulation** refers to large-scale overturning circulation, involving formation of dense, cold water and return of warm water to the sinking regions
- **Meridional Overturning Circulation** refers to essentially the same class of circulation, with emphasis on the north-south and vertical circulation, often abbreviated as “MOC”
- **Abyssal circulation** also refers to essentially the same circulation, with emphasis on the circulation of the deep sea.

Meridional overturning stream function

Remember mass continuity of the incompressible fluid: $\nabla \cdot \mathbf{u} = 0$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0$$

We zonally (east-west) integrate it.

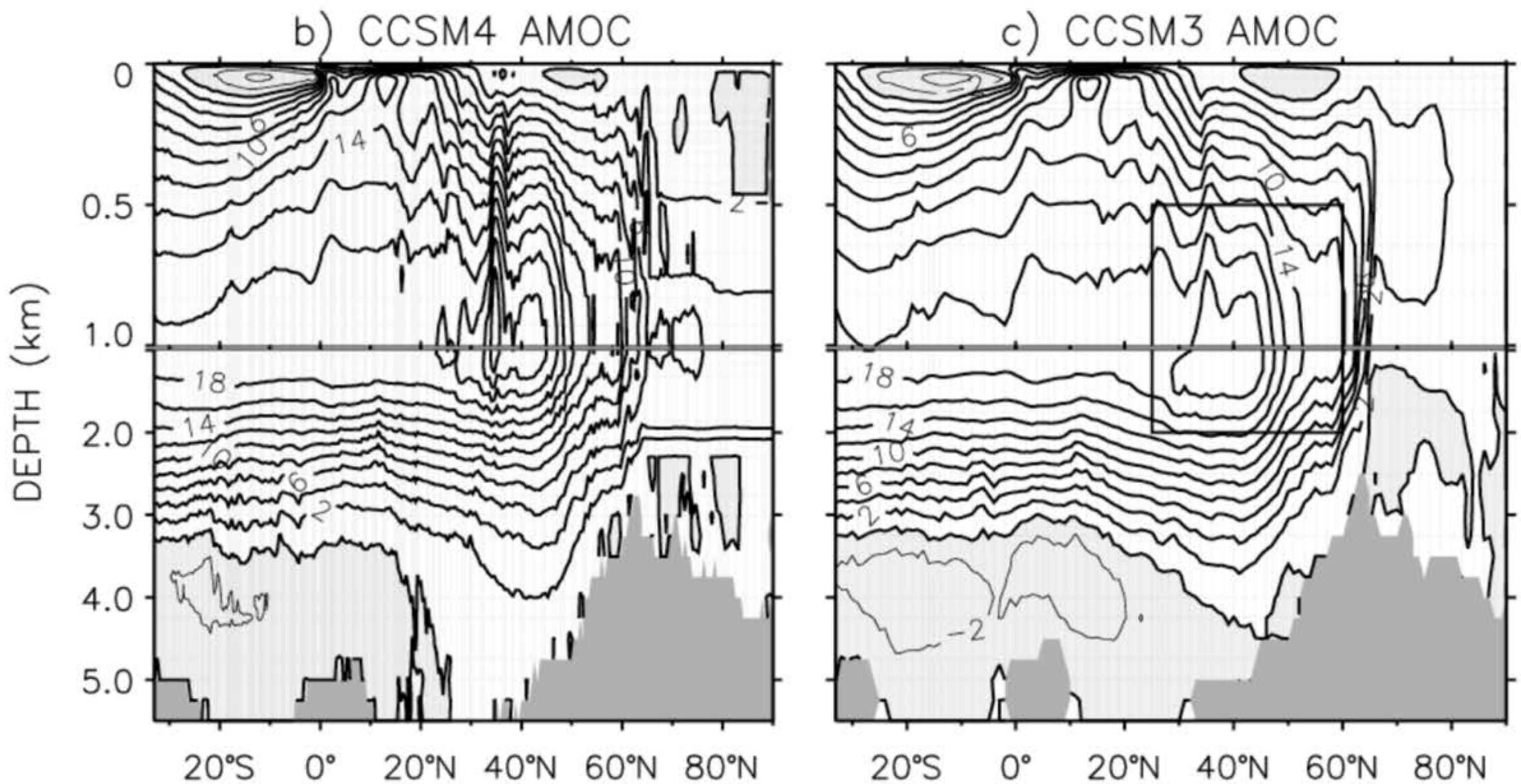
$$\frac{\partial V}{\partial y} + \frac{\partial W}{\partial z} = 0$$

We can define a meridional overturning stream function, Φ, so that the (V, W) will satisfy the mass continuity equation.

$$V = -\frac{\partial \Phi}{\partial z}, W = \frac{\partial \Phi}{\partial y}$$

Simulated meridional overturning stream function

Mean AMOC

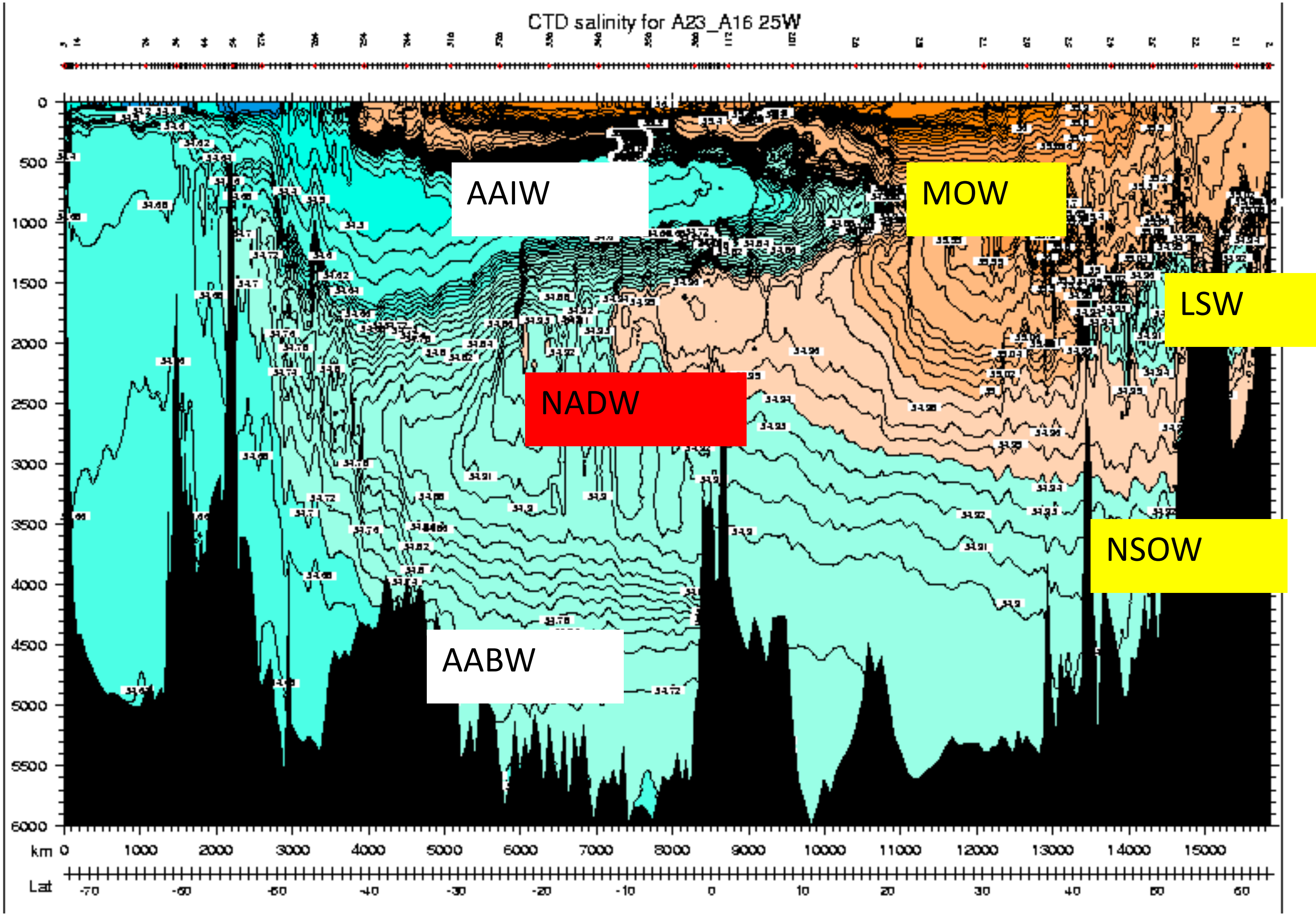


Danabasoglu et al. (2011, J. Climate)

North Atlantic MOCs

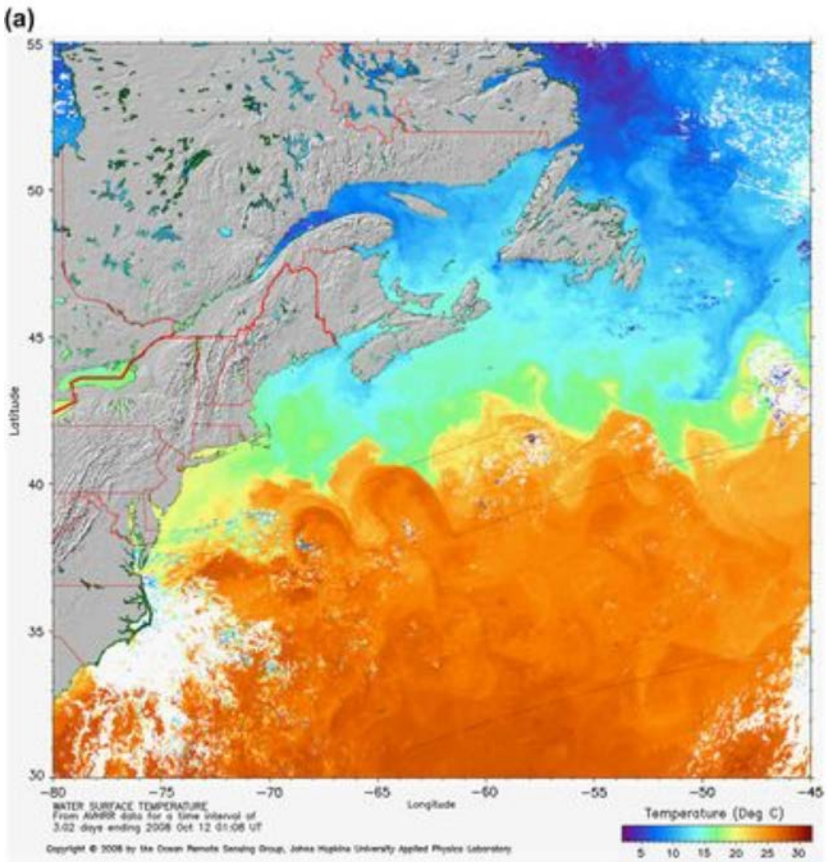
- Subtropical cells: upper ocean, shallow MOCs that involves sinking at mid-latitudes and upwelling in tropics.
- Atlantic Meridional Overturning Circulation (AMOC)
  - Sinking and export NADW from the N. Atlantic
- Antarctic Bottom Water (AABW) cell
  - Sinking in the polar Southern Ocean

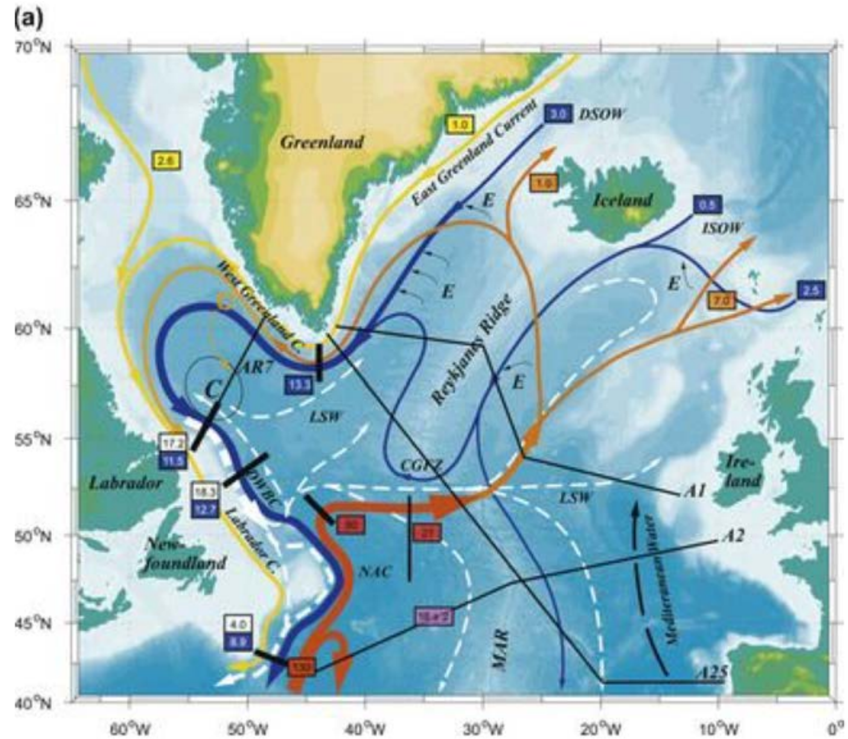
Atlantic Meridional Overturning Circulation (AMOC)



Subpolar North Atlantic

Gulf Stream becomes North Atlantic Current (NAC)



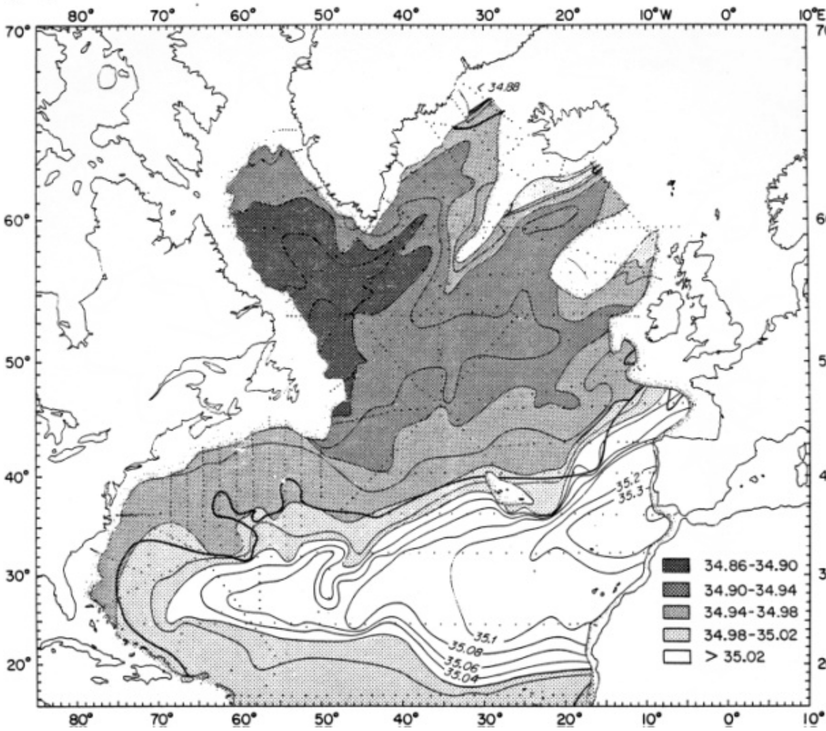


Gulf stream ~ 90 Sv becomes NAC and much of it recirculates back to the south as the subtropical gyre circulation.

Poleward flowing NAC feeds the warm water to the sinking region ~ 15-20 Sv

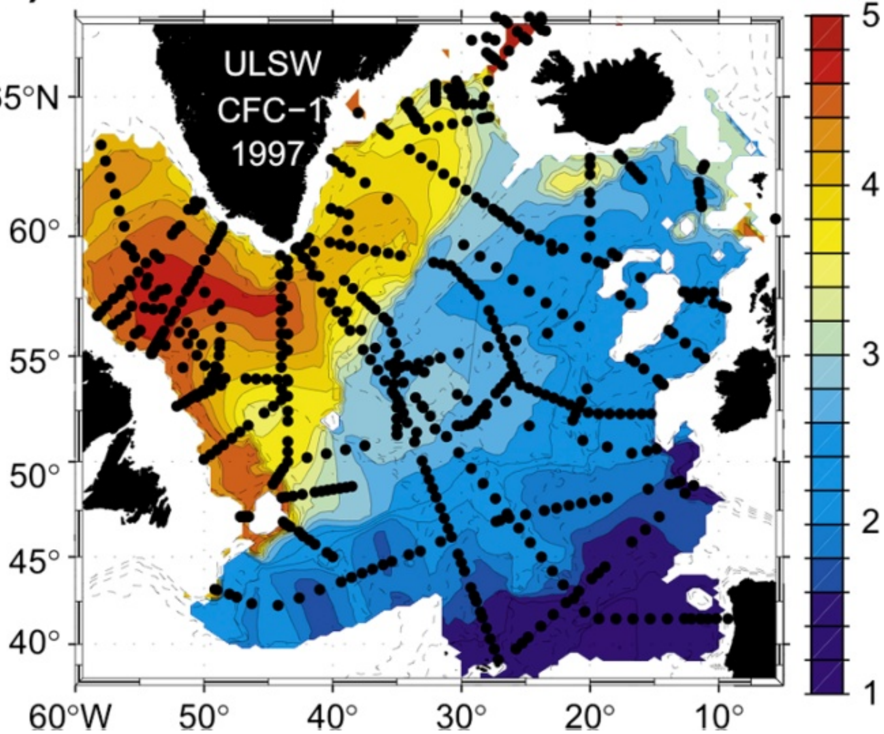
NADW source: Labrador Sea Water

(a)



Salinity on isopycnal typical of LSW: fresh in Labrador Sea

(b)



Chlorofluorocarbons in the Labrador Sea Water, highs indicating recent ventilation

Labrador Sea hydrography

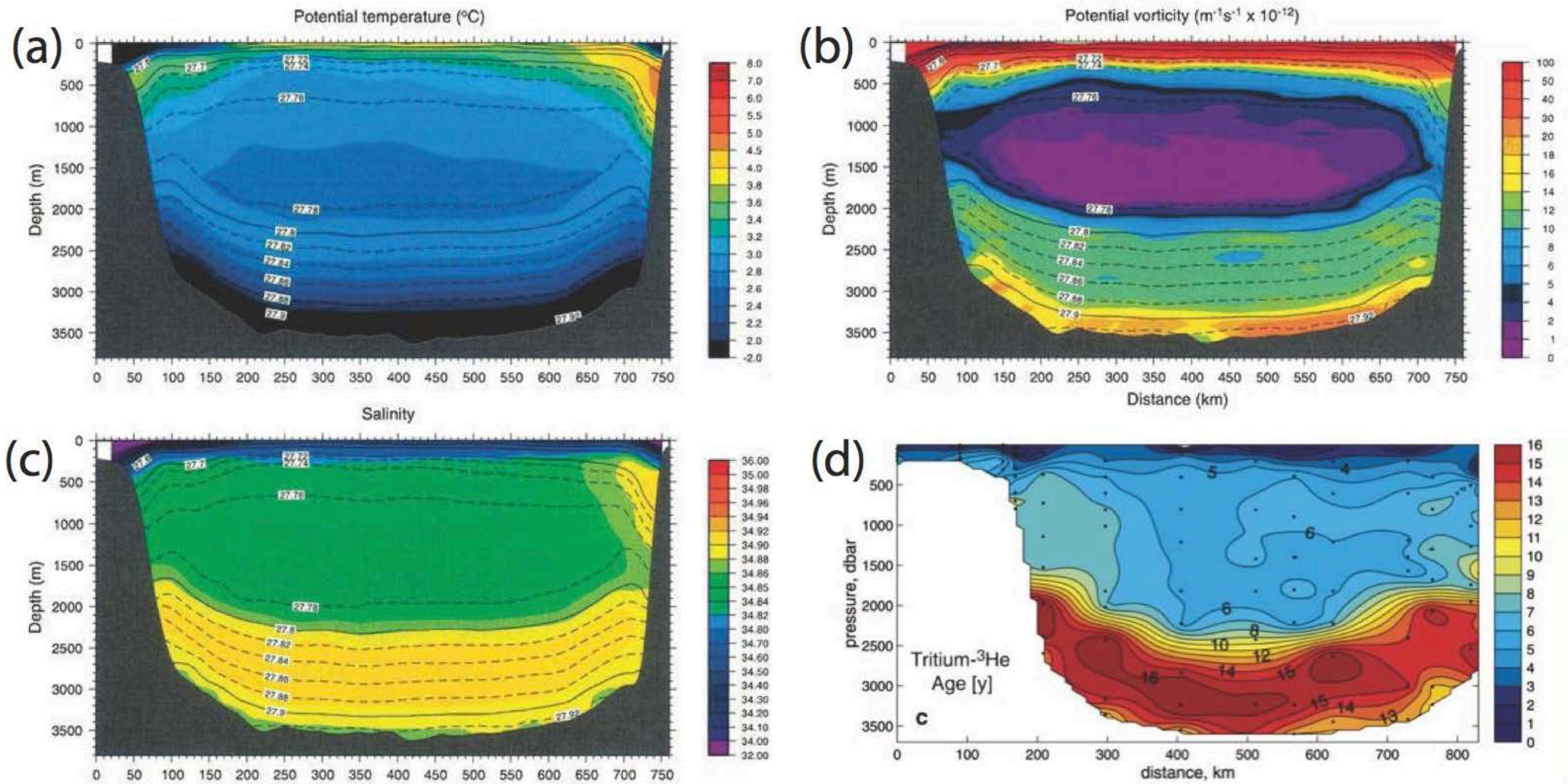
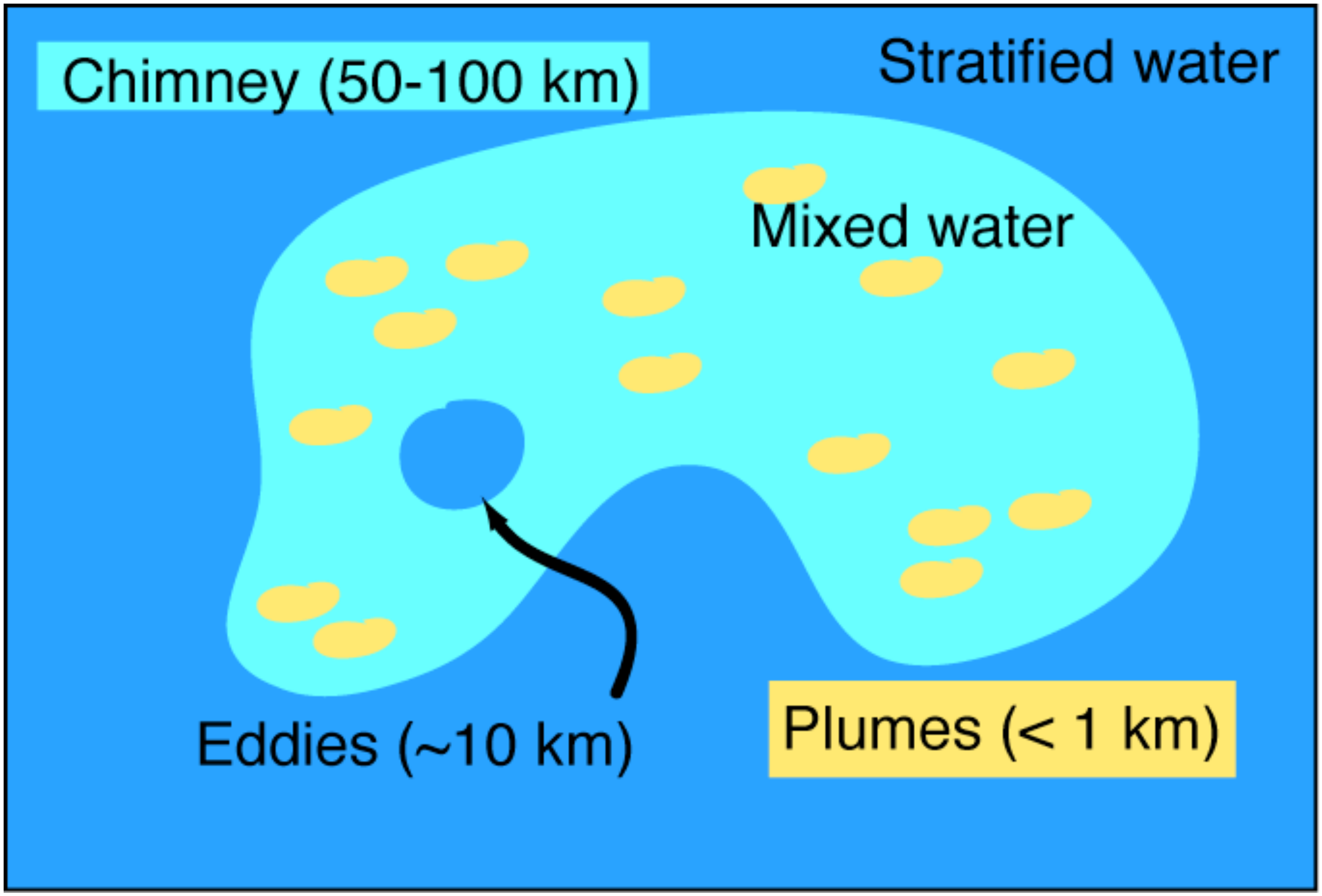


Figure 2 Hydrographic observations (a. potential temperature; b. potential vorticity; c. salinity) from the AR7W line averaged over the 1990s (*Pickart and Spall, 2007*). (d) Tritium-Helium age observed in 1993 (Khatiwala et al., 2002).

Open Ocean Deep convection

Lateral structure of convection regions



DPO S7.39

NADW sources: Nordic Seas Overflow Water

Deep convection in the Greenland Sea.

Overflow into the N. Atlantic over sills that are about 500-600 m deep.

Pathways: (1) Denmark Strait between Greenland and Iceland and
(2) Iceland-Faroe Ridge

Dickson and Brown (1995)

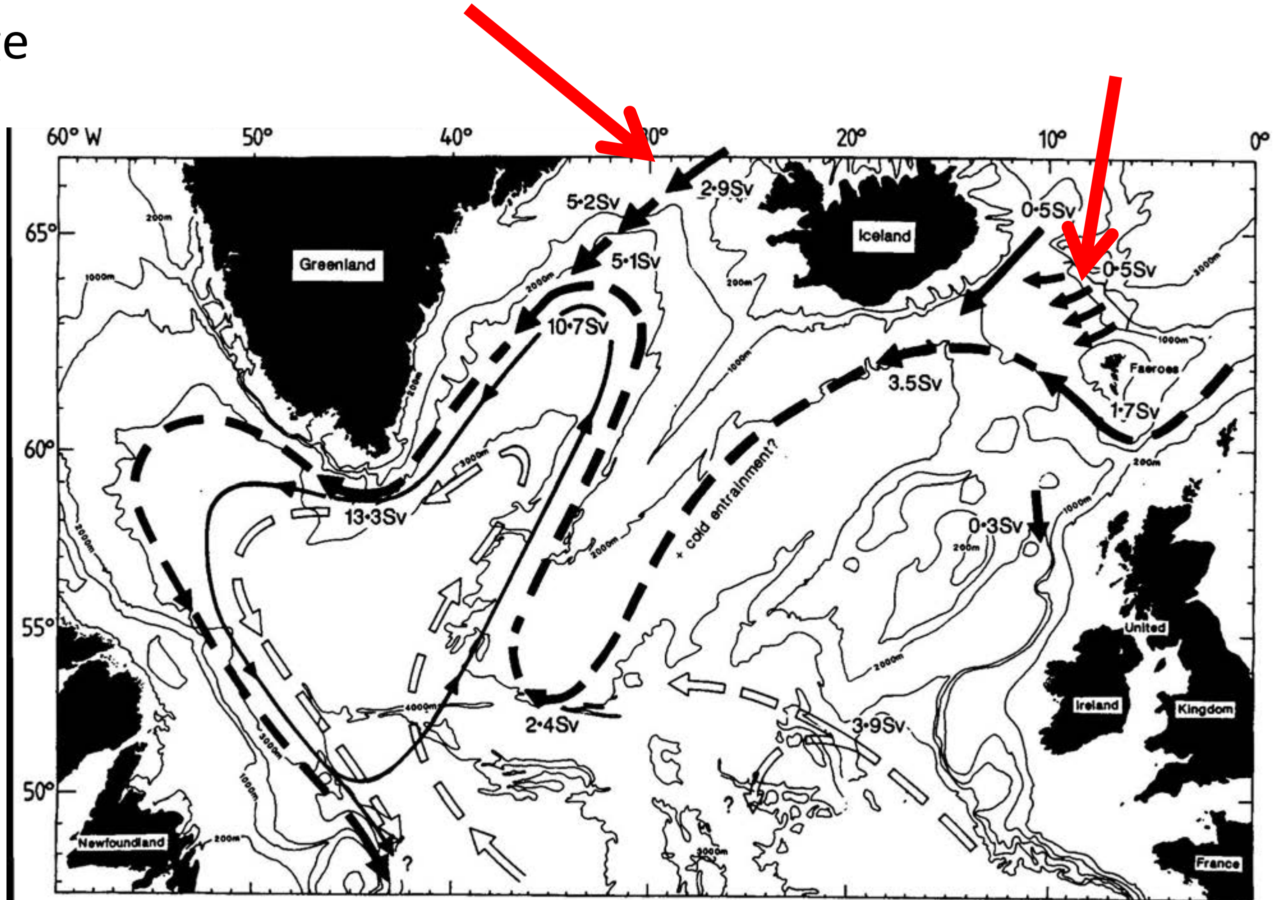
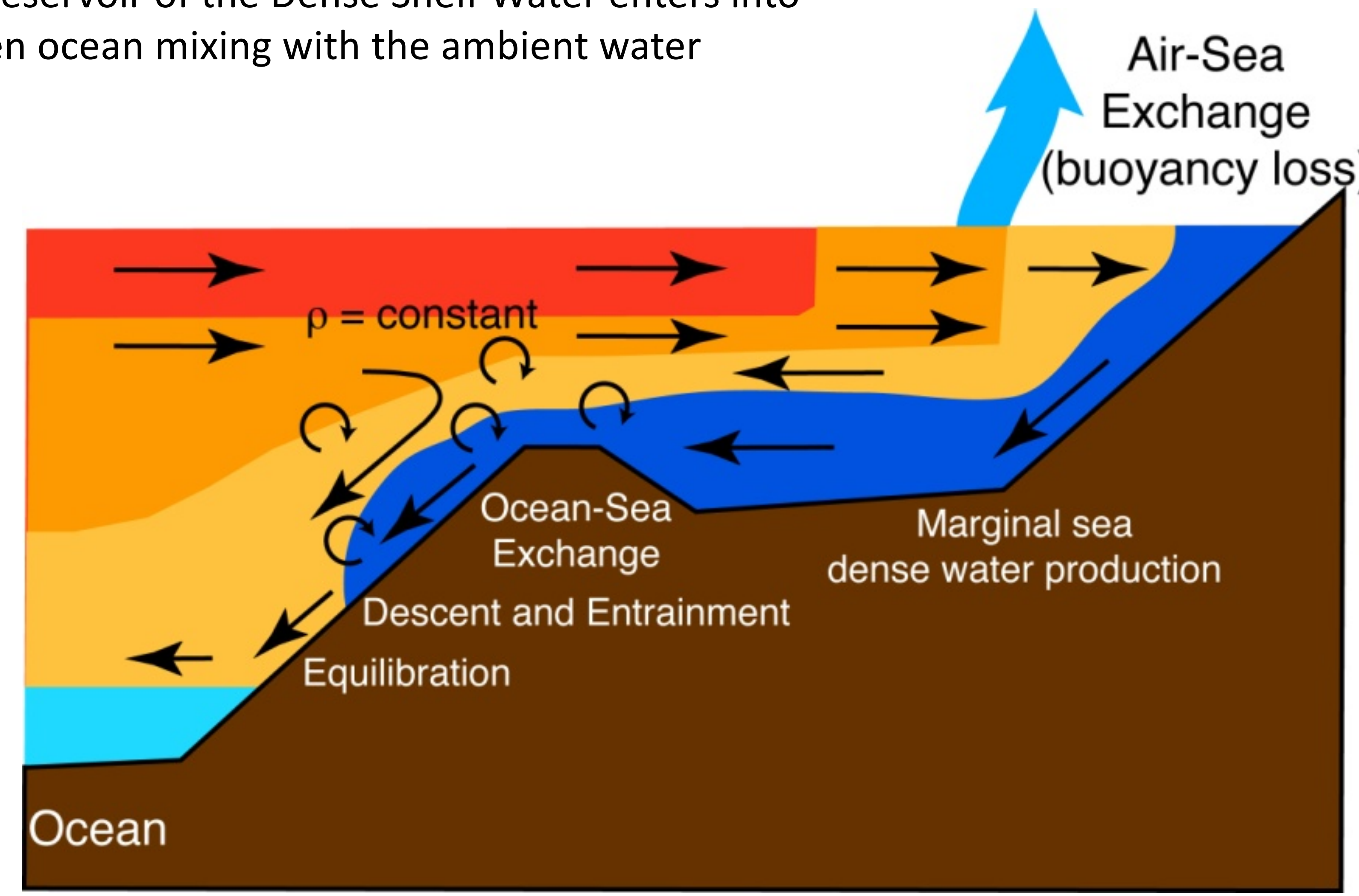


Figure 13. Proposed transport scheme for waters denser than $\sigma\theta = 27.80$ in the northern North Atlantic, based on all available measurements.

What happens in the "Overflow"?

1. Buoyancy loss by air-sea flux and/or brine rejection
2. Formation of Dense Shelf Water
3. The reservoir of the Dense Shelf Water enters into the open ocean mixing with the ambient water

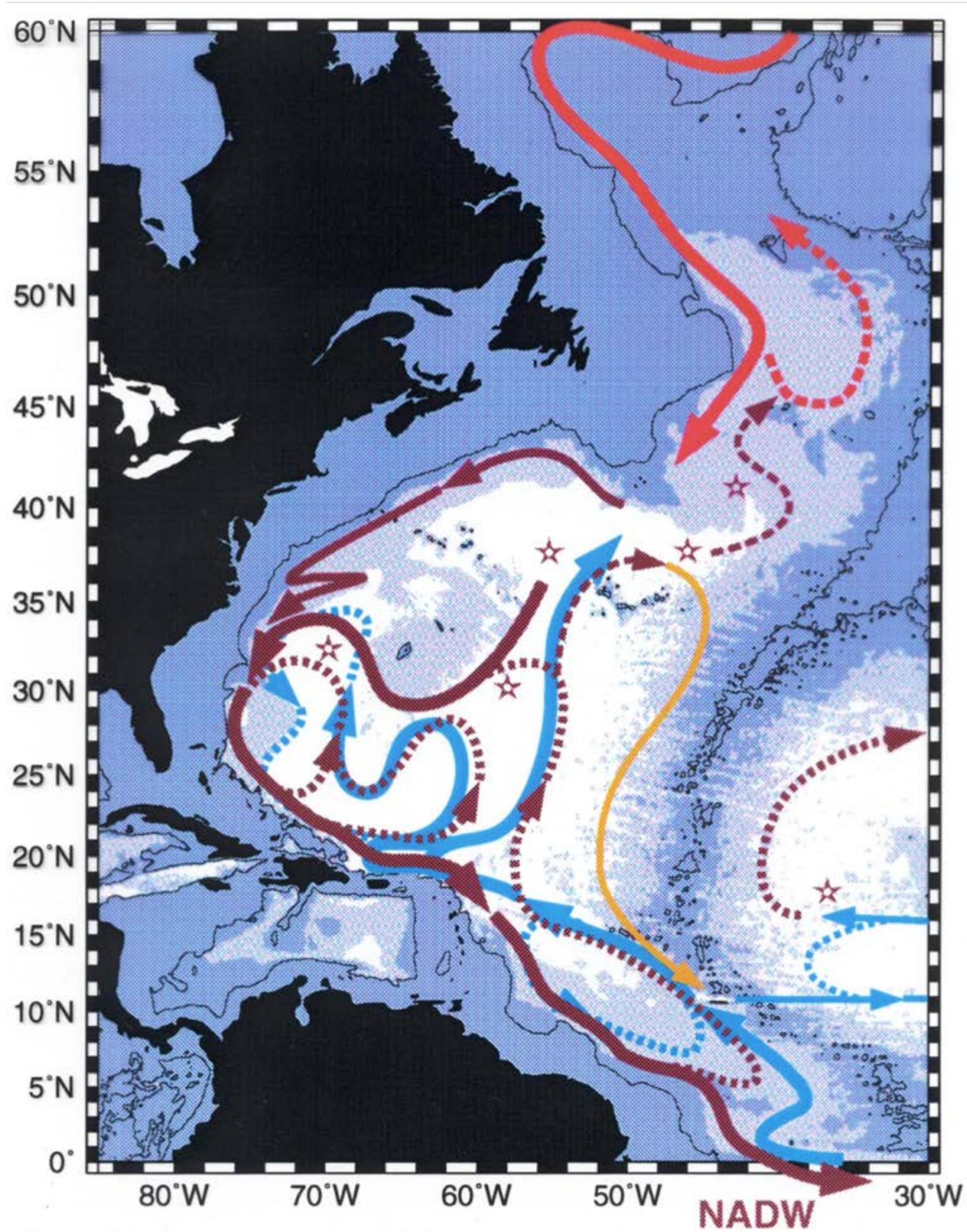


Deep Western Boundary Current

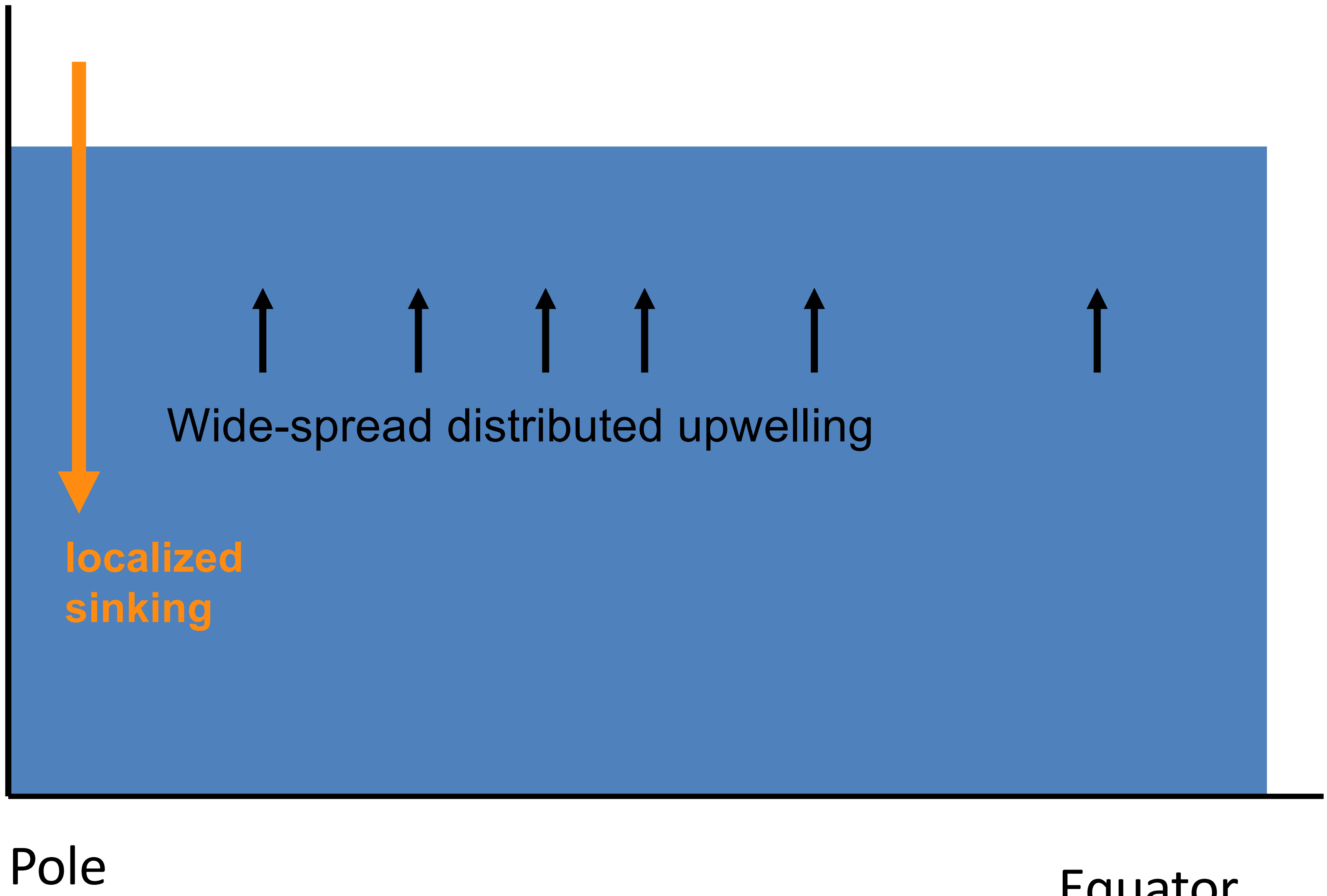
The export of NADW from the subpolar North Atlantic to the southern basins occurs through the deep western boundary current.

Stommel-Arons theory (1958) explains this phenomenon. It is the application of Sverdrup balance to the abyssal circulation.

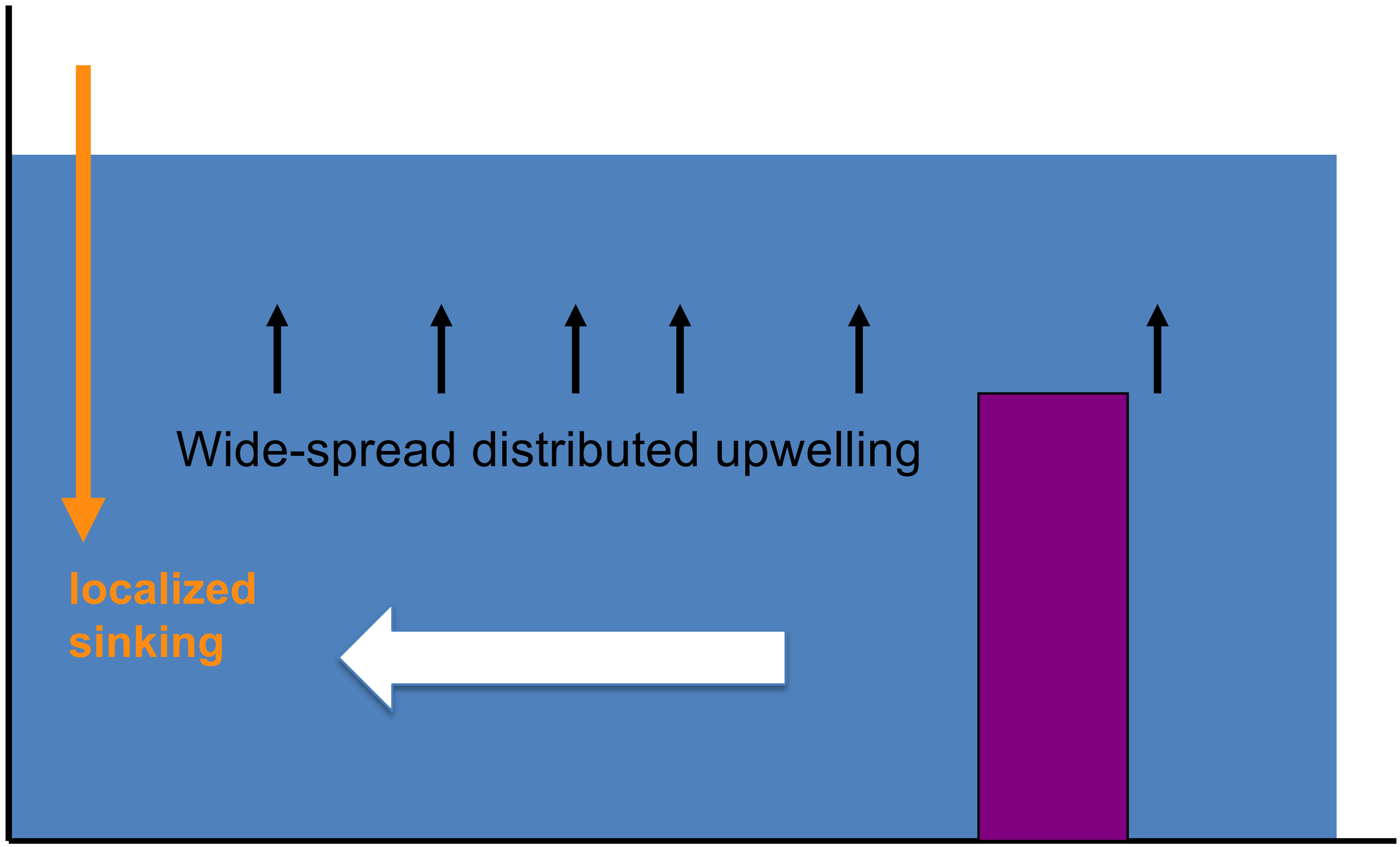
$$\beta v_g = f \frac{\partial w}{\partial z}$$



Abyssal circulation dynamics



Abyssal circulation dynamics

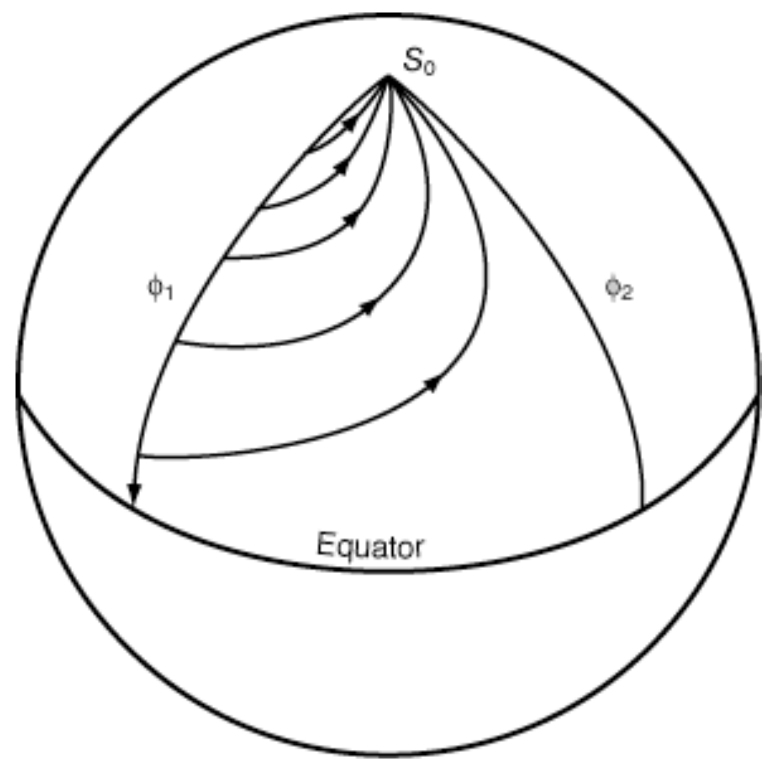


Water column is "stretching" due to upwelling. It tends to increase its absolute vorticity by moving poleward.

Deep western boundary current

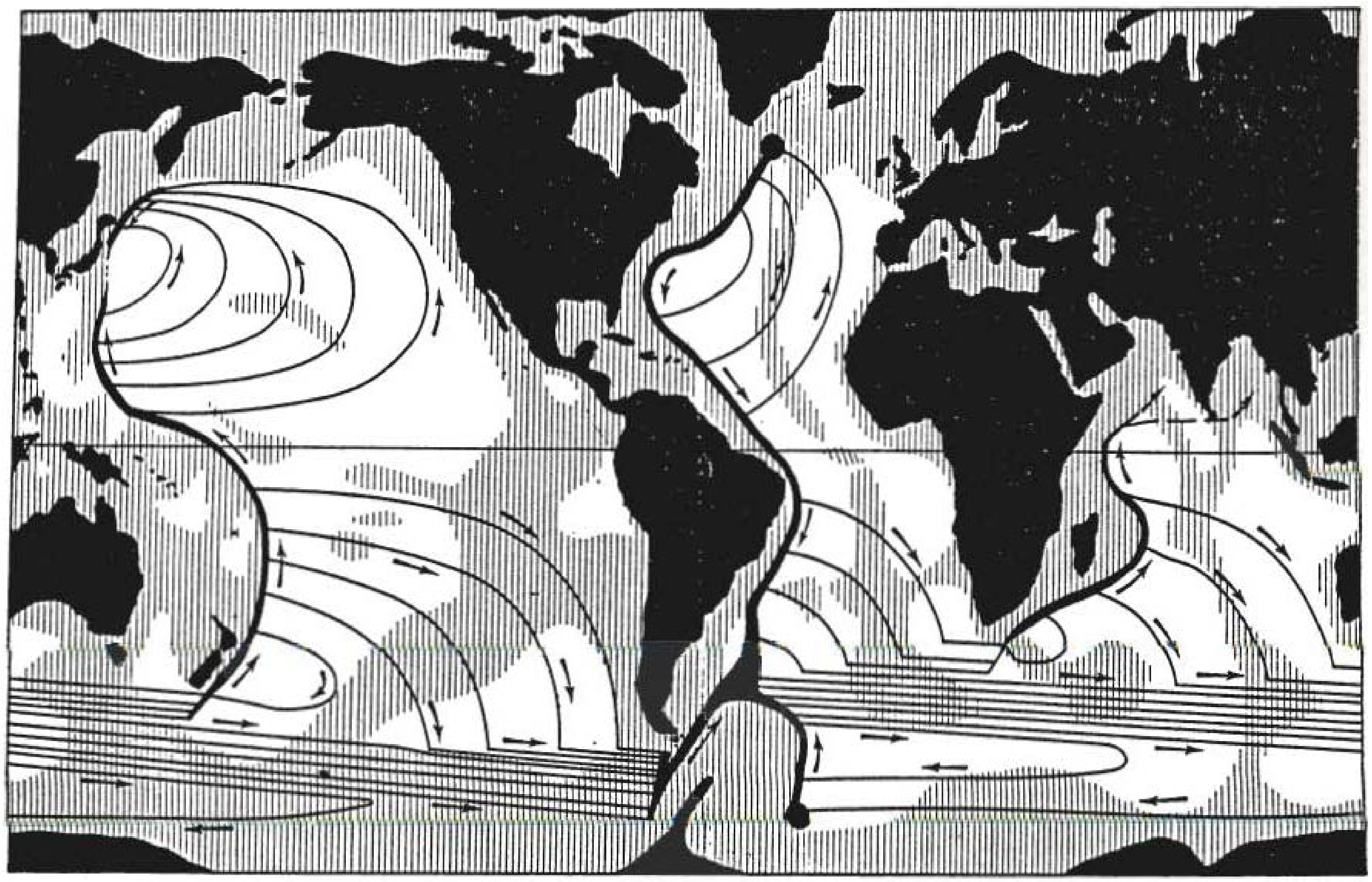
Uniformly distributed upwelling leads to the poleward Sverdrup circulation, which then returns equatorward along the western boundary of the basin. This is joined by the water sinking at the polar latitude.

Thus the intensity of the deep WBC can be much greater than the rate of deep water formation.



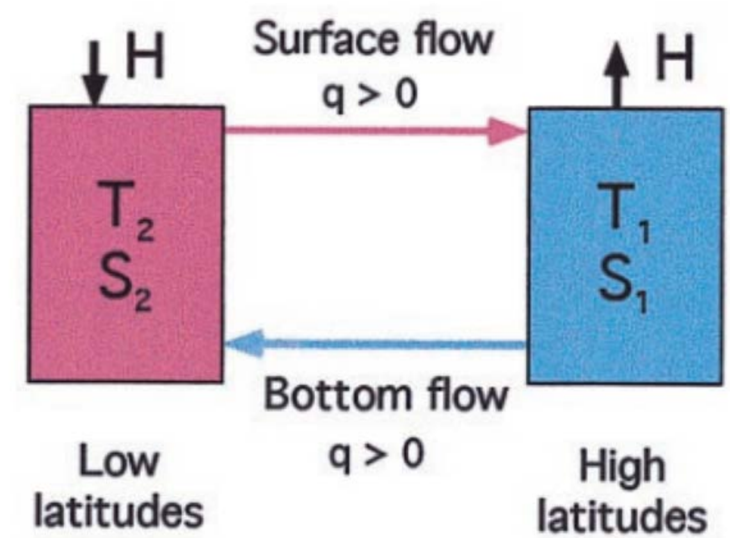
Stommel-Arons theory (1958)

Deep water formation is localized but the upwelling occurs more uniformly.
Upwelling drives poleward geostrophic flow in the deep ocean due to the beta-effect.



How stable is the thermohaline circulation?

- Stommel (1961)'s two-box model
  - Re-discovered in 1980s
- Two stable solutions: sinking in high versus low latitudes under the same atmospheric condition (thermal gradient and E-P)
- Different initial states lead to different solution for MOC



Redrawn by Marotzke (2000)

Increasing precipitation (E-P) at high latitudes lead to a shift between the two states

Multiple equilibria happens in 3D ocean model

NATURE VOL. 323 25 SEPTEMBER 1986 ARTICLES 301

High-latitude salinity effects and interhemispheric thermohaline circulations

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A general circulation model for the ocean is used to investigate the interaction between the global-scale thermohaline circulation and the salinity distribution. It is shown that an equatorially asymmetric circulation can be maintained even under equatorially symmetric basin geometry and surface forcing. Multiple equilibrium solutions are obtained for the same forcing by perturbing the high-latitude salinity field in an otherwise equatorially symmetric initial condition. The timescale of the transition from the symmetric circulation to an asymmetric circulation depends critically on the sign of the initial salinity perturbation.

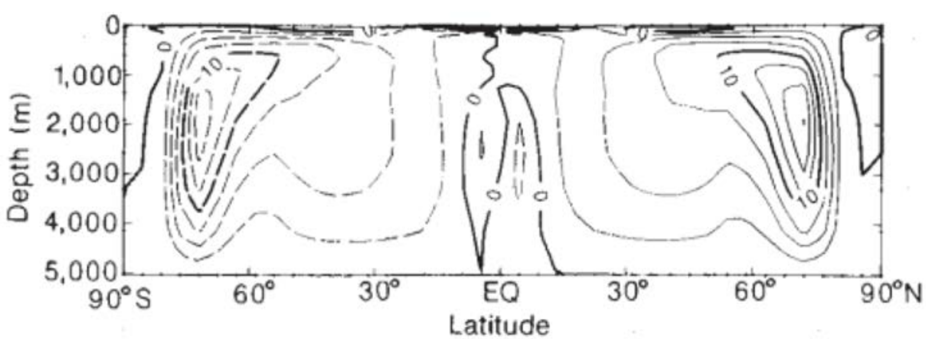

Fig. 2 Stream-function for the zonally integrated meridional overturning circulation for experiment 1 at the end of the integration. Positive values indicate a clockwise circulation; contour interval, $2.5 \times 10^6\ m^3\ s^{-1}$.

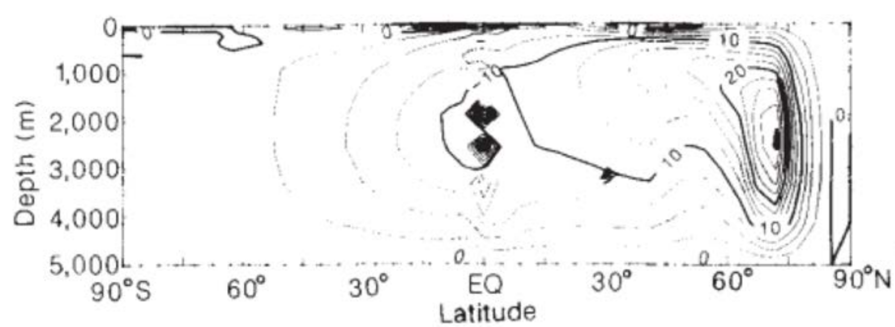

Fig. 5 Stream-function for the zonally integrated meridional overturning circulation for experiment 4 at the end of the integration. Positive values indicate a clockwise circulation; contour interval, $2.5 \times 10^6\ m^3\ s^{-1}$.

Insights from paleoceanographic data

NATURE VOL. 330 5 NOVEMBER 1987 ARTICLES 35

North Atlantic thermohaline circulation during the past 20,000 years linked to high-latitude surface temperature

Edward A. Boyle* & Lloyd Keigwin†

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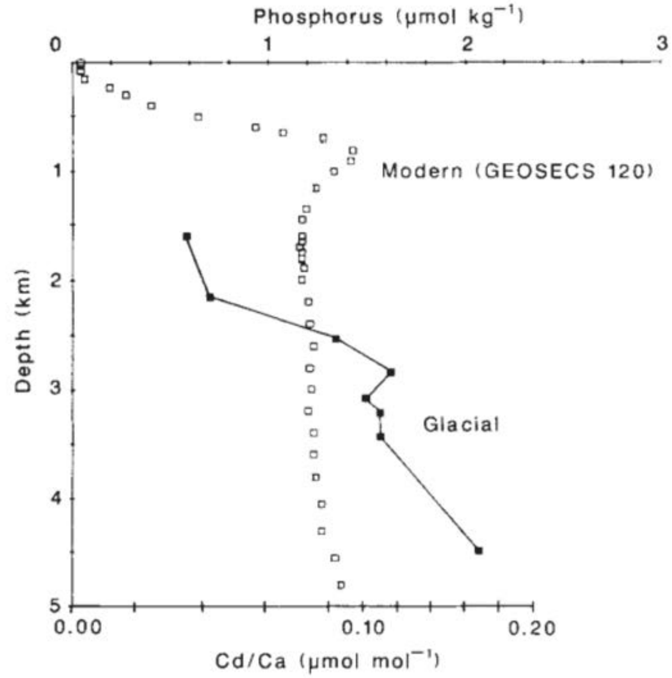
† Woods Hole Oceanographic Institution, Woods Hole, Massachusetts 02543, USA

During a surface cooling event 10,000 to 12,000 years ago, higher Cd/Ca and lower $^{13}C/^{12}C$ ratios are observed in benthic foraminifera shells from rapidly accumulating western North Atlantic sediments. Data from sediment cores show that marked nutrient depletion of intermediate waters occurs in association with reduced glacial North Atlantic Deep Water flux. It is proposed that cold high-latitude sea surface temperatures enhance intermediate-water formation at the expense of deep-water formation.

Boyle and Keigwin (1987)



REVIEW & COMMENT

THE GREAT OCEAN CONVEYOR

By Wallace S. Broecker

(Broecker , 1991)

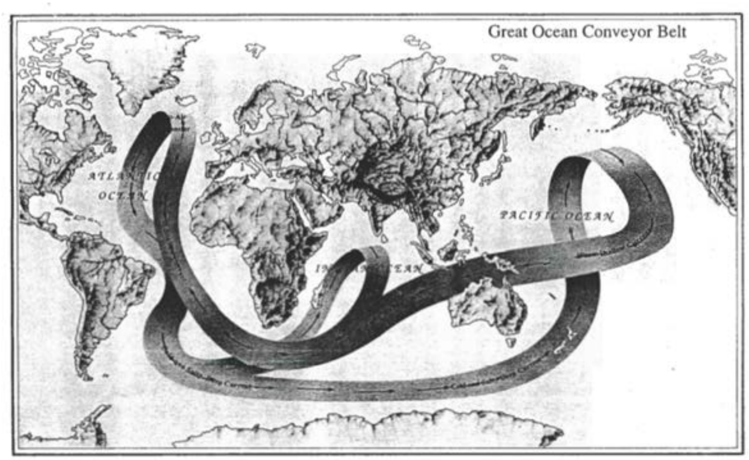


Fig. 1: The great ocean conveyor logo (Broecker, 1987). (Illustration by Joe Le Monnier, Natural History Magazine.)

Can global warming shut down AMOC?

Climate sensitivity due to increased CO_2: experiments with a coupled atmosphere and ocean general circulation model*

Warren M Washington and Gerald A Meehl

National Center for Atmospheric Research** Boulder, CO 80307-3000, USA

Washington and Meehl (1989, Clim. Dyn.)

Interhemispheric asymmetry in climate response to a gradual increase of atmospheric CO_2

R. J. Stouffer, S. Manabe & K. Bryan

Geophysical Fluid Dynamics Laboratory/NOAA, Princeton University, Princeton, New Jersey 08542, USA

Stouffer et al. (1989, Nature)

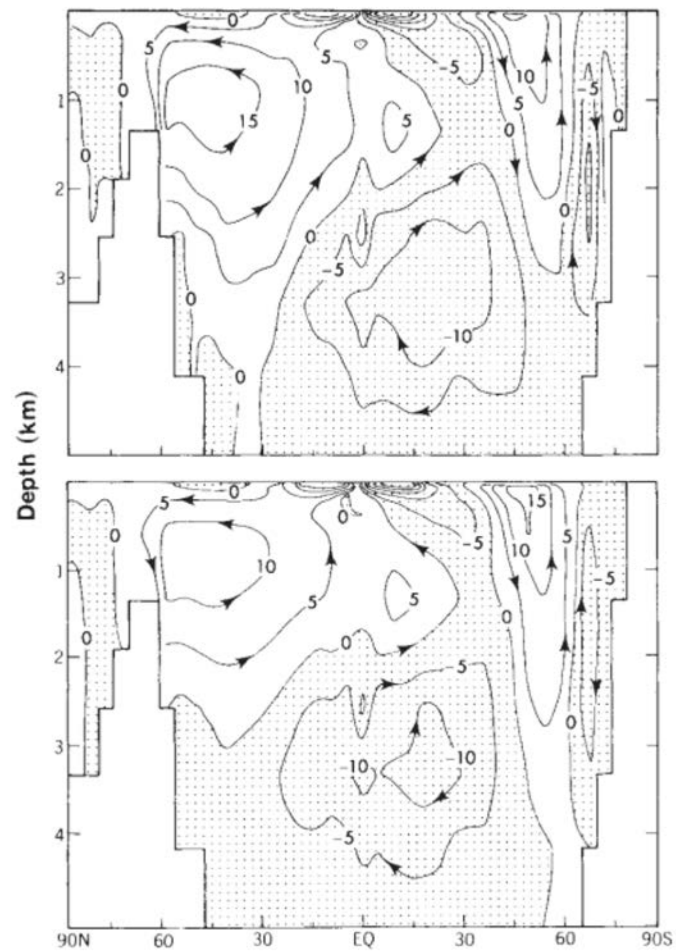


FIG. 4 Streamlines of zonal-mean meridional oceanic circulation (in Sverdrups) averaged over the seventh decade (that is, years 61–70) of the experiment. Top, the control integration with constant CO_2; bottom, the integration in which atmospheric CO_2 concentration is increased with the rate of 1% yr^{-1}.

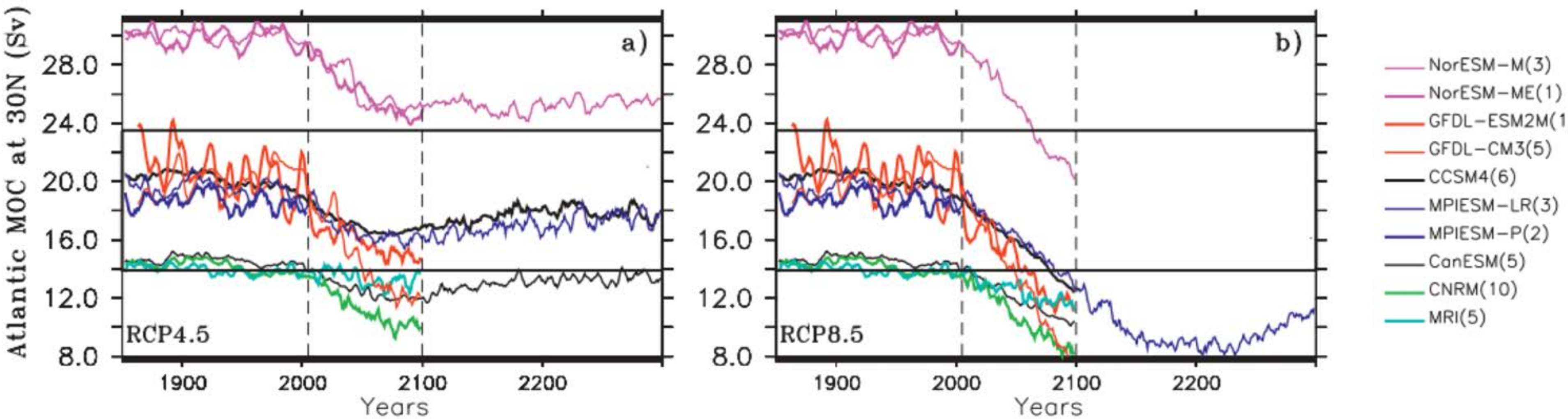
Current climate model projections

Atlantic Meridional Overturning Circulation (AMOC) in CMIP5 Models: RCP and Historical Simulations

WEI CHENG

Joint Institute for the Study of the Atmosphere and Ocean, University of Washington, and Pacific Marine Environmental Laboratory, Seattle, Washington

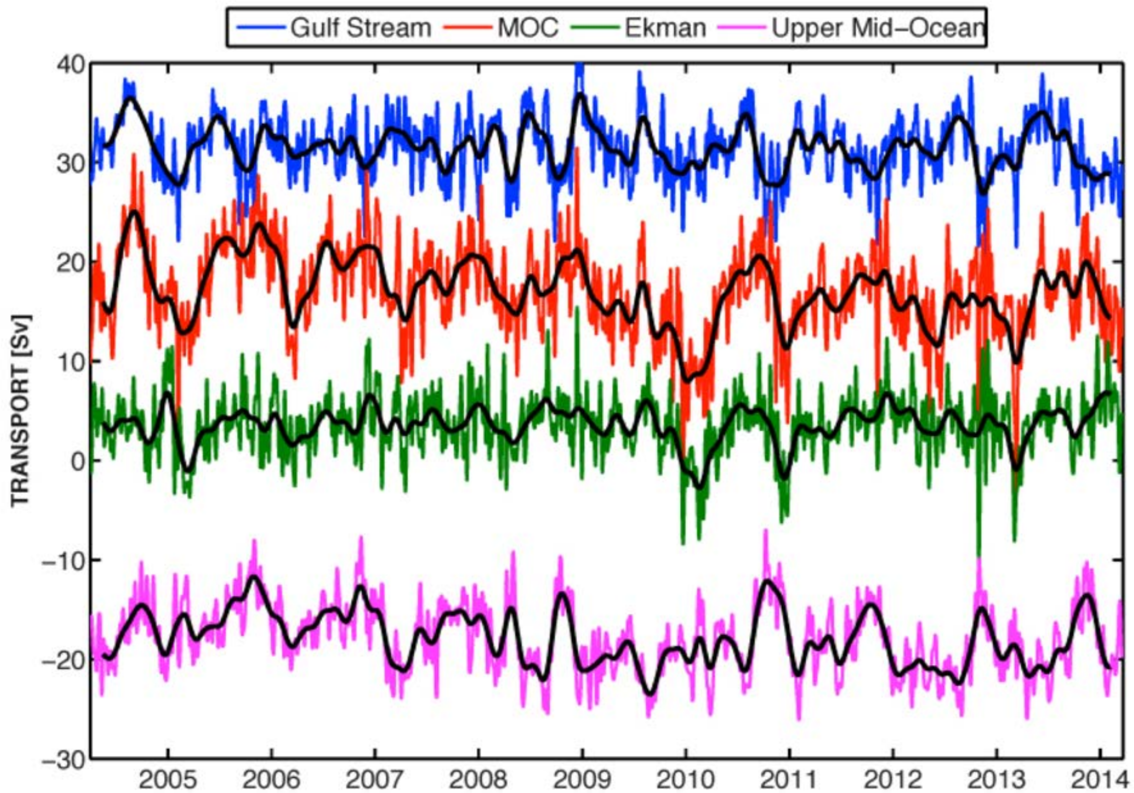
Cheng et al., (2013) J. Climate



RCP4.5

RCP8.5

Observation of AMOC



Blue = Florida strait (telephone cable data)

Green = Ekman (wind data)

Pink = Upper Mid-Ocean (mooring data)

26N Rapid observation (2004-)

UK-RAPID