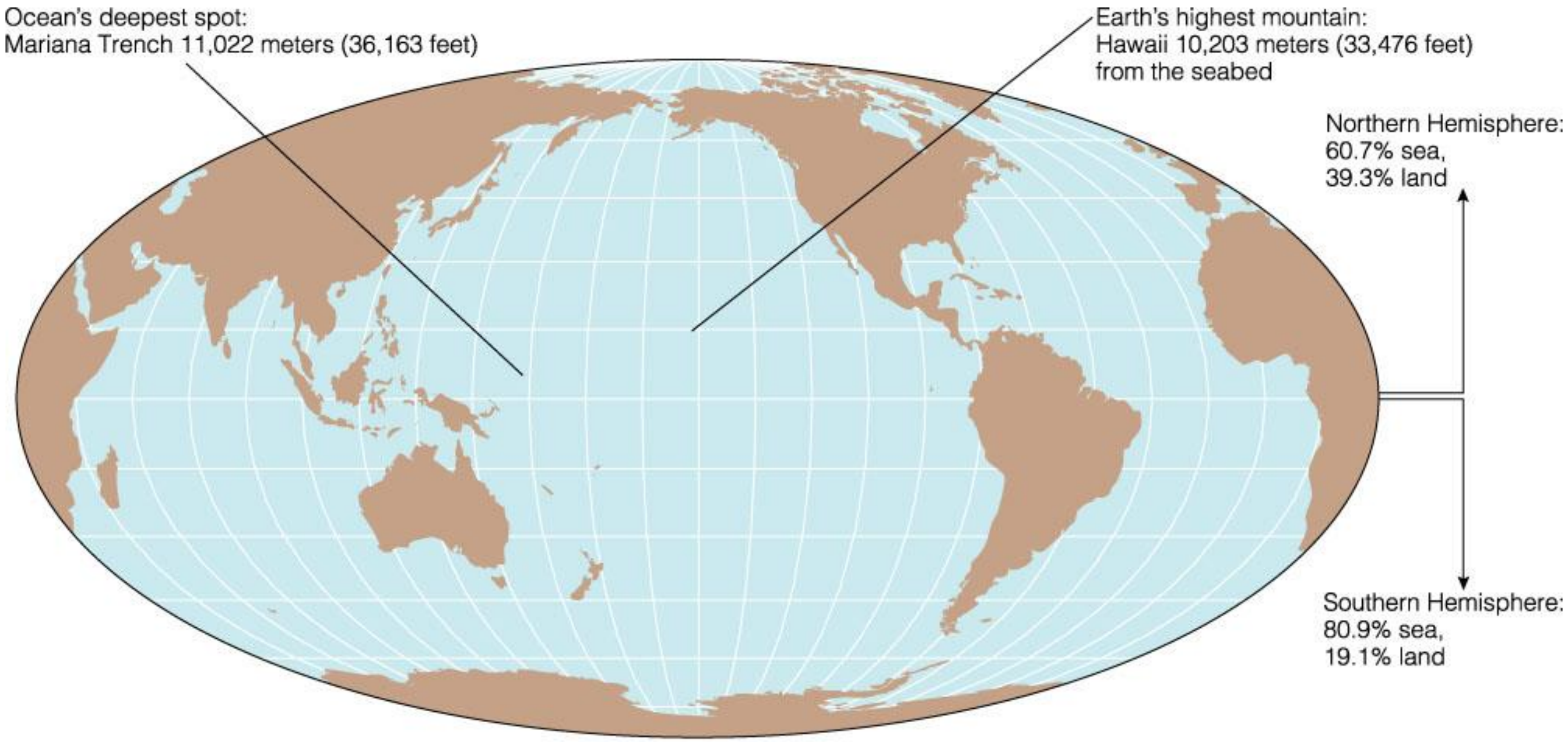
Physical Oceanography

1. Ocean, Origin, Geology
2. Geography, Topography
3. Temperature, Salinity, Density and Oxygen – Horizontal/Vertical
4. Mixed Layer
5. Water Masses
6. General Circulation of the Oceans
   - Thermohaline Circulation
   - Wind driven circulation
7. Upper ocean processes, heat fluxes

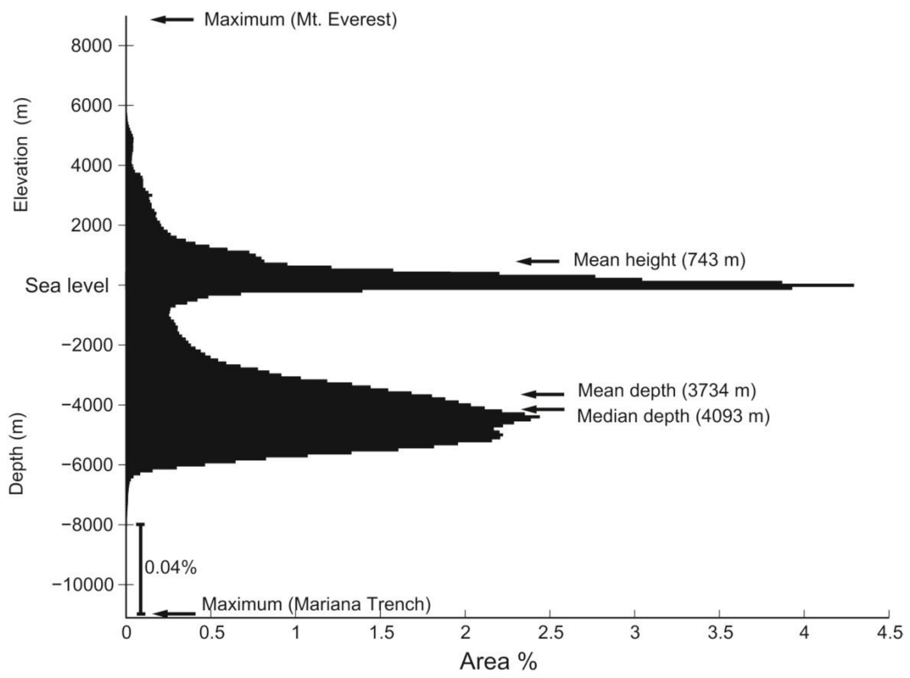
MSc Atmospheric Sciences 2016
IITM-University of Pune

Roxy Mathew Koll :: CCCR/IITM

The Ocean



Area: 361,100,000 square kilometers (139,400,000 square miles)
Volume: 1,370,000,000 cubic kilometers (329,000,000 cubic miles)
Average depth: 3,796 meters (12,451 feet)
Average temperature: 3.9°C (39.0°F)
Average salinity: 34,482 grams per kilogram (0.56 ounce per pound), 3.4%
Average land elevation: 840 meters (2,772 feet)
Age: About 4 billion years
Future: Uncertain

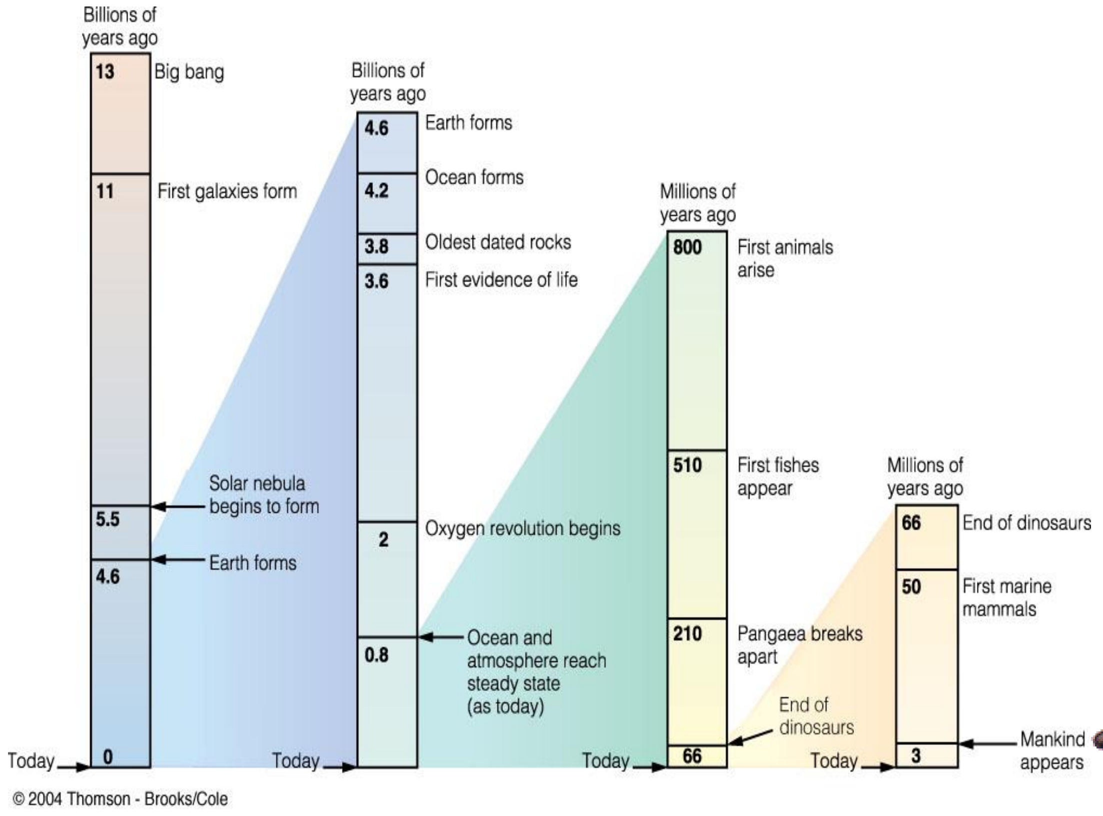


Areas of Earth's surface above and below sea level as a percentage of the total area of Earth (in 100m intervals). Data from Becker et al. (2009).

1. The ocean covers 71% of Earth's surface
2. The ocean contains 97% of the water on Earth
3. The ocean is Earth's most important feature
4. They are only temporary features of a single world ocean
5. Average ocean depth is about 4½ times greater than the average height of the continents above sea level

Origin of the Ocean

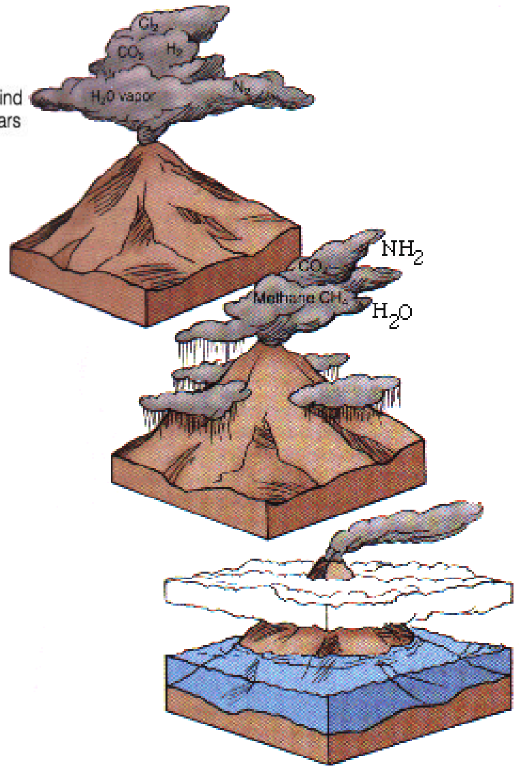
1. **Source**: The major trapped volatile during the formation of Earth was water (H2O).
   Others included nitrogen (N2), the most abundant gas in the atmosphere, carbon dioxide (CO2), and hydrochloric acid (HCl), which was the source of the chloride in sea salt (mostly NaCl).



2. **Outgassing**: Ocean had its origin from the prolonged escape of water vapor and other gases from the molten igneous rocks of the Earth to the clouds surrounding the cooling Earth.

3. **Cooling and Rains**: After the Earth's surface had cooled to a temperature below the boiling point of water, rain began to fall and continued to fall for centuries.

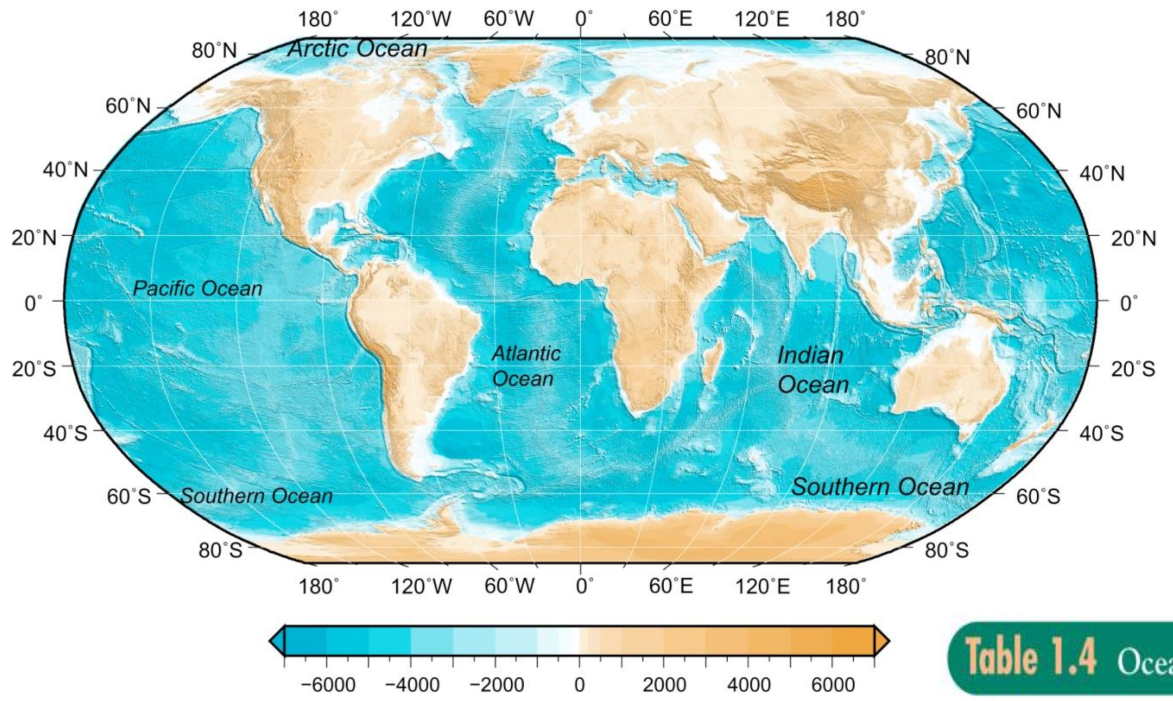
4. **Present state**: As the water drained into the great hollows in the Earth's surface, the primeval ocean came into existence, about 4billion years ago. The forces of gravity prevented the water from leaving the planet.



Origin of the Ocean : Etymology

Greek French English
ōkeanós -> occean -> ocean

Oceans, Distribution and Topography



1. Pacific
2. Atlantic
3. Indian
4. Arctic

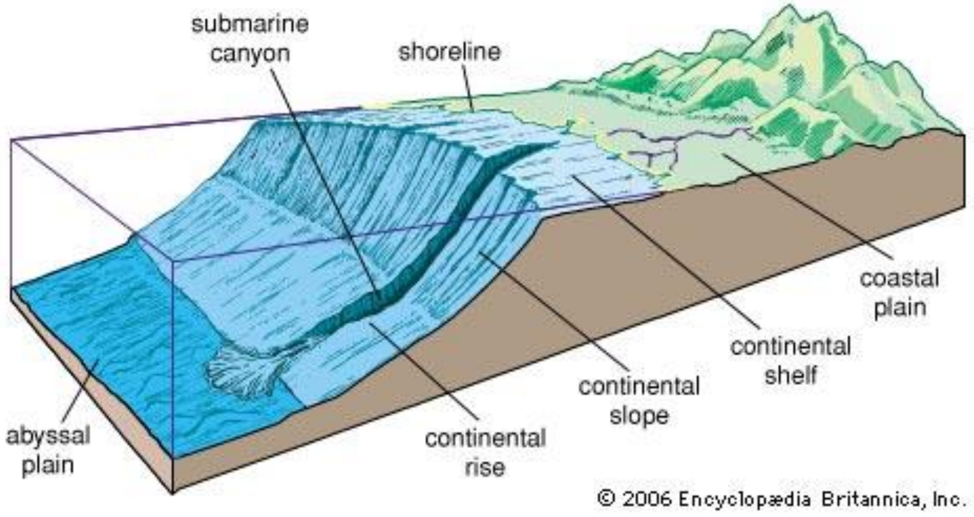
Southern Ocean

The 4 oceans are interconnected through smaller, marginal seas, such as:
the Mediterranean, Gulf of Mexico, South China Sea, Red Sea, Caribbean, Baltic Sea, Bering Sea, etc.

Table 1.4 Ocean Depths, Areas, and Volumes

Ocean	Average Depth	Area	Volume	Percent of Ocean Surface Area	Percent of Earth's Surface Area
Pacific	4028 m (13,216 ft)	179.7×10^6 km^2 (69.3×10^6 mi^2)	723.8×10^6 km^3 (173.3×10^6 mi^3)	50.1	35.5
Atlantic	3332 m (10,932 ft)	93.7×10^6 km^2 (36.1×10^6 mi^2)	312.2×10^6 km^3 (74.8×10^6 mi^3)	25.9	18.4
Indian	3897 m (12,786 ft)	73.6×10^6 km^2 (28.4×10^6 mi^2)	286.8×10^6 km^3 (68.7×10^6 mi^3)	20.4	14.4
Arctic	1117 m (3665 ft)	14.1×10^6 km^2 (5.4×10^6 mi^2)	15.7×10^6 km^3 (3.8×10^6 mi^3)	3.9	2.8
All Oceans	3795 m (12,451 ft)	361.1×10^6 km^2 (139.3×10^6 mi^2)	1338.5×10^6 km^3 (320.6×10^6 mi^3)	100	70.8

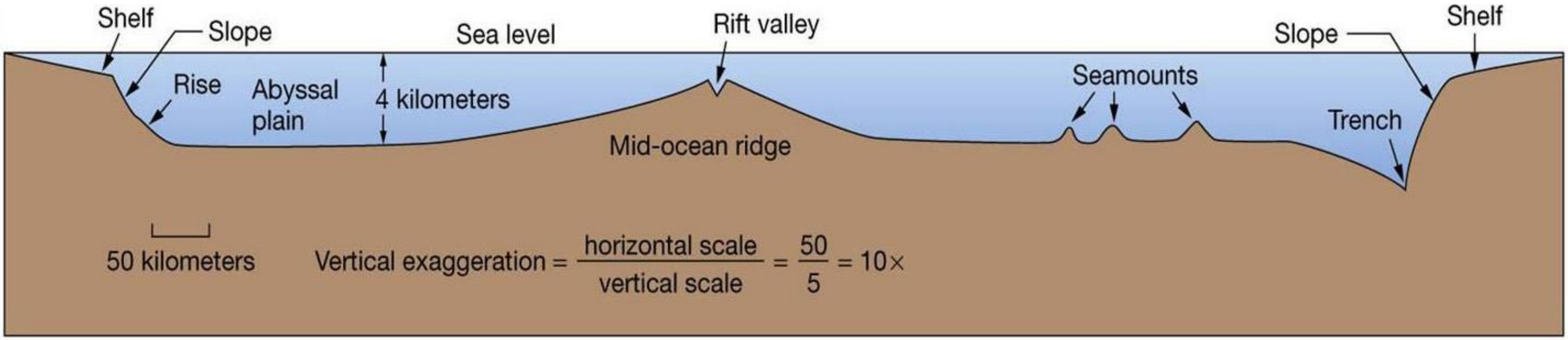
Oceans, Topography

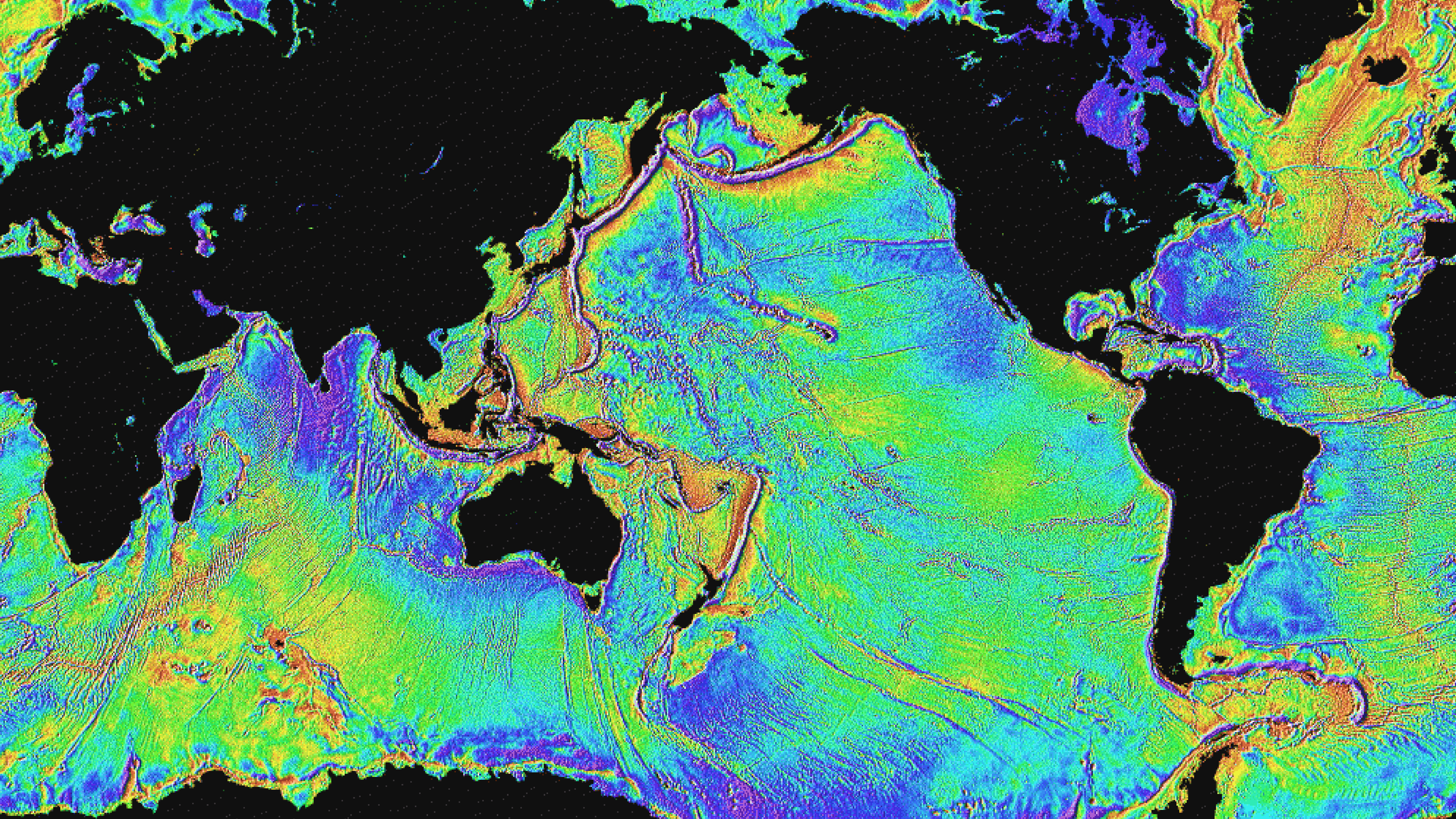


The **continental shelf** extends seaward from the shore with an average gradient of 1 in 500. Its outer limit (the shelf break) is set where the gradient increases to about 1 in 20 (on average) to form the continental slope down to the deep sea bottom. The shelf has an average width of 65 km.

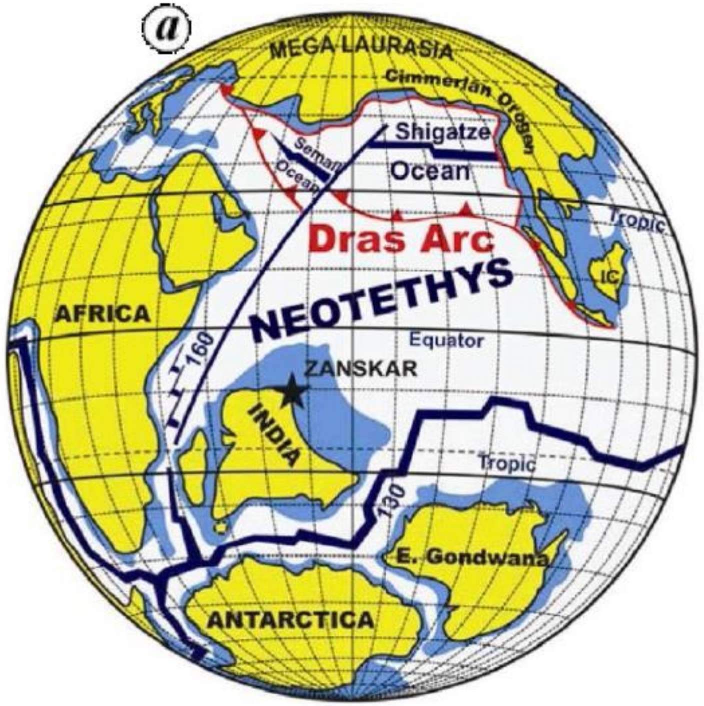
The average depth at the shelf break is about 130 m. Most of the world's fisheries are located on the continental shelves for a multitude of reasons including proximity of estuaries, depth of penetration of sunlight compared with bottom depth, and upwelling of nutrient-rich waters onto some shelves, particularly those off western coasts.

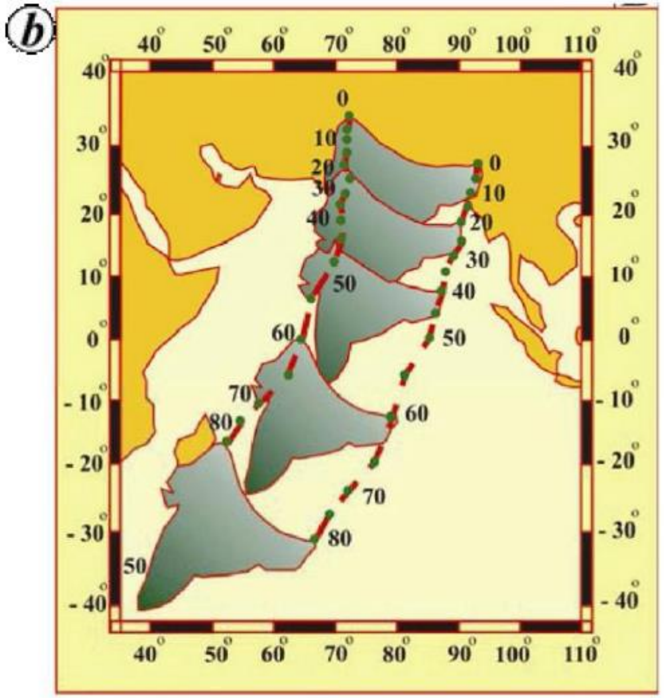
The **continental slope** averages about 4000m vertically from the shelf to the sea bottom. The lower part of the slope, where it grades into the deep-sea bottom, is referred to as the **continental rise**.

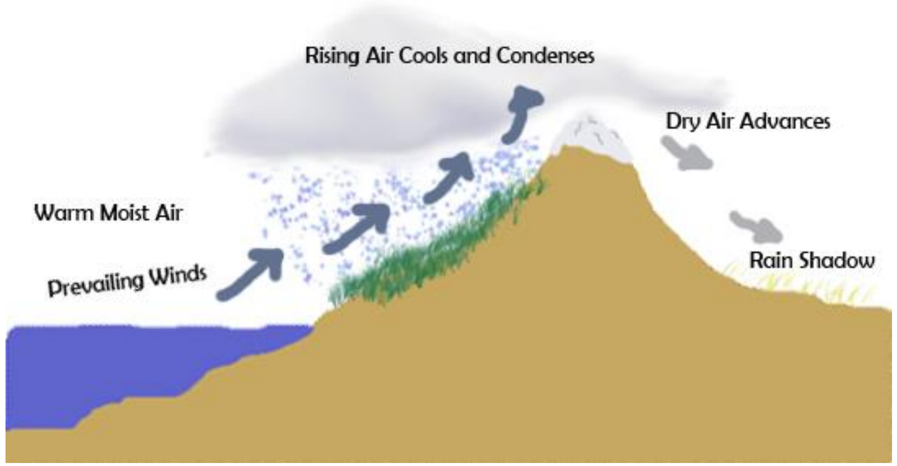




Indian Ocean, Monsoon







Indian plate motion: rate of ~ 15cm/yr
– collided with Asia c.20-30 Mya.

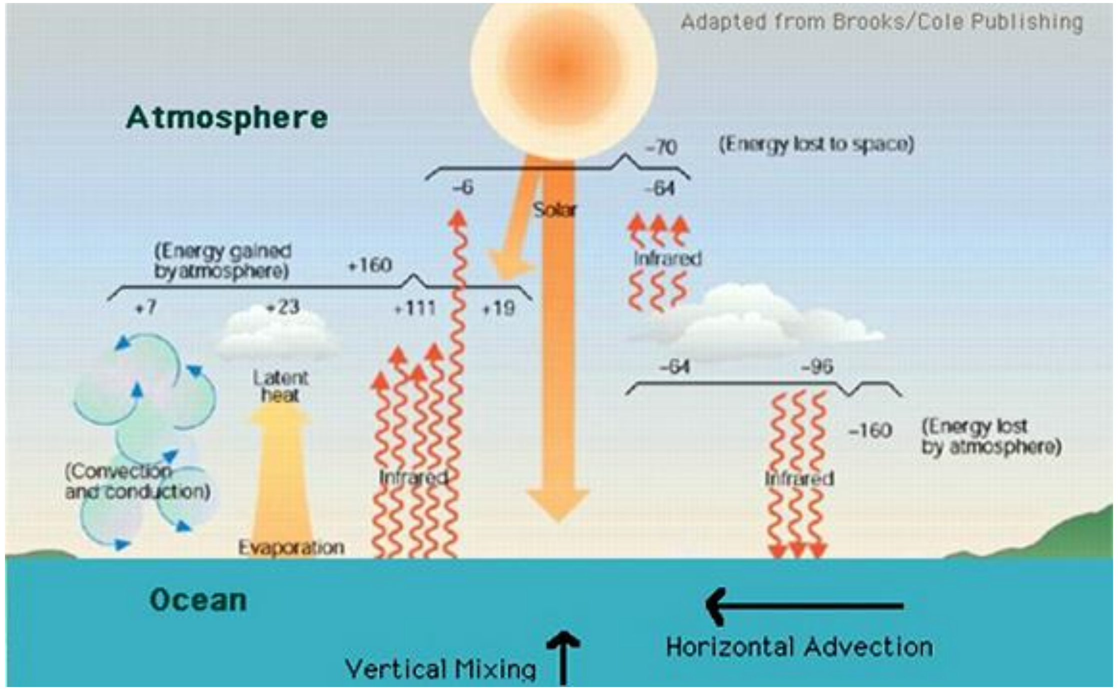
Rajasaurus Narmadensis
-T.rex contemporary

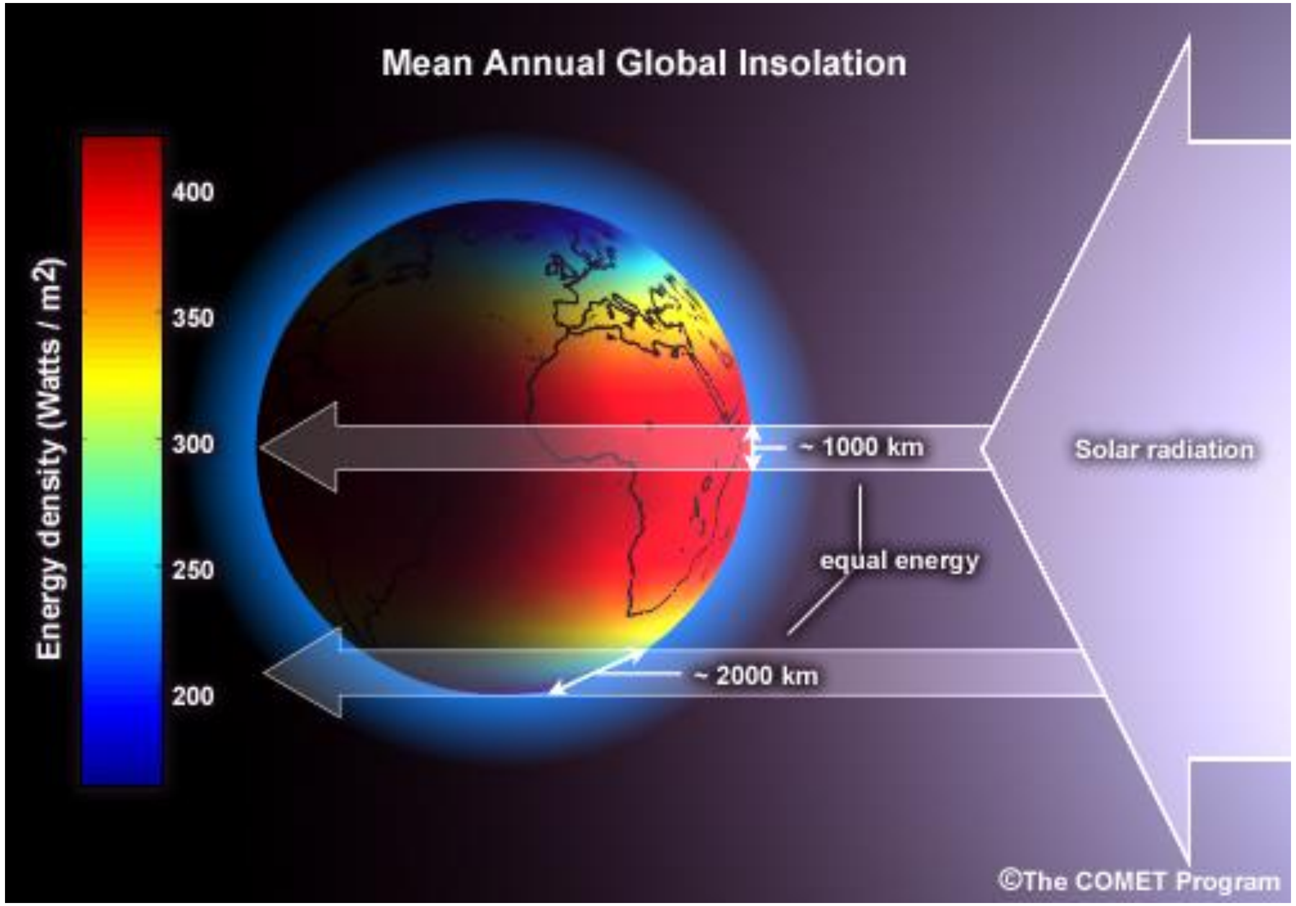
Copley et al. *JGR,* 2010; Jain *Curr.Sci.*, 2014

Some remnants of Deccan volcanism

Layers of basaltic lava deposits – Mahabaleshwar, Maharashtra

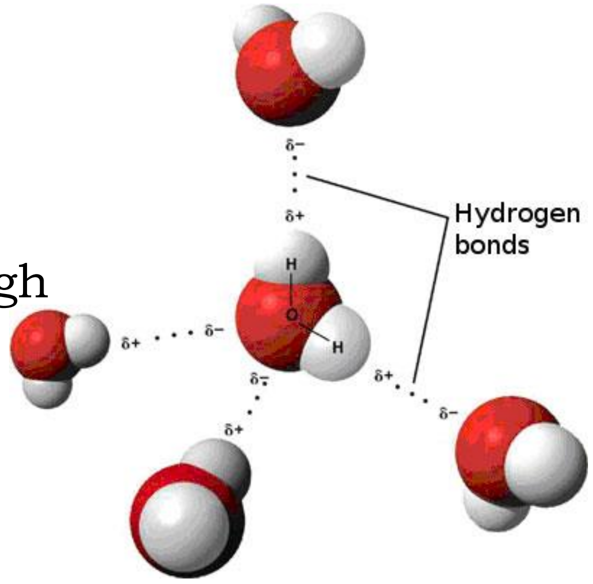
Properties of Seawater Temperature: Solar Radiation





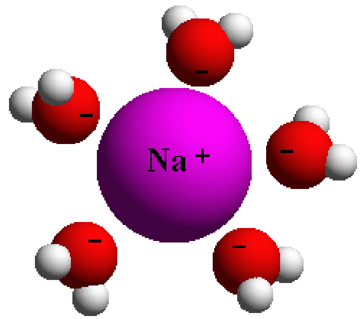
Motion of the atmosphere is driven by the uneven distribution of solar energy. Solar heating is greatest in equatorial regions and causes water in the oceans to evaporate and the moist air to rise. The warm, humid air forms a belt of equatorial clouds and heavy rainfall. It is bordered in the middle latitudes by high-pressure zones that are cloud-free and contain dry descending air.

Properties of Seawater

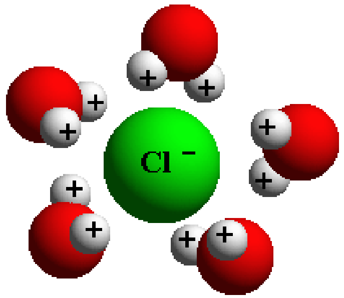


Polarity of water: Consisting of two positively charged hydrogen ions and a single negatively charged oxygen ion, water is arranged as a polar molecule having positive and negative sides. This molecular polarity leads to water's high dielectric constant (ability to withstand or balance an electric field).

Water is able to dissolve many substances because the polar water molecules align to shield each ion, resisting the recombination of the ions. The ocean's salty character is due to the abundance of dissolved ions.

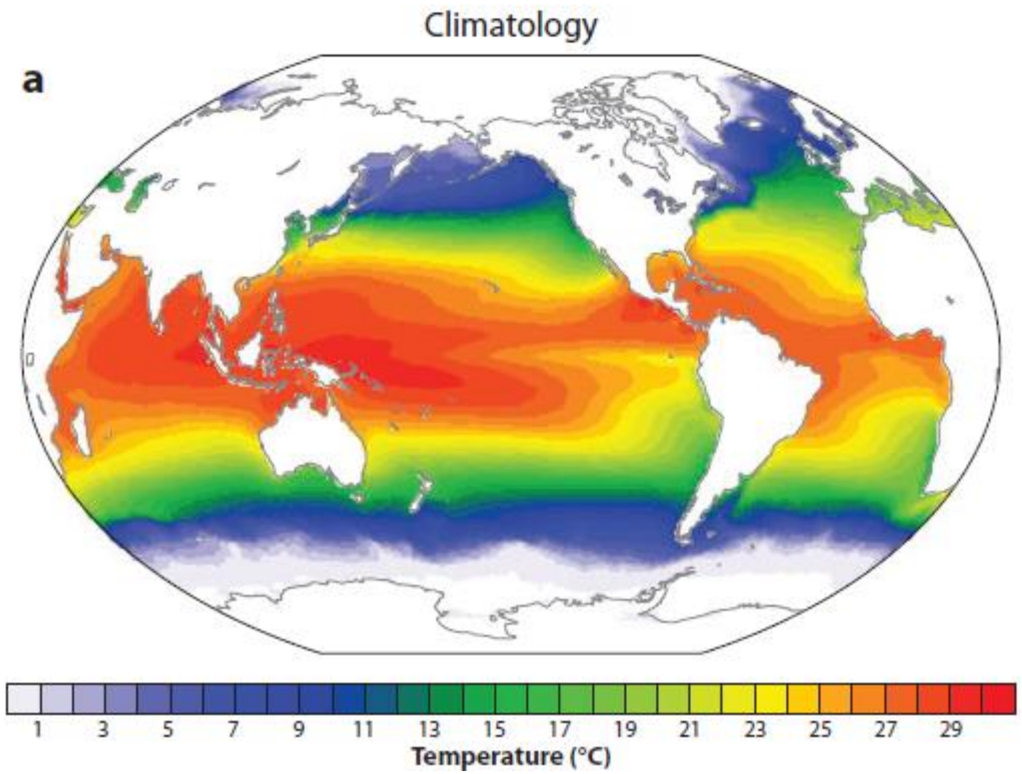


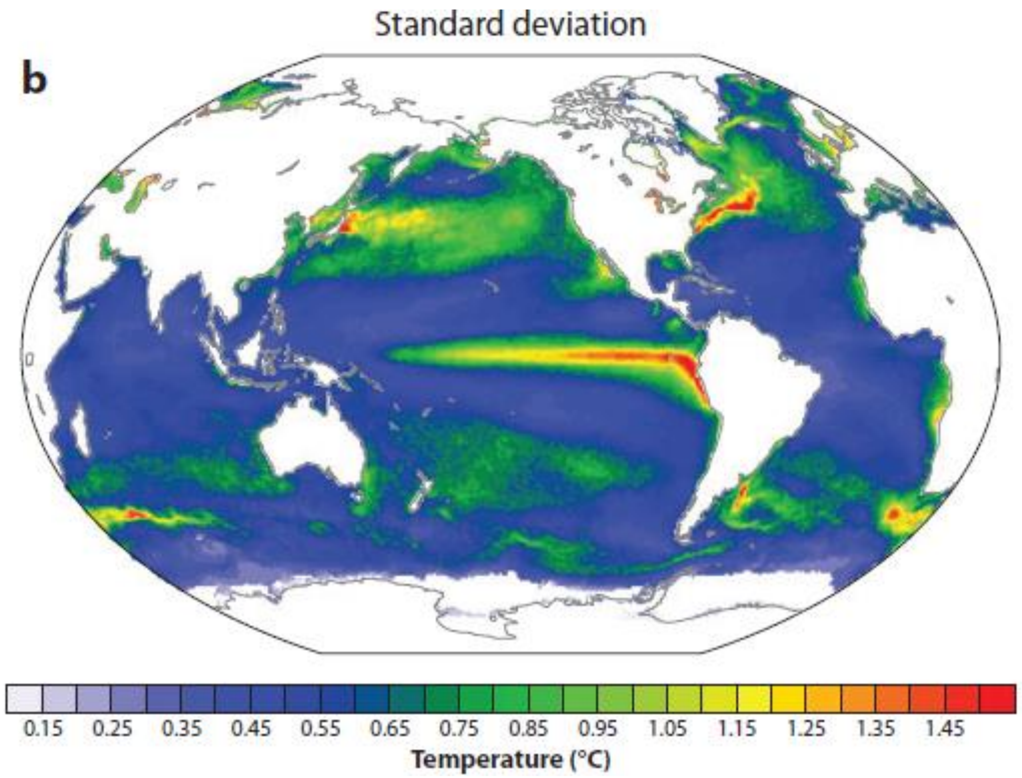
Heat capacity of water: The polar nature of the water molecule causes it to form polymer-like chains of up to eight molecules. Approximately 90% of the water molecules are found in these chains. Energy is required to produce these chains, which is related to water's heat capacity. Water has the highest heat capacity of all liquids except ammonia.



Ocean as a heat reservoir: This high heat capacity is the primary reason the ocean is so important in the world climate system. Unlike the land and atmosphere, the ocean stores large amounts of heat energy it receives from the sun. This heat is carried by ocean currents, exporting or importing heat to various regions. Approximately 90% of the anthropogenic heating associated with global climate change is stored in the oceans, because water is such an effective heat reservoir.

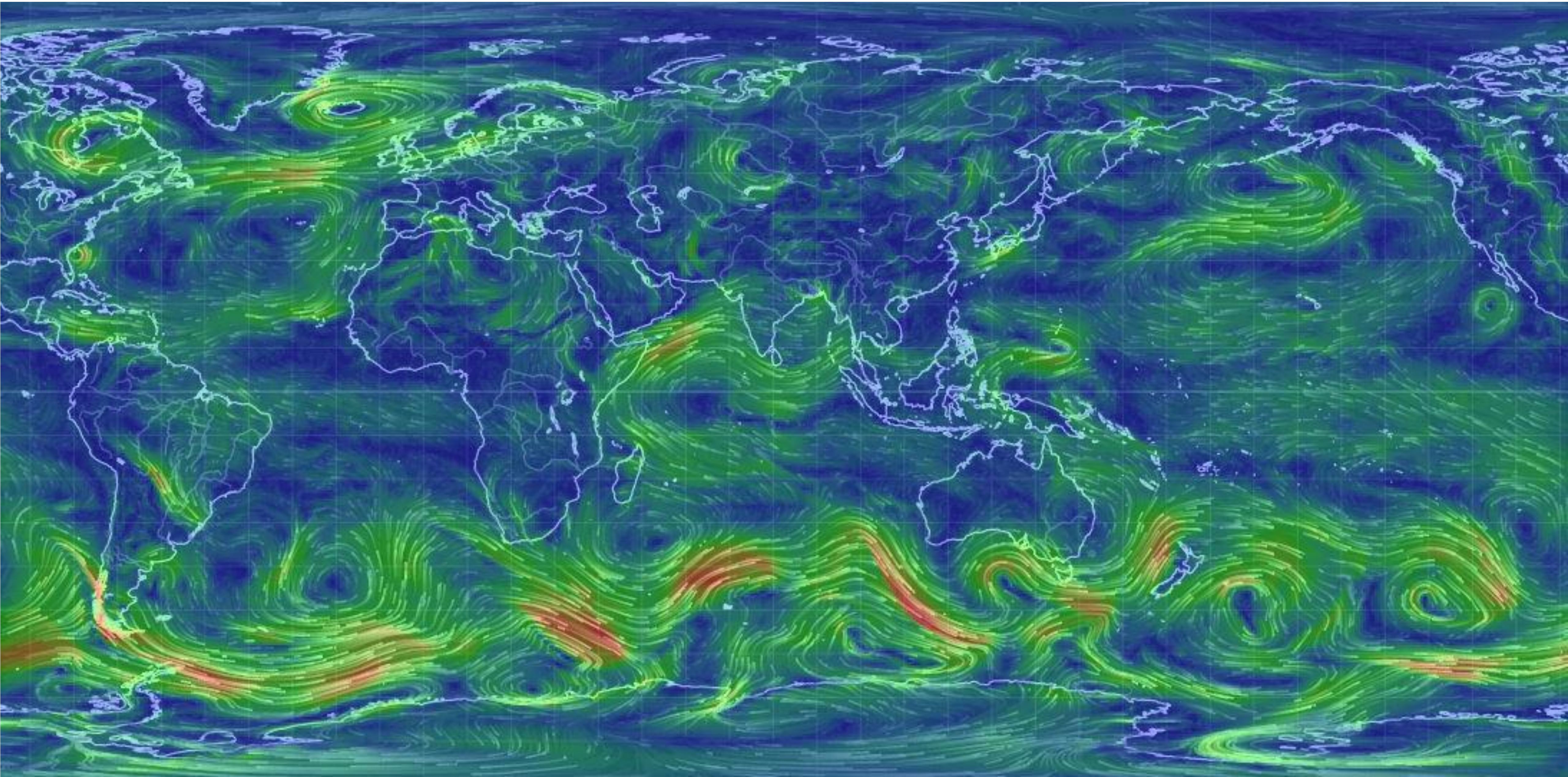
Properties of Seawater
Sea Surface Temperature [SST]





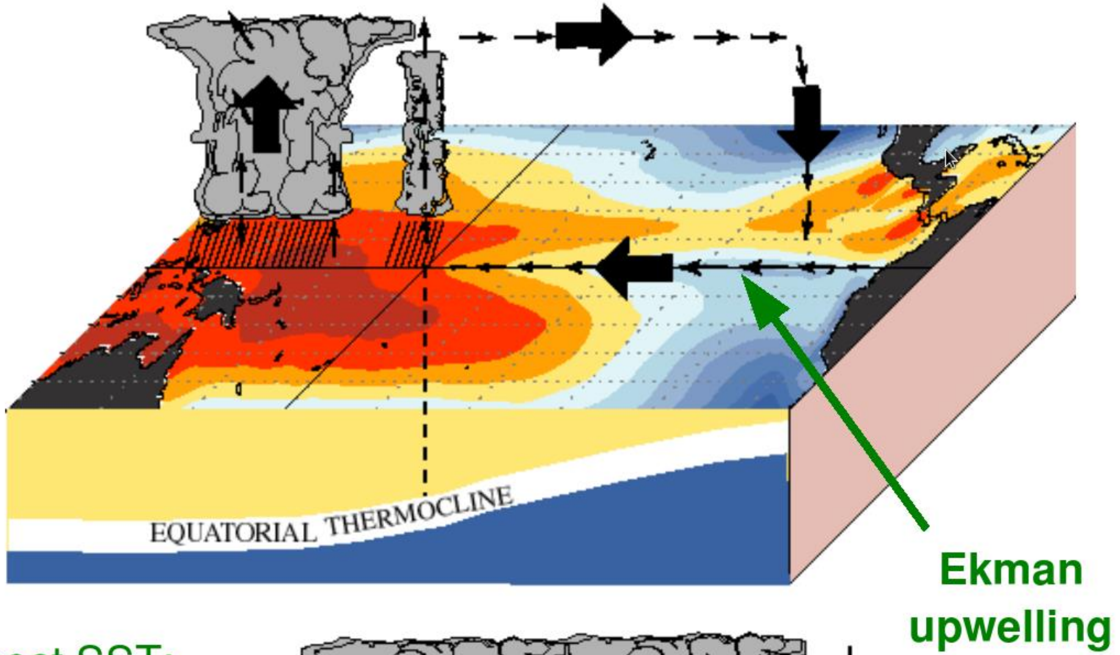
Deser et al 2010. note where the 4°C isotherm occurs (most ocean volume is colder than this)

Trade Winds, ENSO and Monsoon



Trade Winds, ENSO and Monsoon

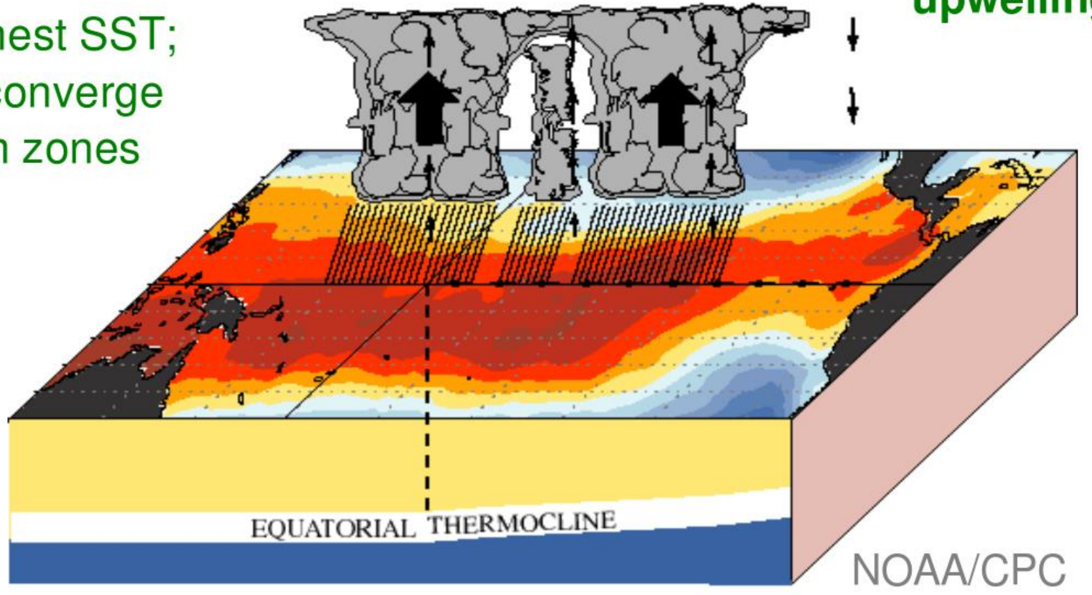
Normal



Asymmetric:
cold tongue, warm pool;
ITCZ north of equator;
easterly trade winds;
sea level slopes **up** to west,
thermocline **down** to west

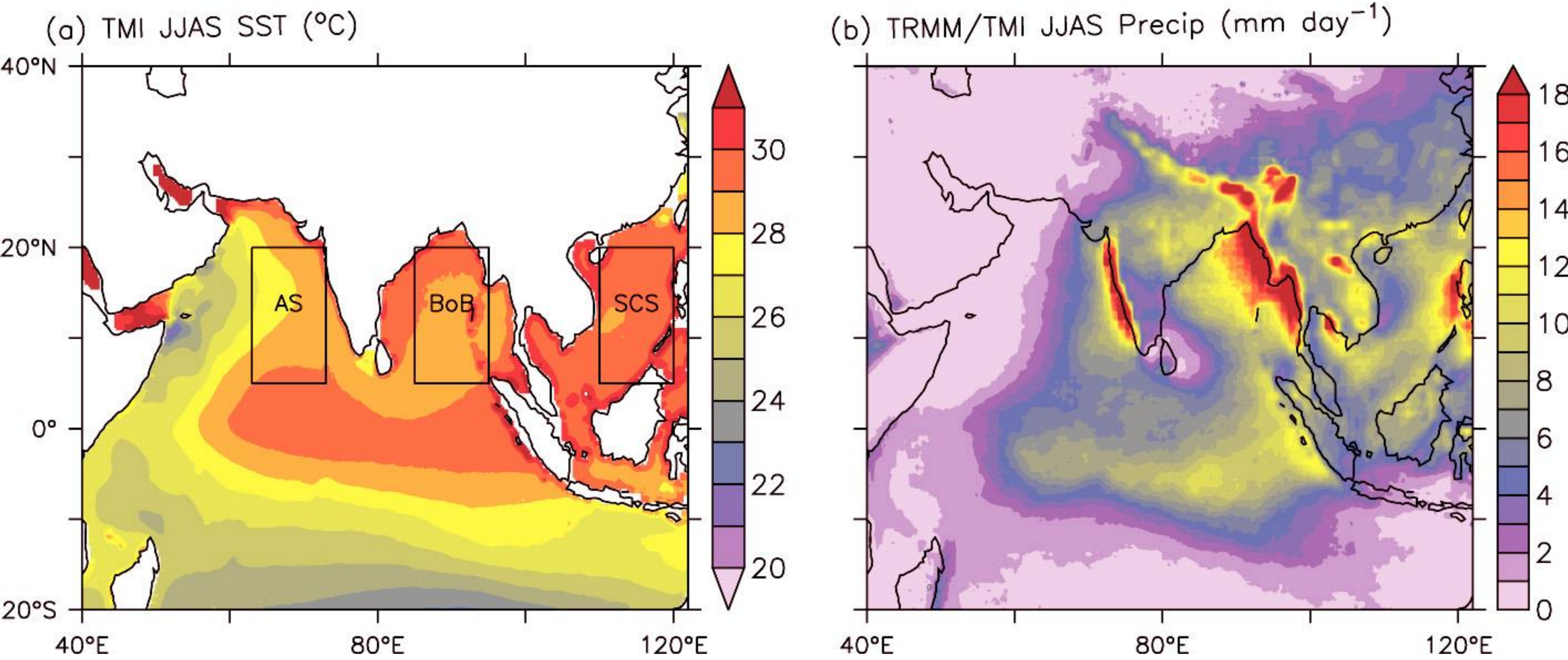
rain follows warmest SST;
surface winds converge
onto rainy/warm zones

El Niño

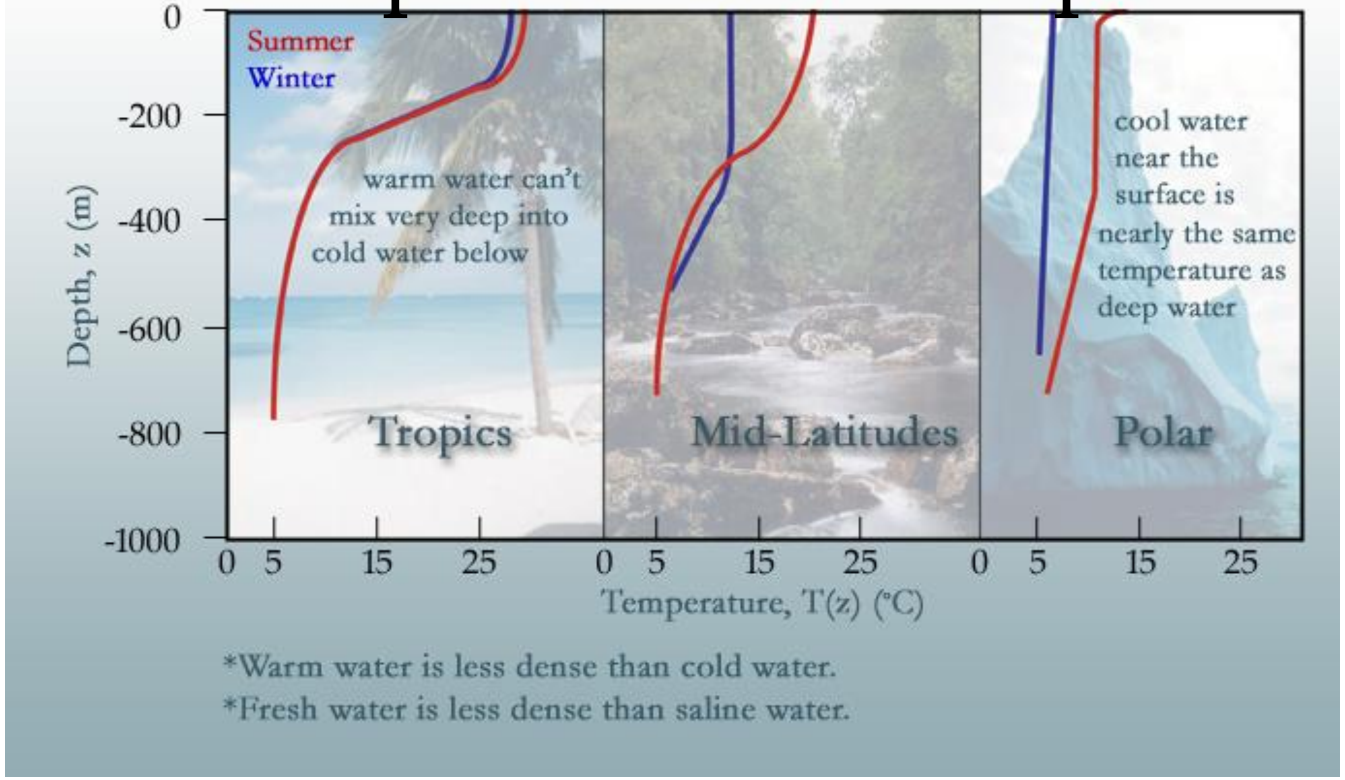


Events occur at
irregular intervals (2-8yr);
peak around Nov-Dec;
last about a year;
often followed by La Niña

SST over the Indian Ocean



Properties of Seawater
Vertical profiles of Temperature

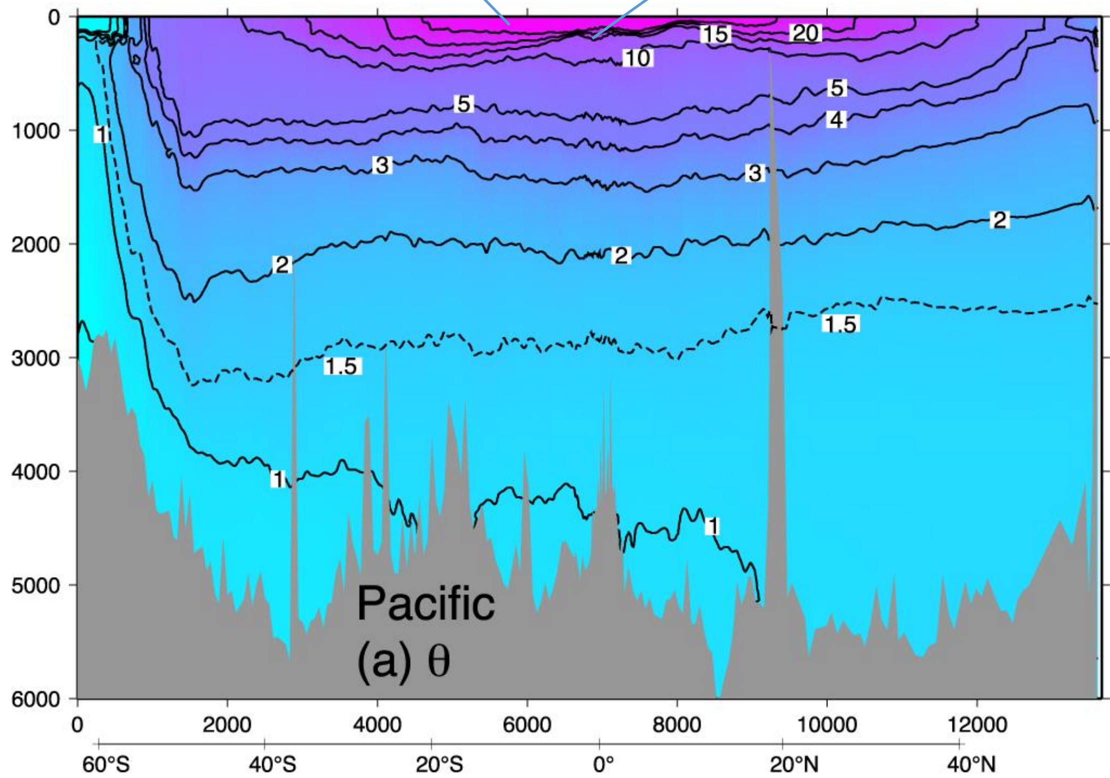


The region of maximum gradient (vertical) in temperature is called **thermocline**.
- the region between warm, light surface water and cold, dense deep water

Mixed layer is the region where temperature mixed and homogeneous to a certain level

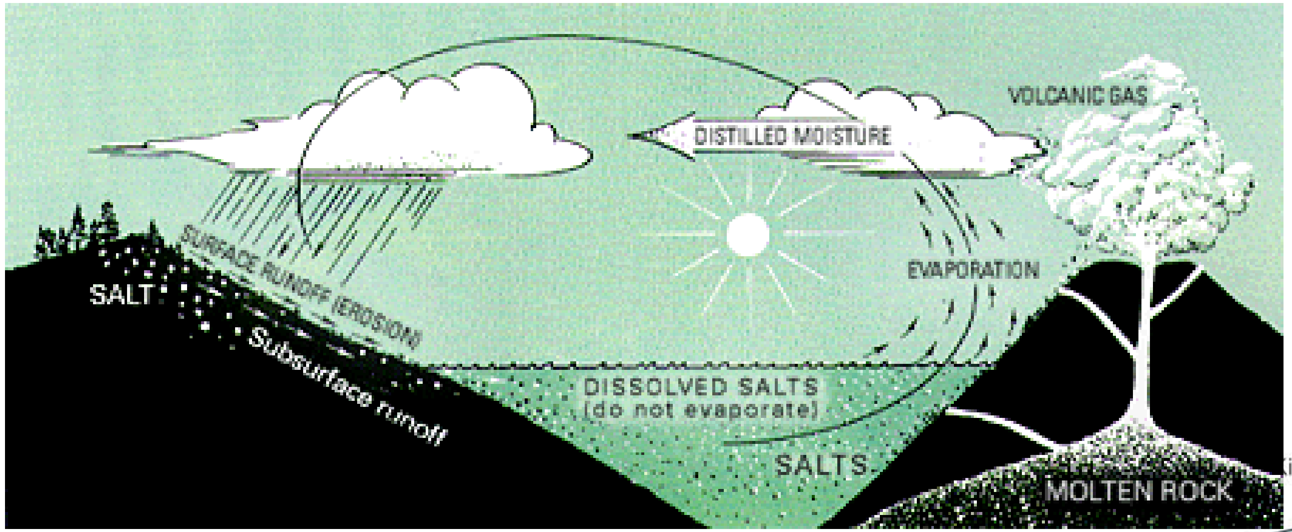
mixed layer

thermocline

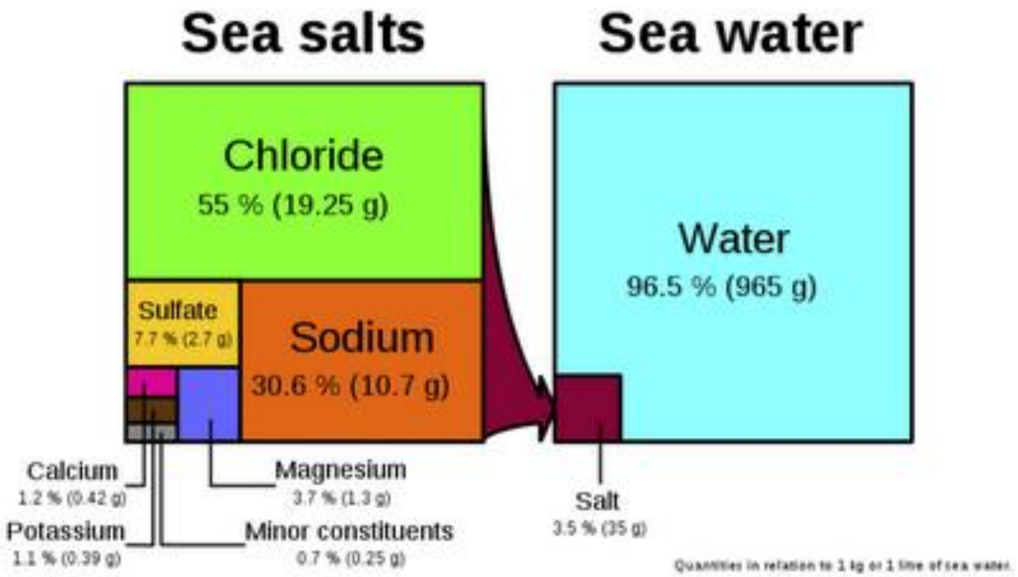


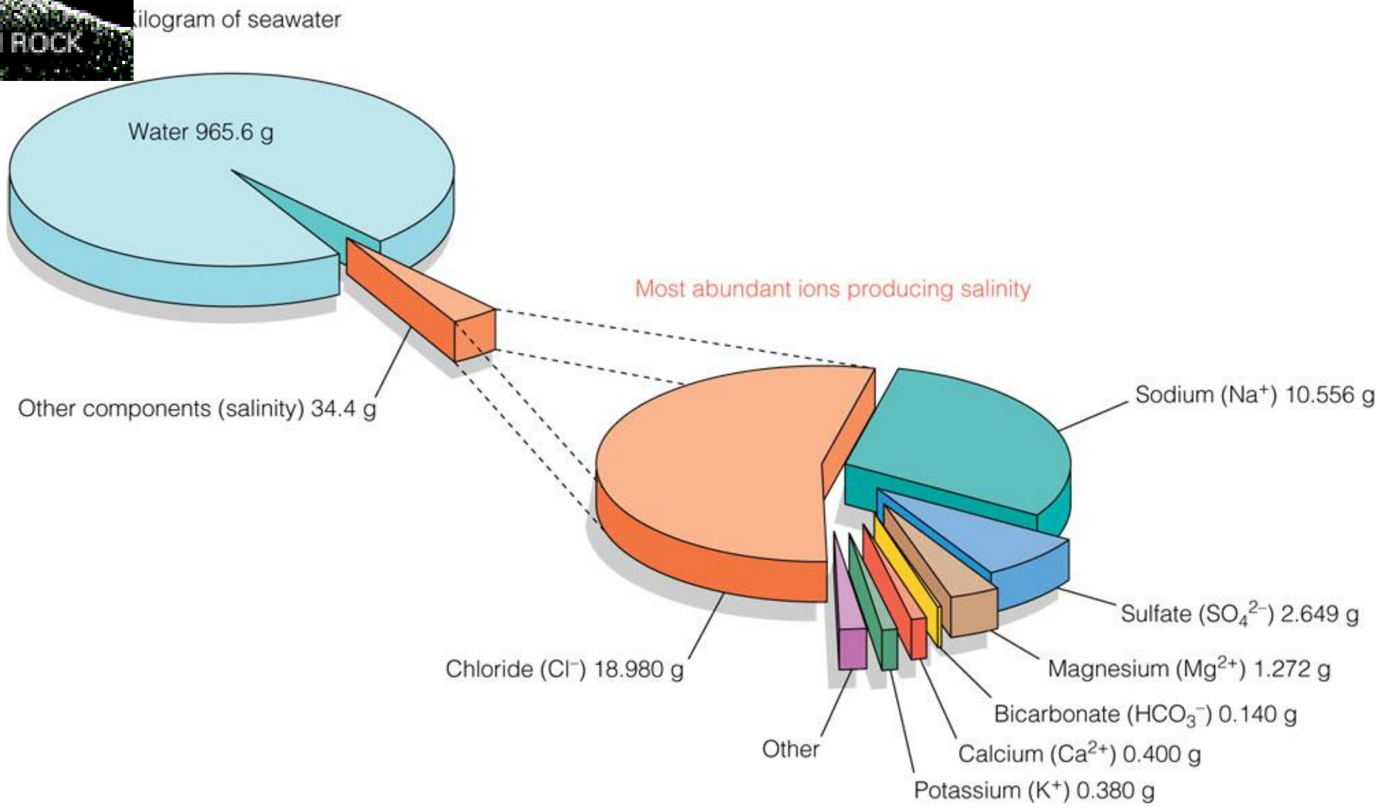
Ocean temperatures vary with depth and with latitude. In most cases, the ocean is stratified into warm surface water and cooler (denser) deep water. Except in polar regions, ocean water becomes markedly cooler with depth. Surface waters are warmest near the equator, but the temperature at 1 km depth varies little with latitude.
The average temperature of the global ocean is 3.6°C.

Why is the ocean salty?

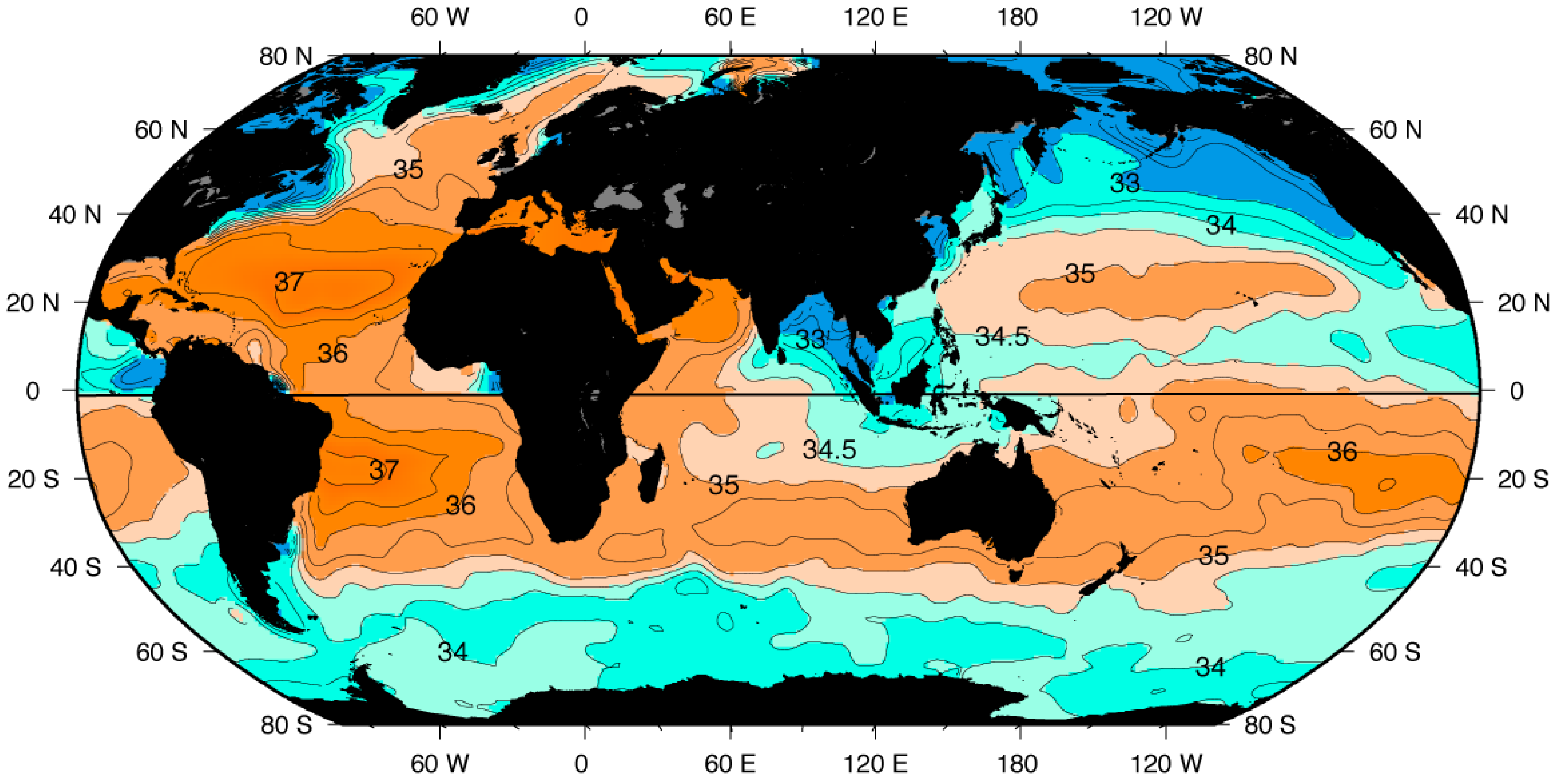


Conservation of salt: Salinity in a given region of the ocean can change due to river discharge, precipitation, freezing/thawing, and evaporation. However, the overall amount of salt in the oceans remains relatively constant, i.e., salinity is a conservative property of sea water.

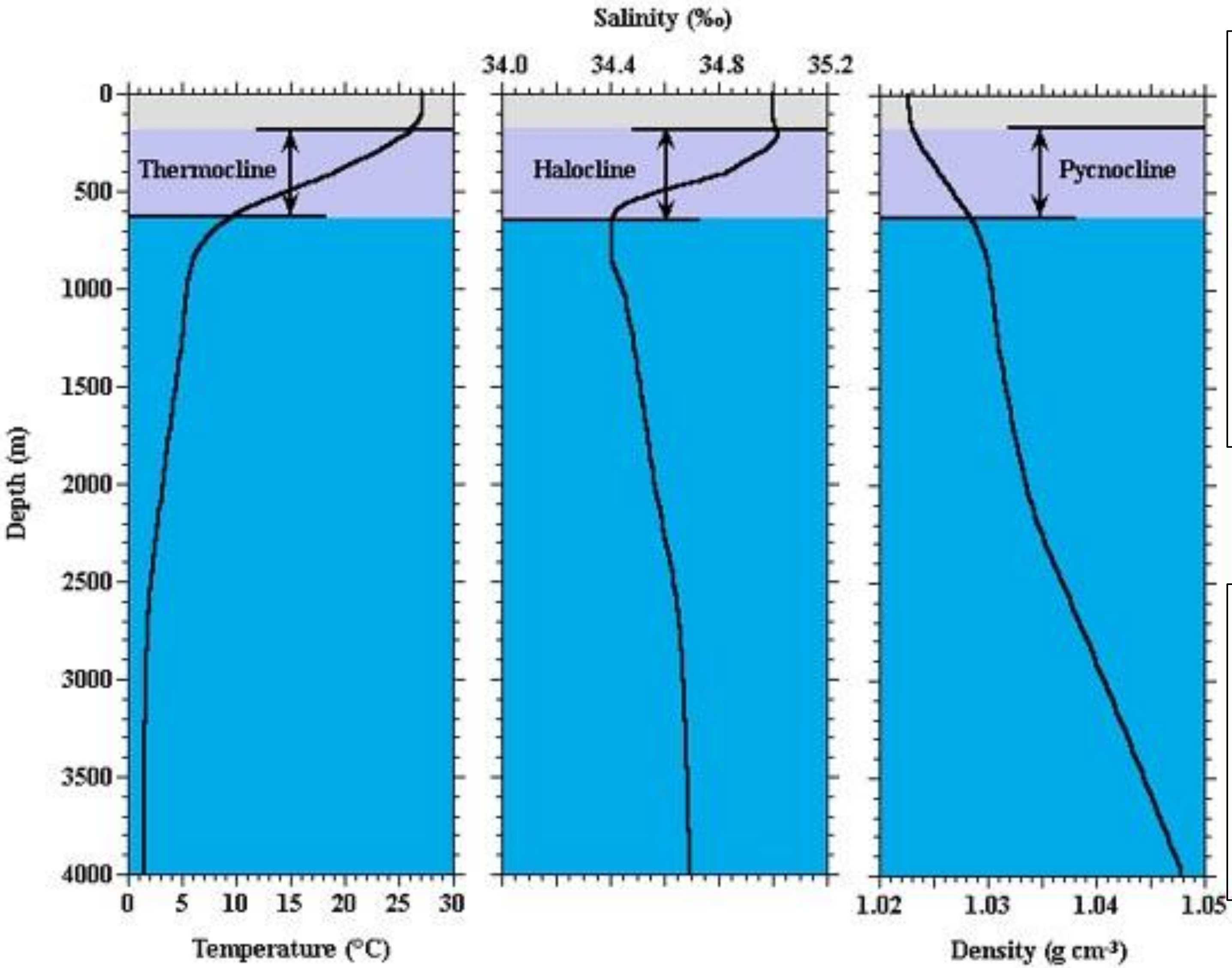




Properties of Seawater
Sea Surface Salinity



Properties of Seawater
Vertical profiles of Temp, Salinity and Density



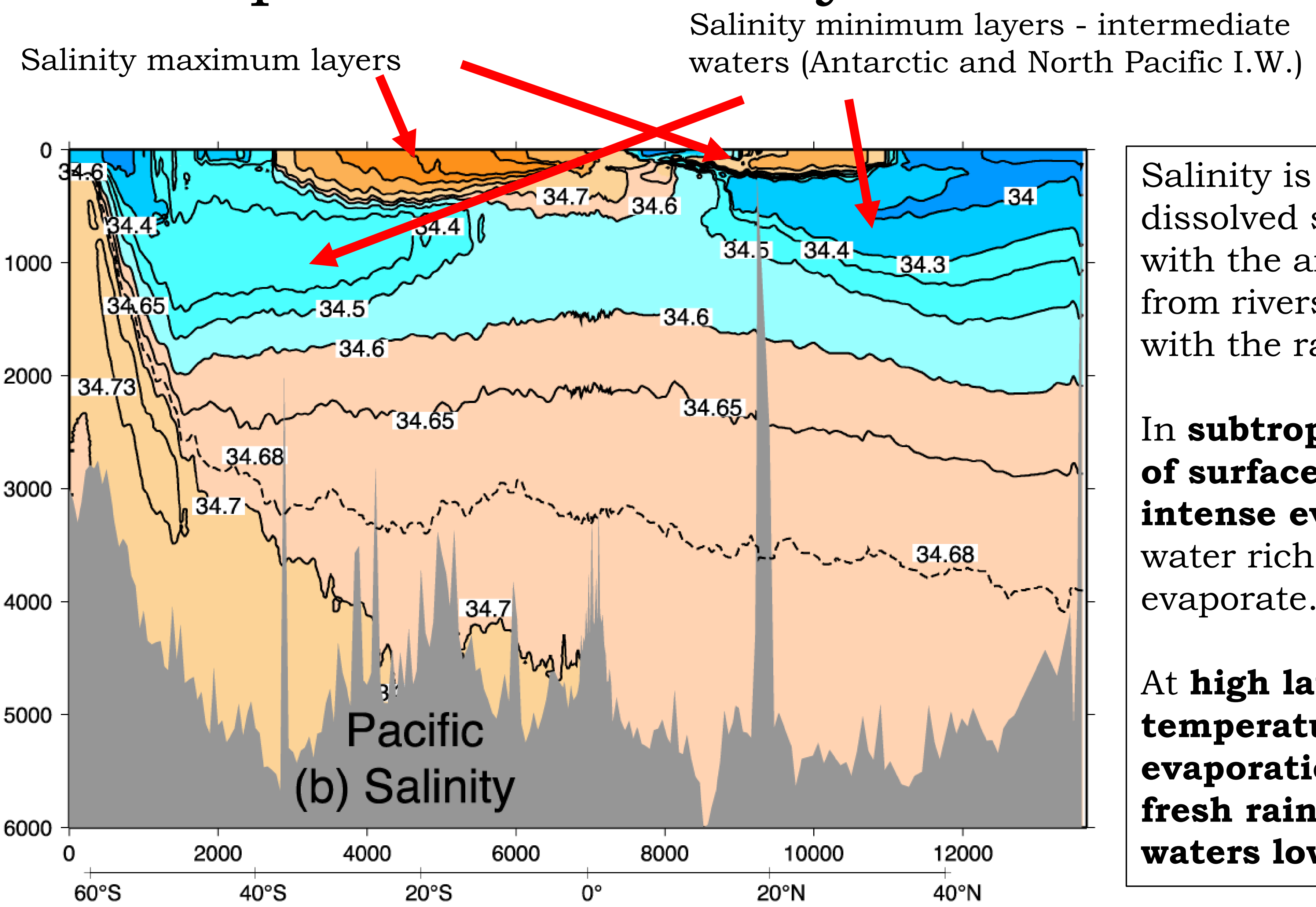
The region of maximum gradient (vertical) in salinity is called **halocline**.

Density - **pycnocline**.

Mixed layer is the region where the properties are mixed and homogeneous to a certain level.

There are two principal layers of oceanic water:
(1) a thin upper layer of warm, well-stirred water and
(2) a thick mass of deeper, colder water that is relatively calm and slow-moving.

Properties of Seawater
Vertical profiles of Salinity

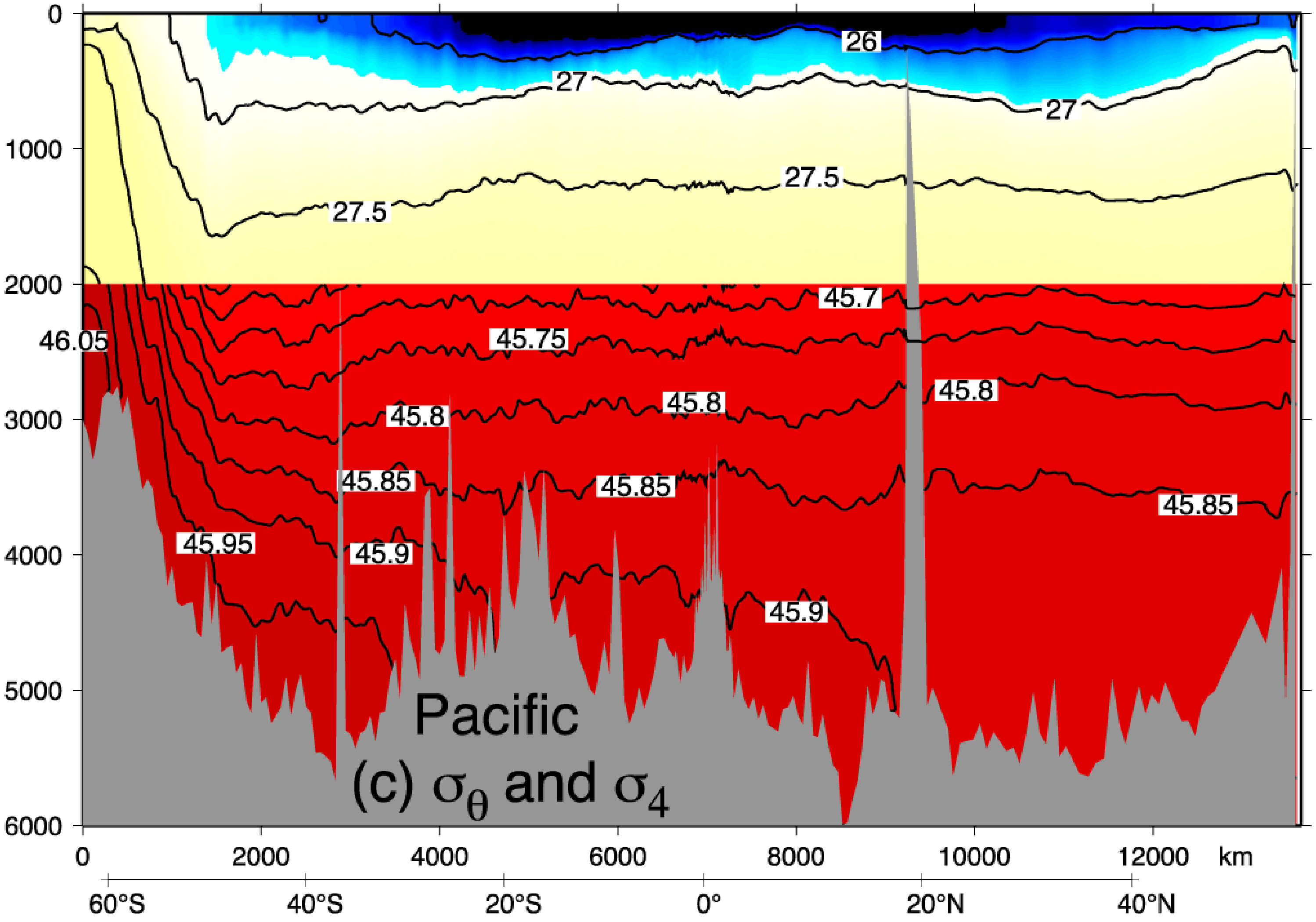


Salinity is a measure of all of the dissolved salts in seawater. It varies with the amount of freshwater input from rivers or melting glaciers and with the rate of evaporation.

In **subtropical regions, the salinity of surface water is high because intense evaporation** leaves the water rich in salts that cannot evaporate.

At **high latitudes where the temperature is lower, the evaporation rate is much lower and fresh rainwater makes the surface waters low in salinity**.

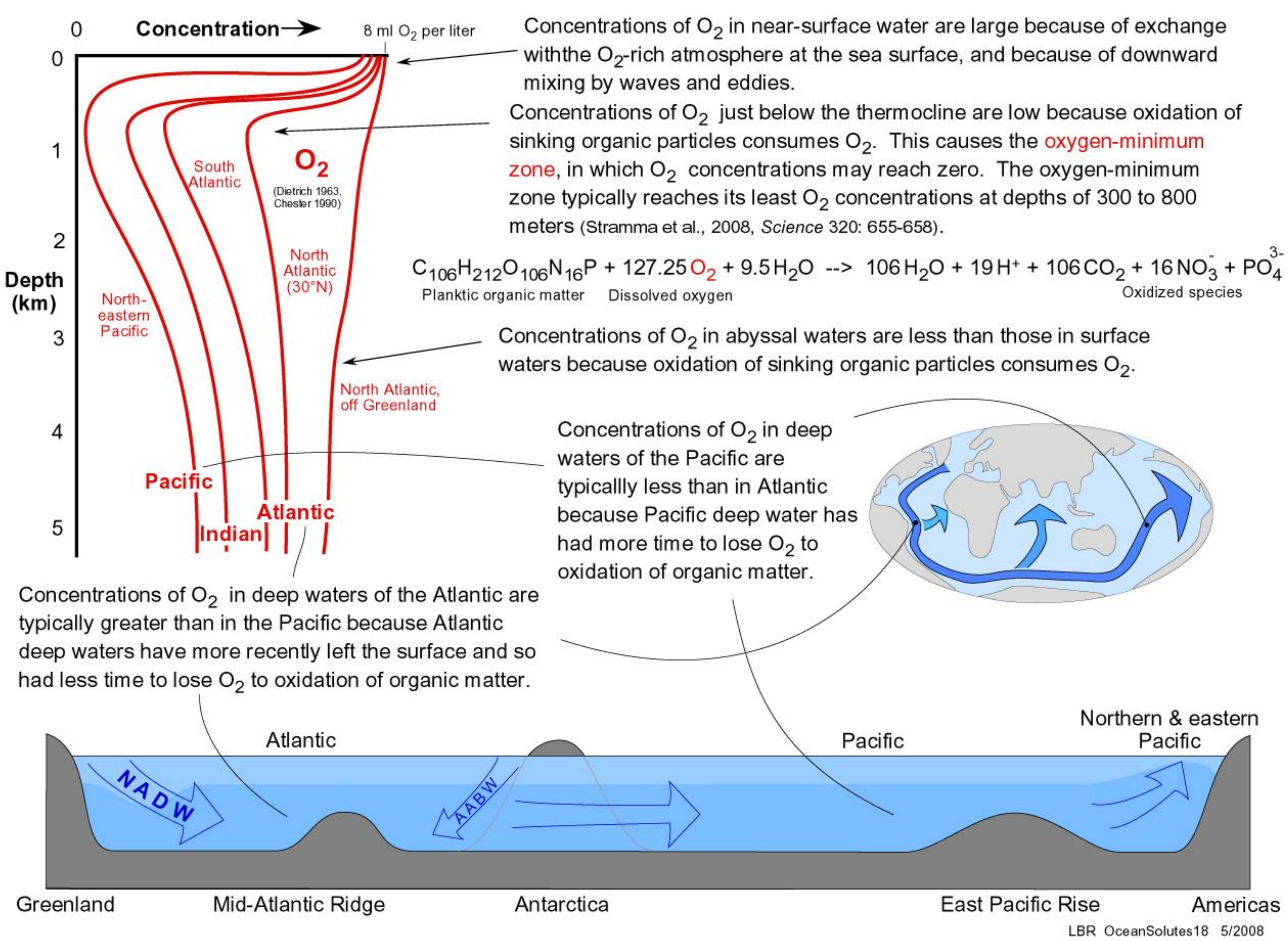
Properties of Seawater
Vertical profiles of Potential Density



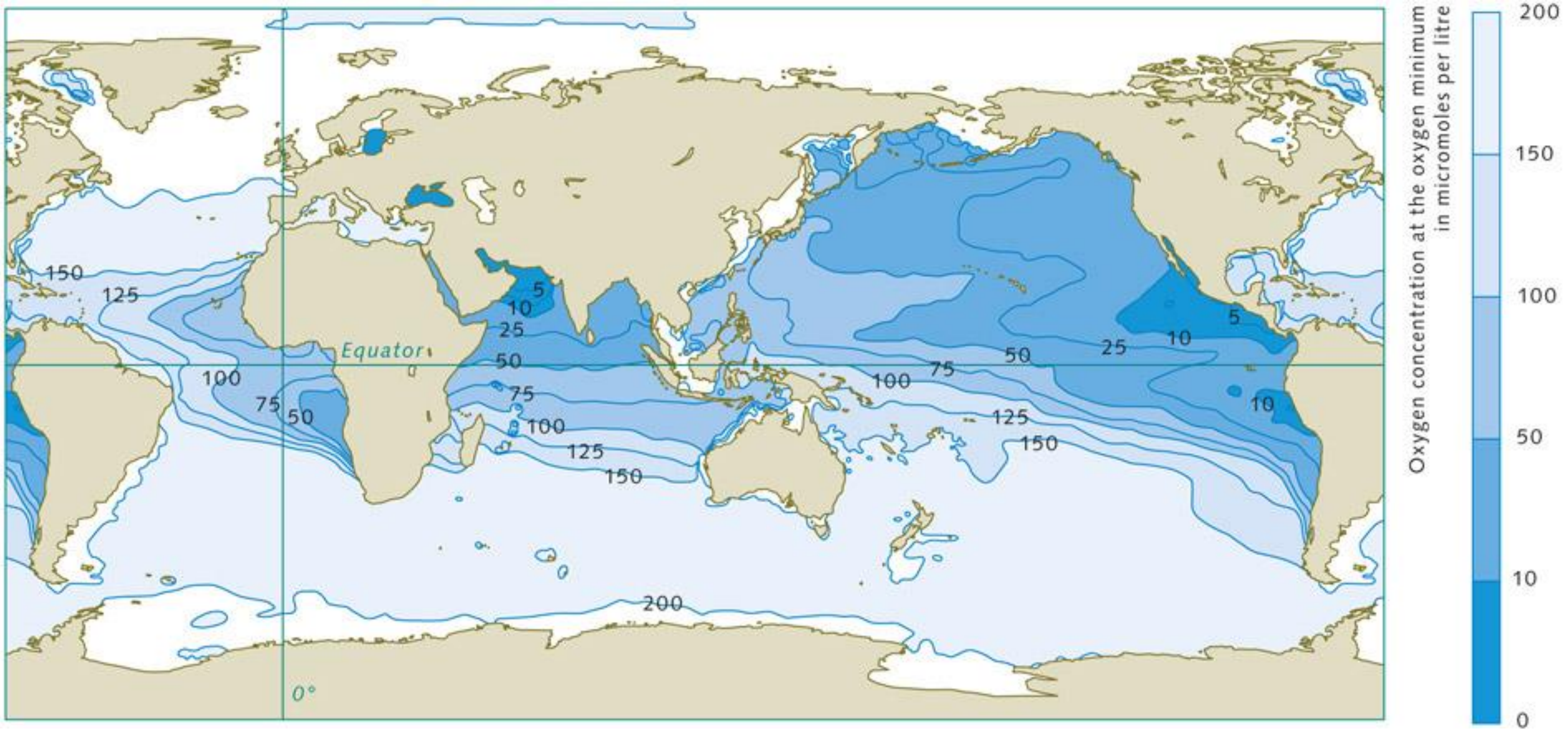
For the normal range of salinity in the ocean, the maximum density occurs at the freezing point, which is depressed to well below 0C.

As seawater is heated, molecular activity increases and thermal expansion occurs, reducing the density.

Properties of Seawater
Vertical profiles of Oxygen

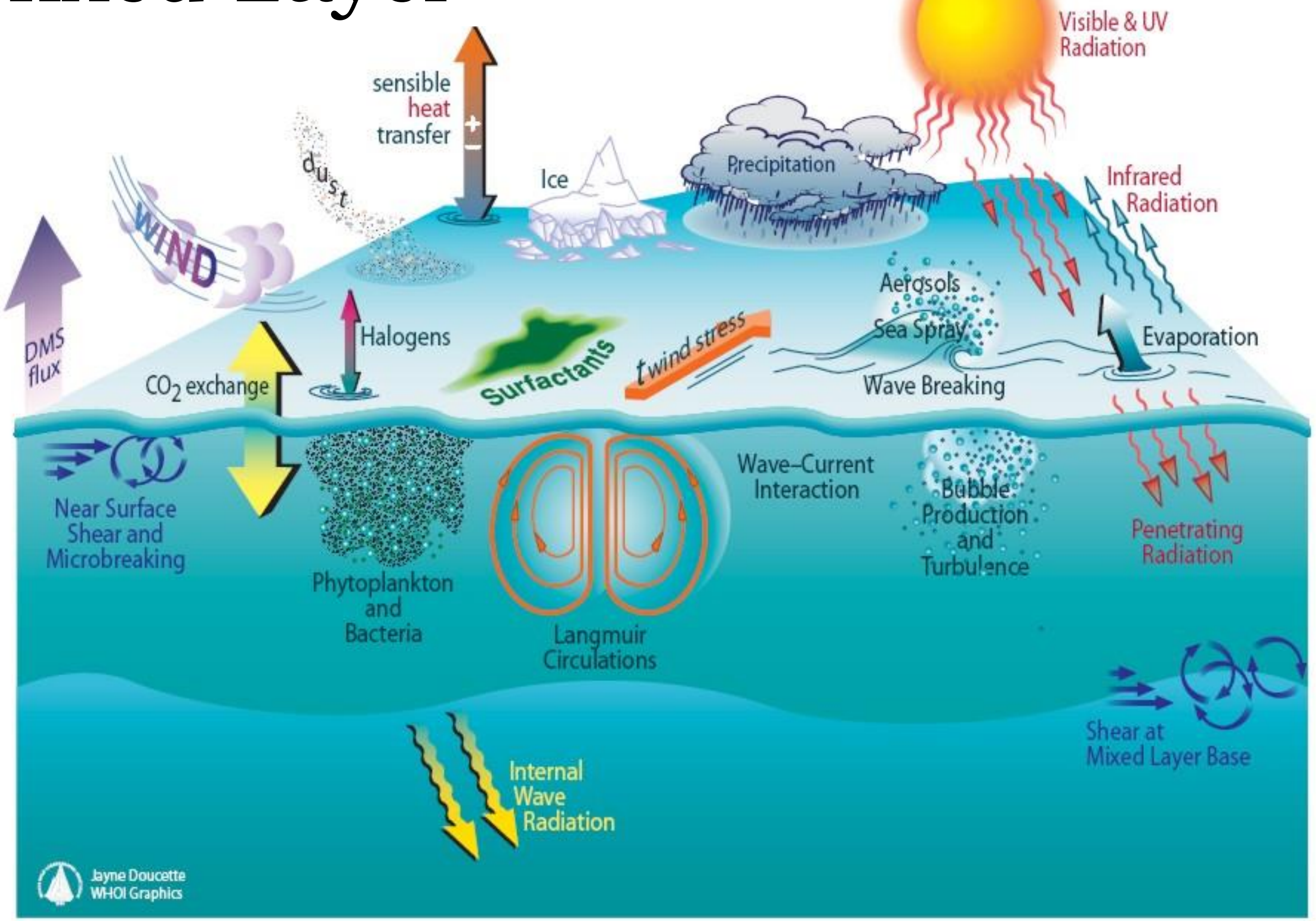


Properties of Seawater
Oxygen at oxygen minimum zone



Marine regions with oxygen deficiencies are mainly located in the mid-latitudes on the west sides of the continents. There is very little mixing here of the warm surface waters with the cold deep waters, so not much oxygen penetrates to greater depths. In addition, high bioproductivity and the resulting large amounts of sinking biomass here lead to strong oxygen consumption at depth, especially between 100 and 1000 metres.

Mixed Layer



Schematic of processes operating at the air-sea interface and in the upper ocean mixed layer

Mixed Layer Depth [MLD]: The term "mixed" refers to a given physical parameter of the ocean state (e.g. temperature, density…) that is assumed to be mixed and homogeneous to a certain level (e.g. regarding some space/time scales), from the surface down to the considered MLD.

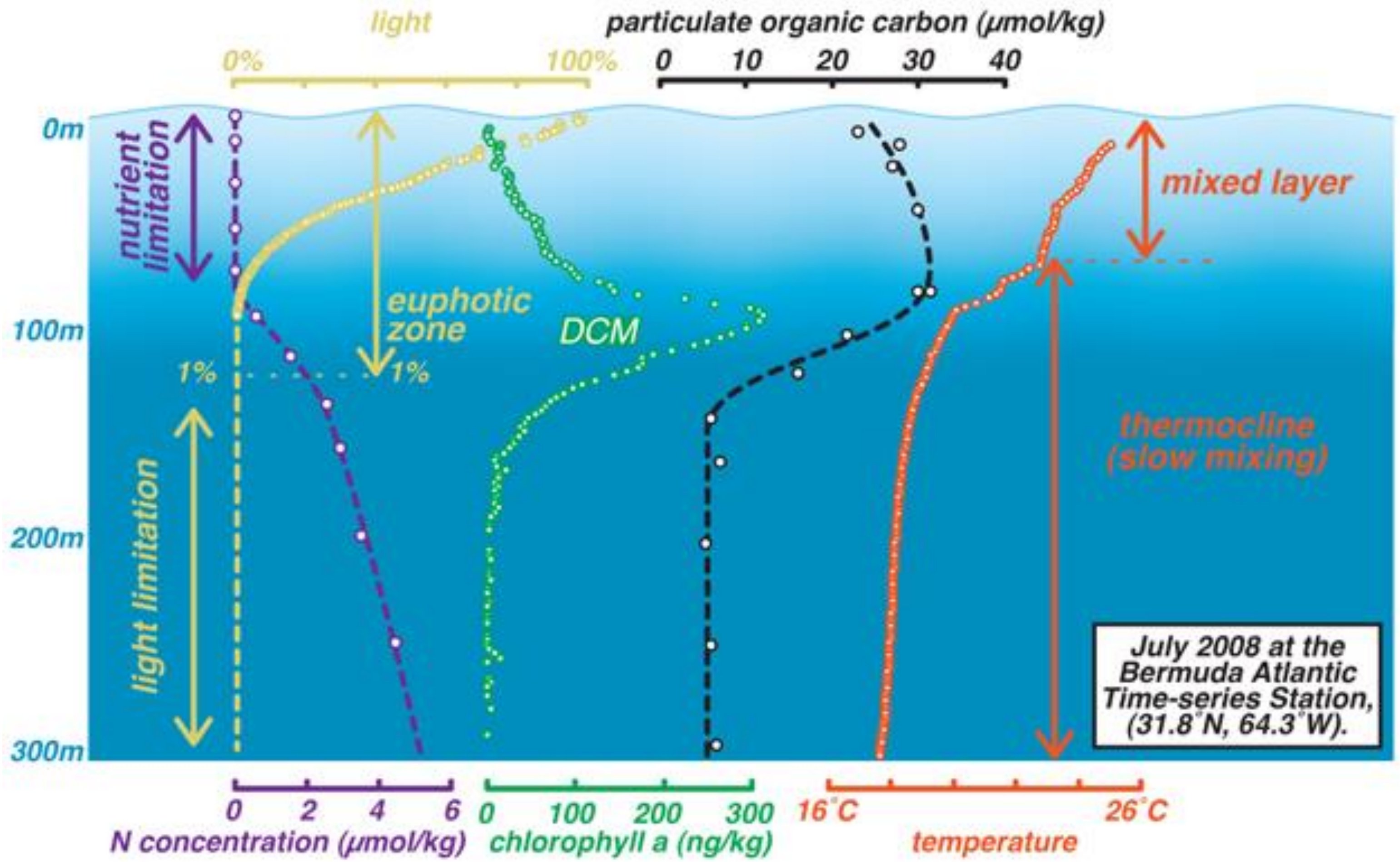
1. Since the specific heat of ocean water is much larger than that of air, the **top 2.5 m of the ocean holds as much heat as the entire atmosphere above it**. Thus the heat required to change a mixed layer of 25 m by 1 °C would be sufficient to raise the temperature of the atmosphere by 10 °C. The depth of the mixed layer is thus very important for determining the temperature range in oceanic and coastal regions.

2. The heat stored within the oceanic mixed layer provides a source for heat that drives global variability such as El Niño.

3. The mixed layer is also important as its depth determines the average level of light seen by marine organisms. In very deep mixed layers, the tiny marine plants known as phytoplankton are unable to get enough light to maintain their metabolism.

Mixing:
1. Wind driven - Waves/Currents
2. Density driven - cooling/evaporation/runoff

Vertical profile of Biological Productivity



The export of organic matter to depth depletes the surface ocean of nutrients, causing the nutrients to accumulate in deep waters where there is no light available for photosynthesis (Figure 2). Because of the density difference between surface water and the deep sea across most of the ocean, ocean circulation can only very slowly reintroduce dissolved nutrients to the euphotic zone.

Physical properties of seawater - summary

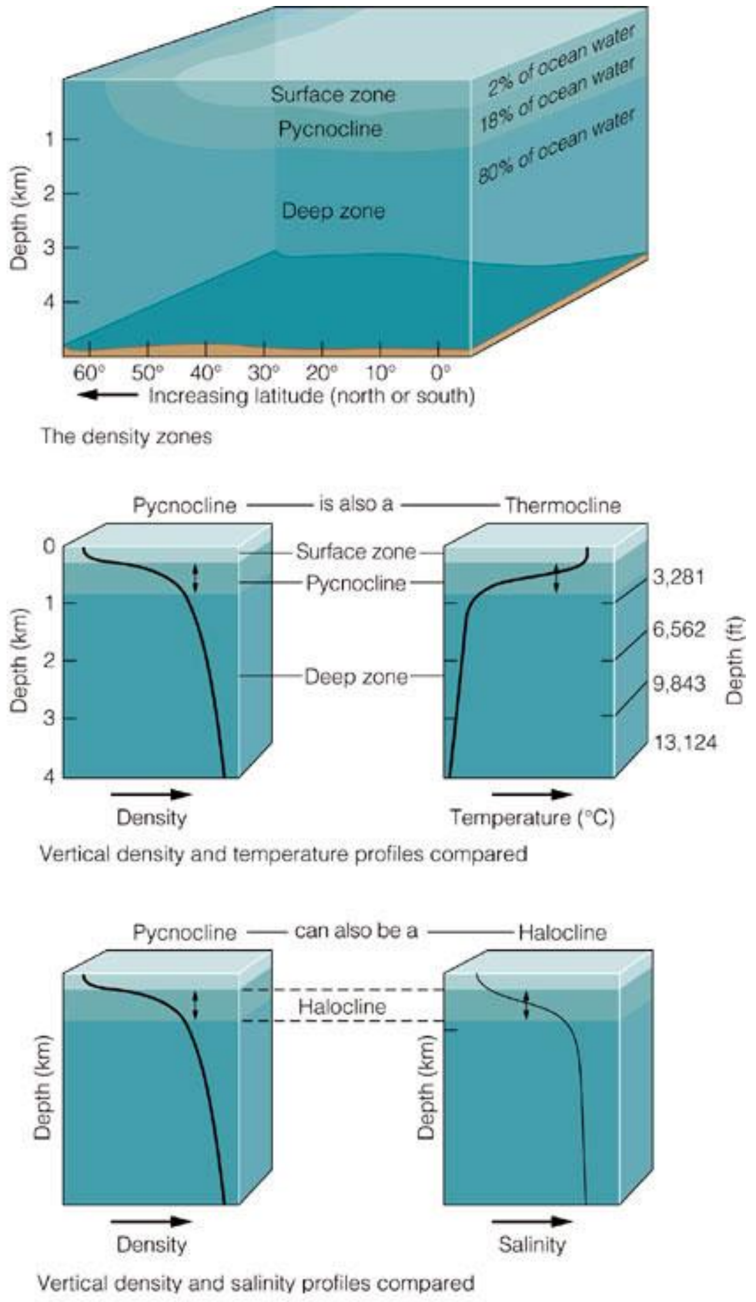
Mixed layer

Thermocline, halocline, pycnocline:

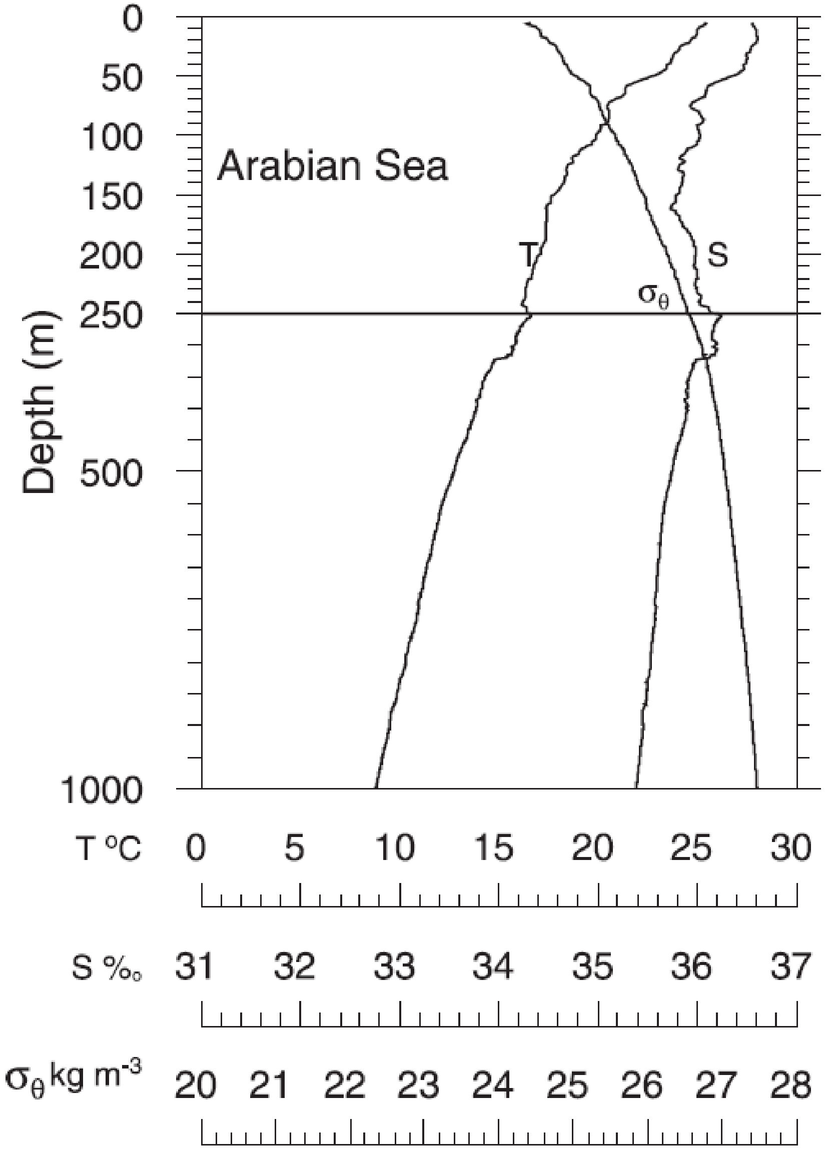
Vertical locations of high gradient
(large $\Delta T/\Delta z$, for thermocline, etc.)

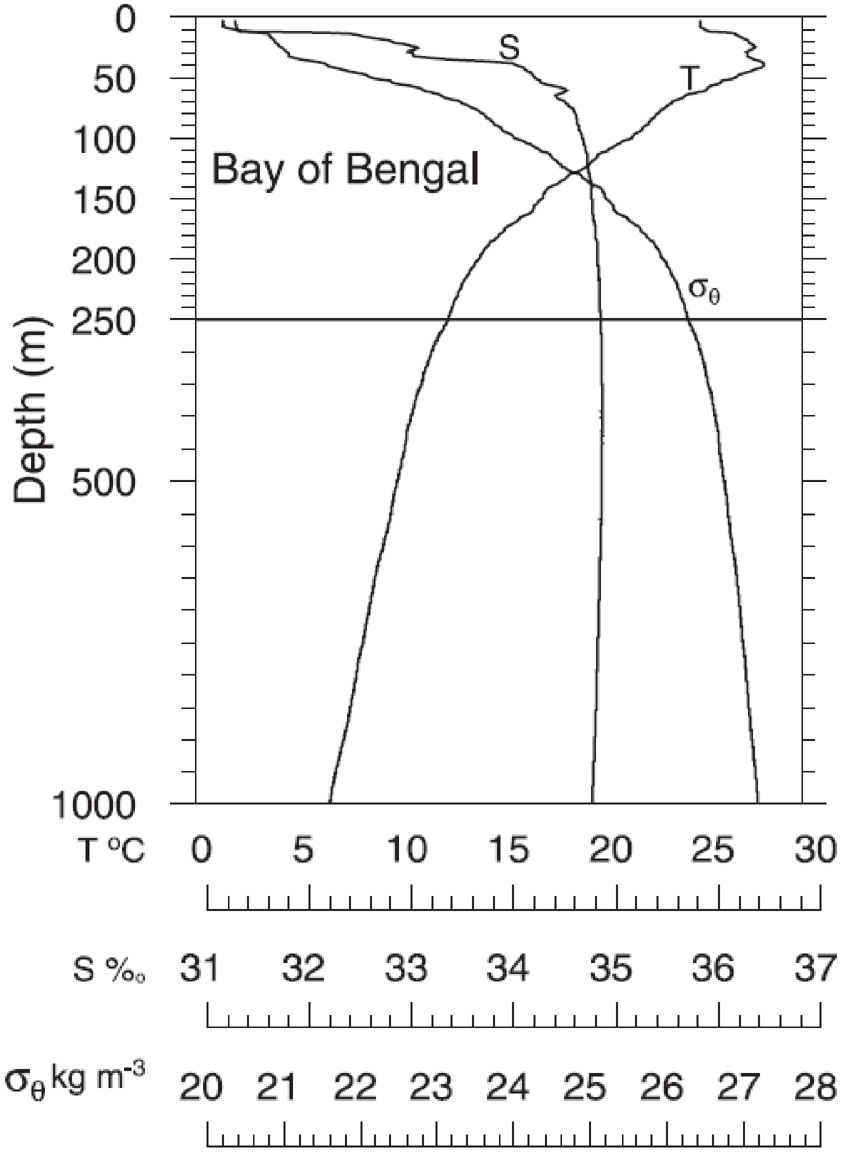
Vertical extrema sometimes have names:
salinity minima, temperature minima or maxima, etc. (e.g. dichothermal layer for very shallow temperature minimum usually in high latitudes

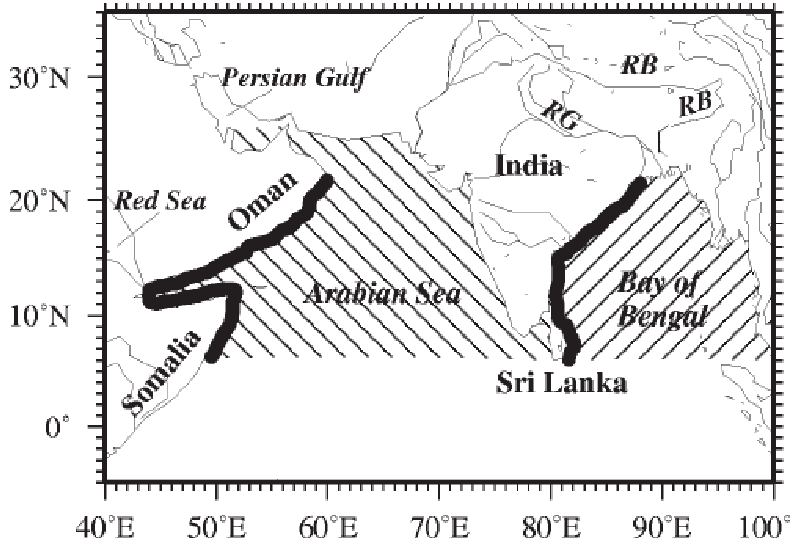
Variable	Ocean Range	Ocean Mean	Required Accuracy
Temperature	-2°C to 40°C	3.5°C	±0.002°C
Salinity	0 g/kg to 42 g/kg	34.9 g/kg	±0.002 g/kg
Pressure	0 dbar to 11000 dbar	1850 dbar	< ±3 dbar
In-situ Density	1000 to 1060 kg/m^3	1036 kg/m^3	±0.004 kg/m^3

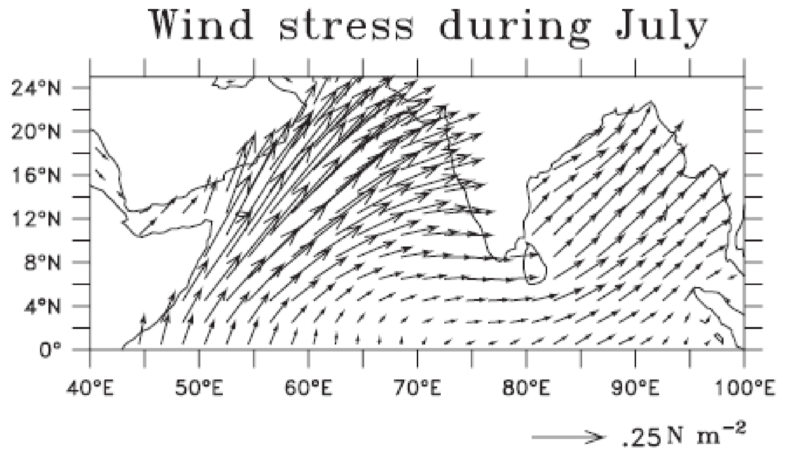


Arabian Sea and Bay of Bengal - Thermodynamics

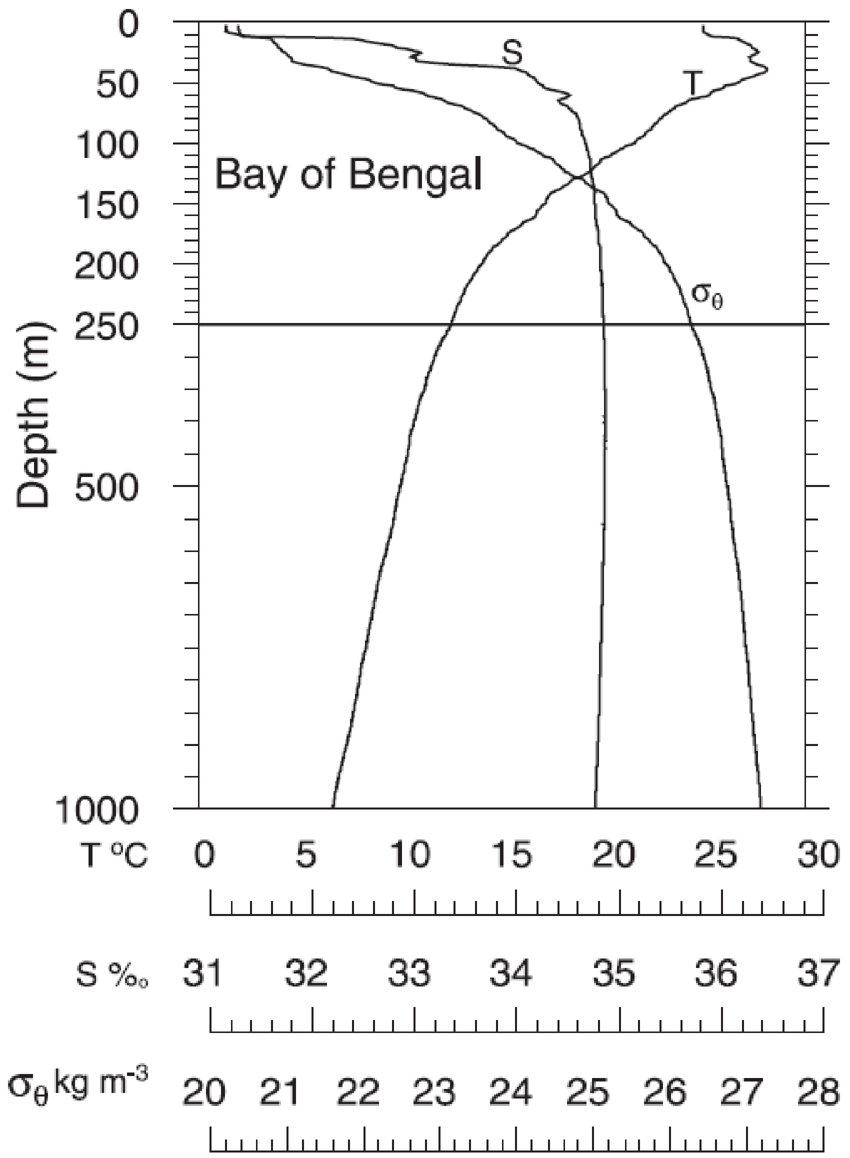








Temperature inversion over Bay of Bengal

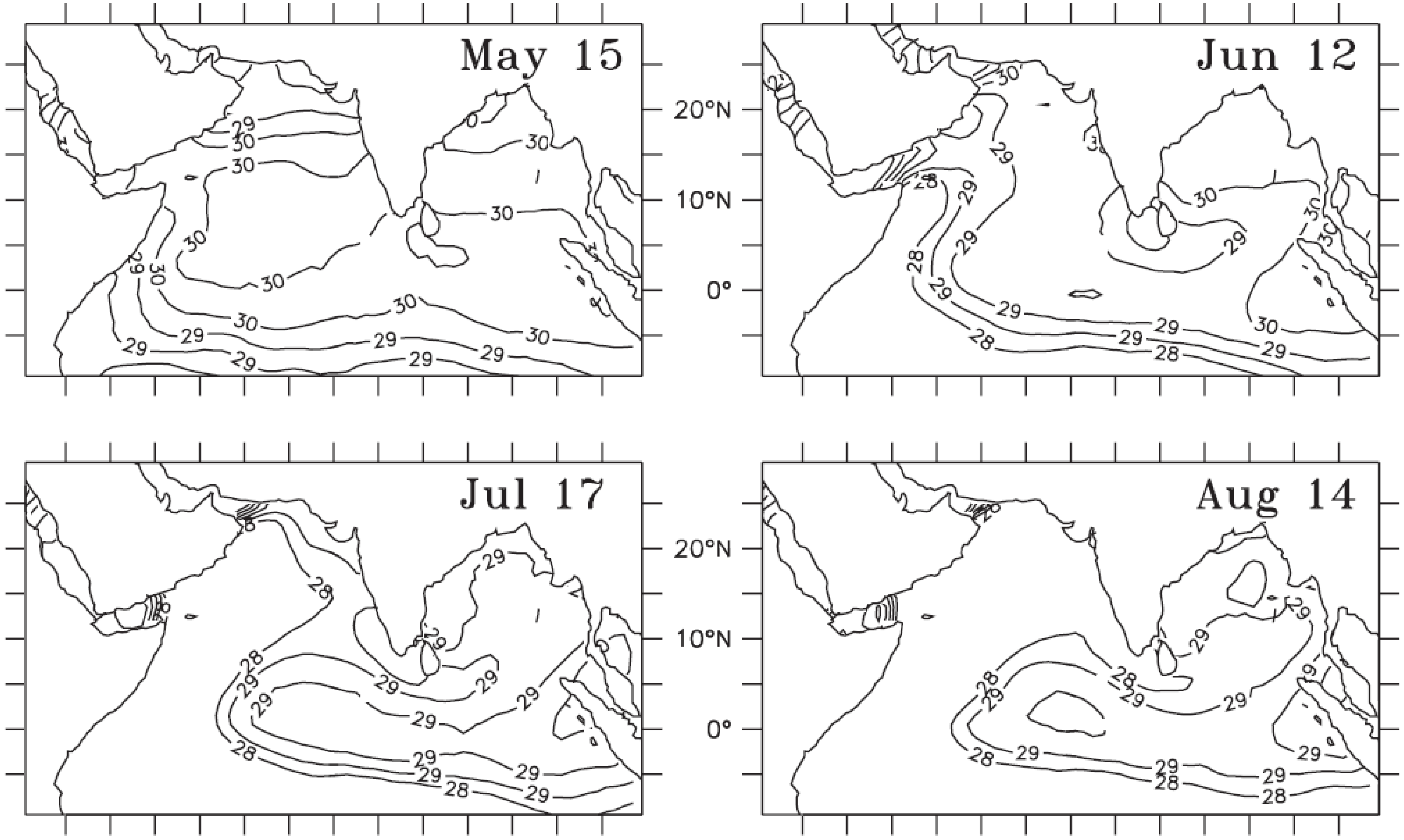


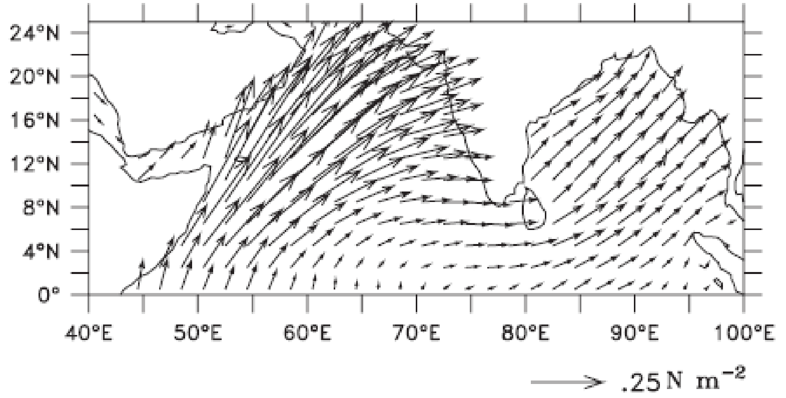
Cold – fresh water stratification

1. Net surface heat loss and associate surface cooling.
2. Advection of cold-fresher water over warm saline water.

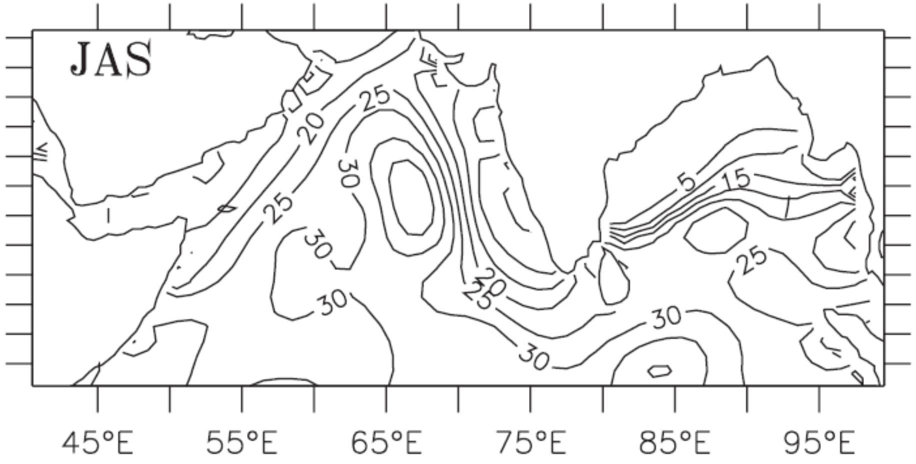
Arabian Sea and Bay of Bengal - Thermodynamics and role on Climate

Evolution of SST during summer monsoon



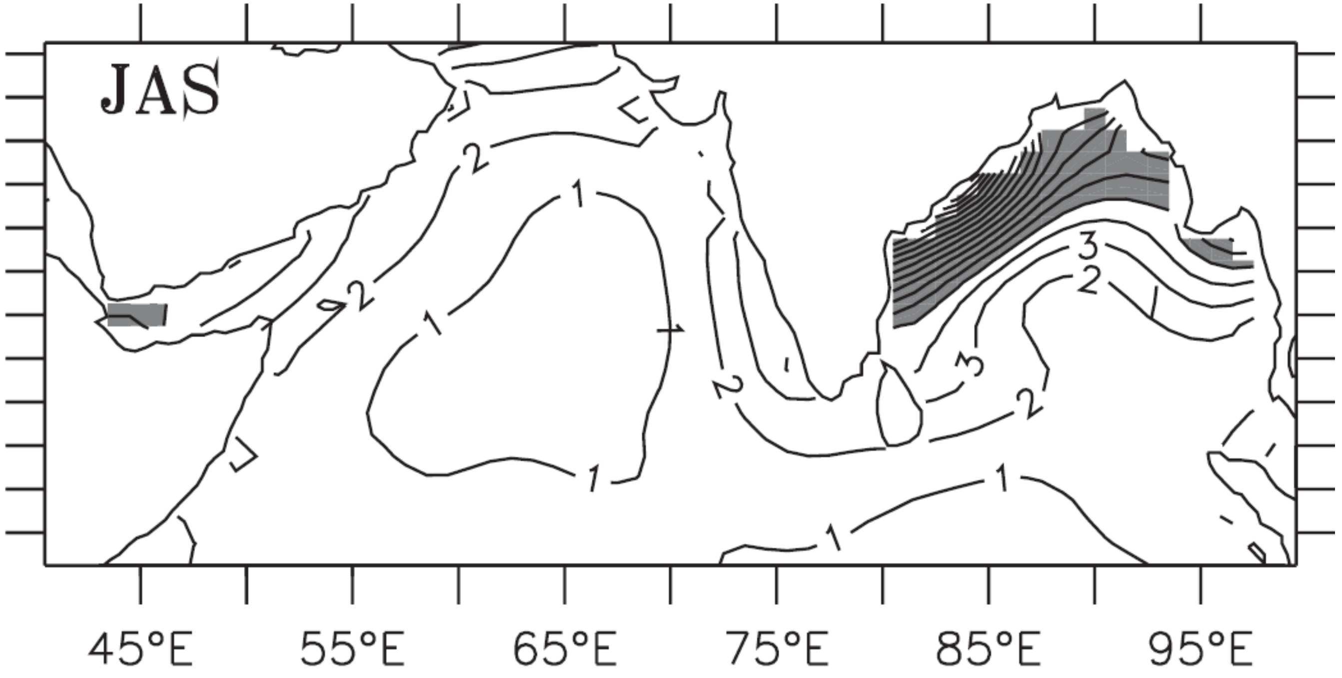


Mixed Layer Depth during summer monsoon

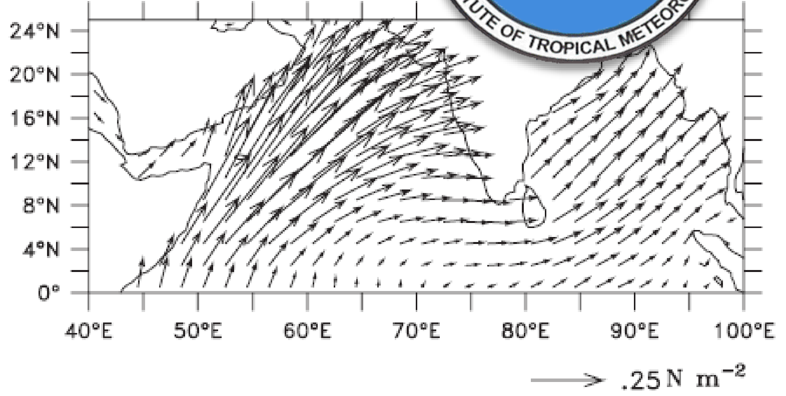


Arabian Sea and Bay of Bengal - Thermodynamics and role on Climate

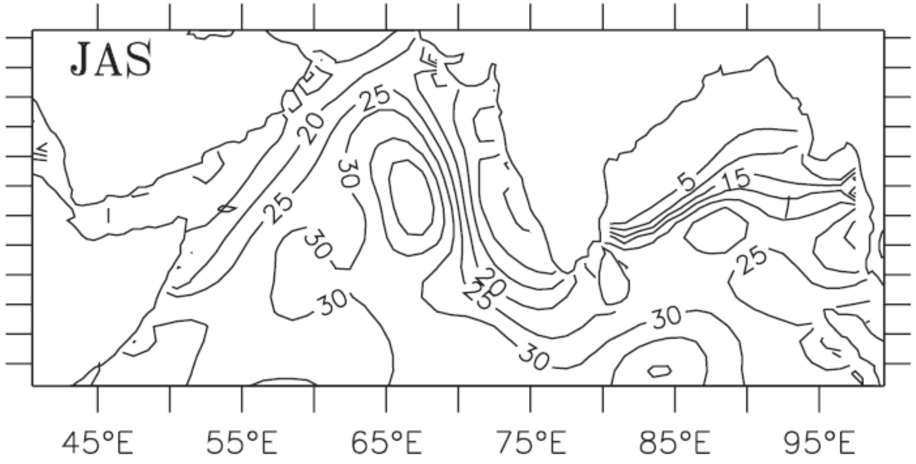
Energy required for mixing (10^3 J/m^2)



Wind stress



Mixed Layer Depth during summer monsoon

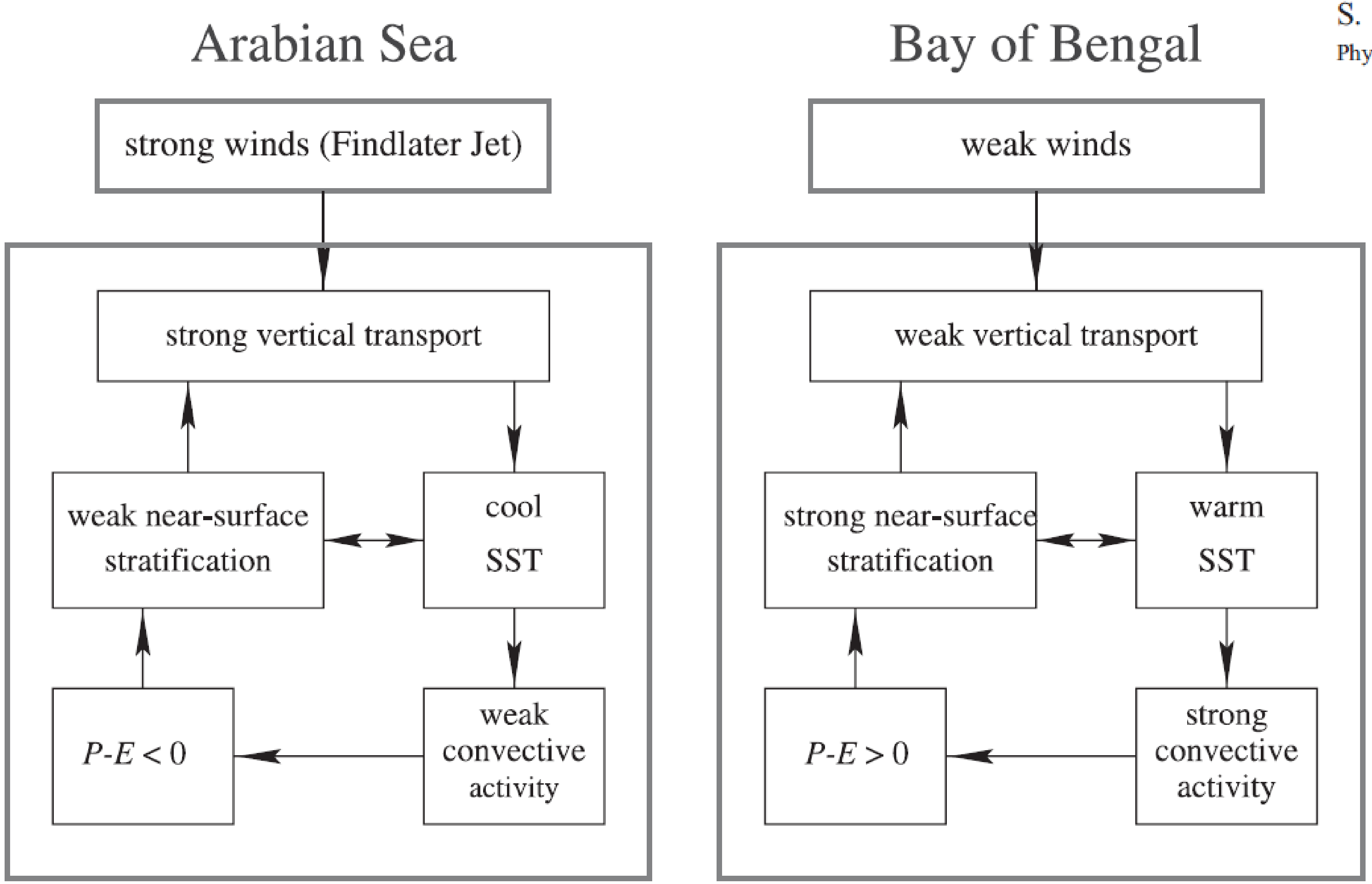


The energy required for mixing (ERM) the water column to 50 m is the difference between the potential energy of a stratified 50 m column and that of the same column when it is unstratified (or mixed vertically).
ERM in the northern bay is >12, but it is <3 in the Arabian Sea. The large difference between the two basins is due to the low salinity of the surface waters in the Bay.

Arabian Sea and Bay of Bengal - Thermodynamics and role on Climate

Differences in heat budgets of the near-surface Arabian Sea and Bay of Bengal: Implications for the summer monsoon

S. S. C. Shenoi, D. Shankar, and S. R. Shetye
Physical Oceanography Division, National Institute of Oceanography, Goa, India



Dynamics of Atmospheres and Oceans 49 (2010) 108–123

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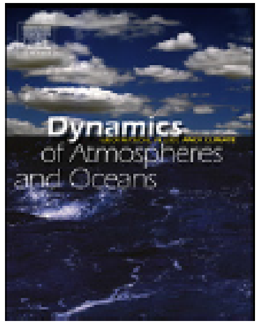

The role of mean ocean salinity in climate

Paul D. Williams[a,*], Eric Guilyardi[a,b], Gurvan Madec[b], Silvio Gualdi[c], Enrico Scoccimarro[c]

[a] National Centre for Atmospheric Science, Department of Meteorology, University of Reading, UK
[b] Laboratoire d'Océanographie et de Climat: Expérimentation et Approche Numérique (LOCEAN/IPSL), CNRS/Université Paris VI, France
[c] Istituto Nazionale di Geofisica e Vulcanologia, Bologna, Italy

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The Dead Sea hydrography from 1992 to 2000

I. Gertman*, A. Hecht

Israel Oceanographic and Limnological Research, National Institute of Oceanography, P.O. Box 8030, Haifa, Israel

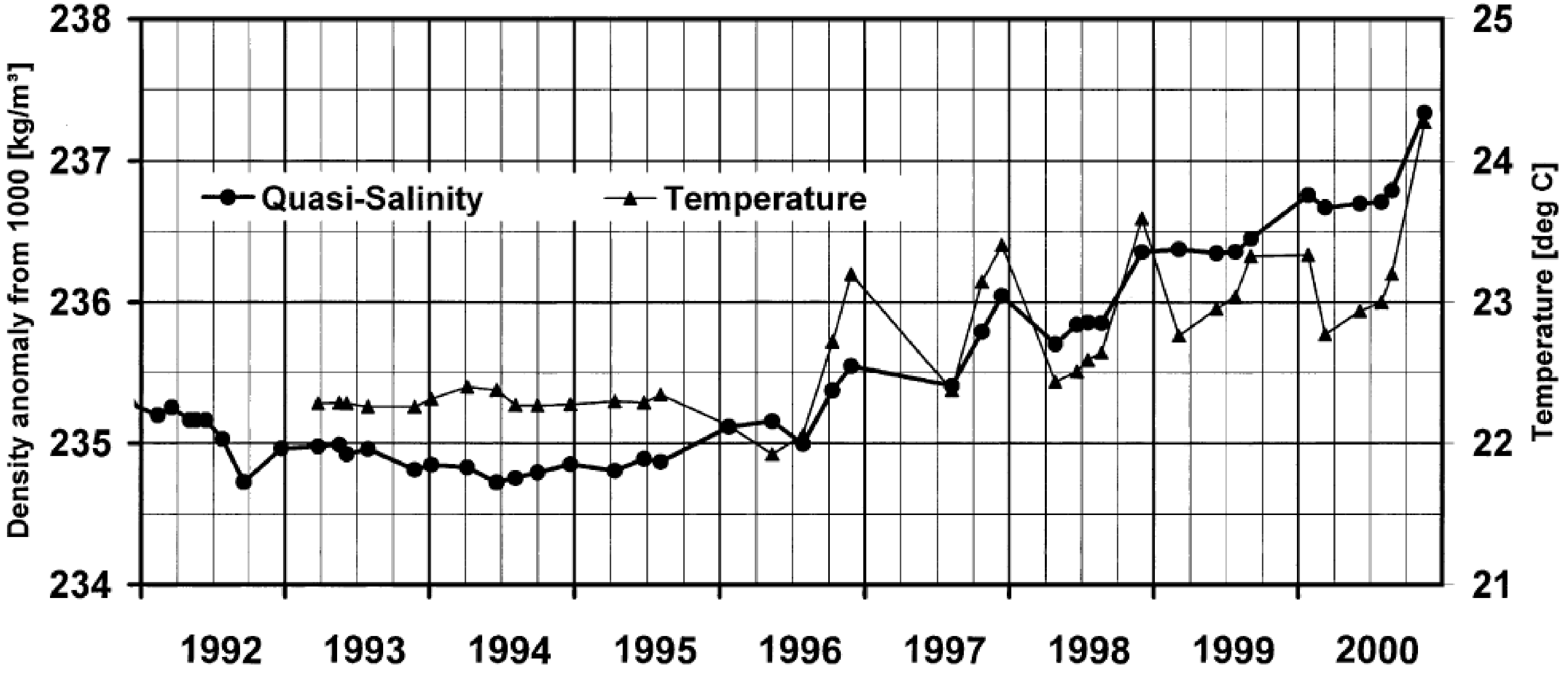


Fig. 8. Long-term changes of temperature and quasi-salinity of the Dead Sea deep water body (below 100 m).

General Circulation – Wind Driven and Thermohaline

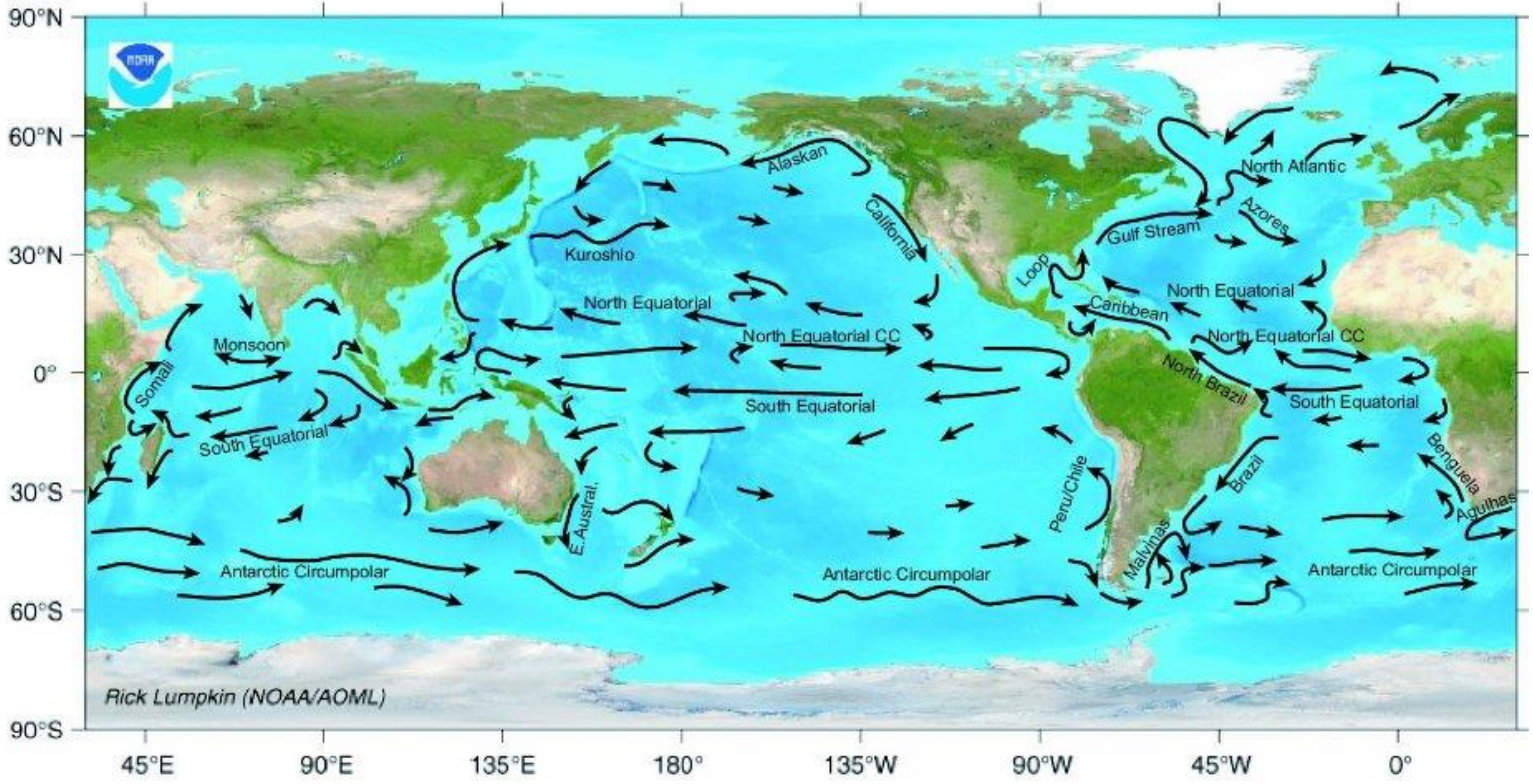
Wind-driven circulation:

Driven by wind.
Most vigorous in the upper ocean down to about 500 to 2000 m, depending on location (deeper, to bottom, near western boundaries).
Space scale: mostly ocean basin.
Time scale: weak seasonal variation to climate time scales.

Thermohaline circulation:

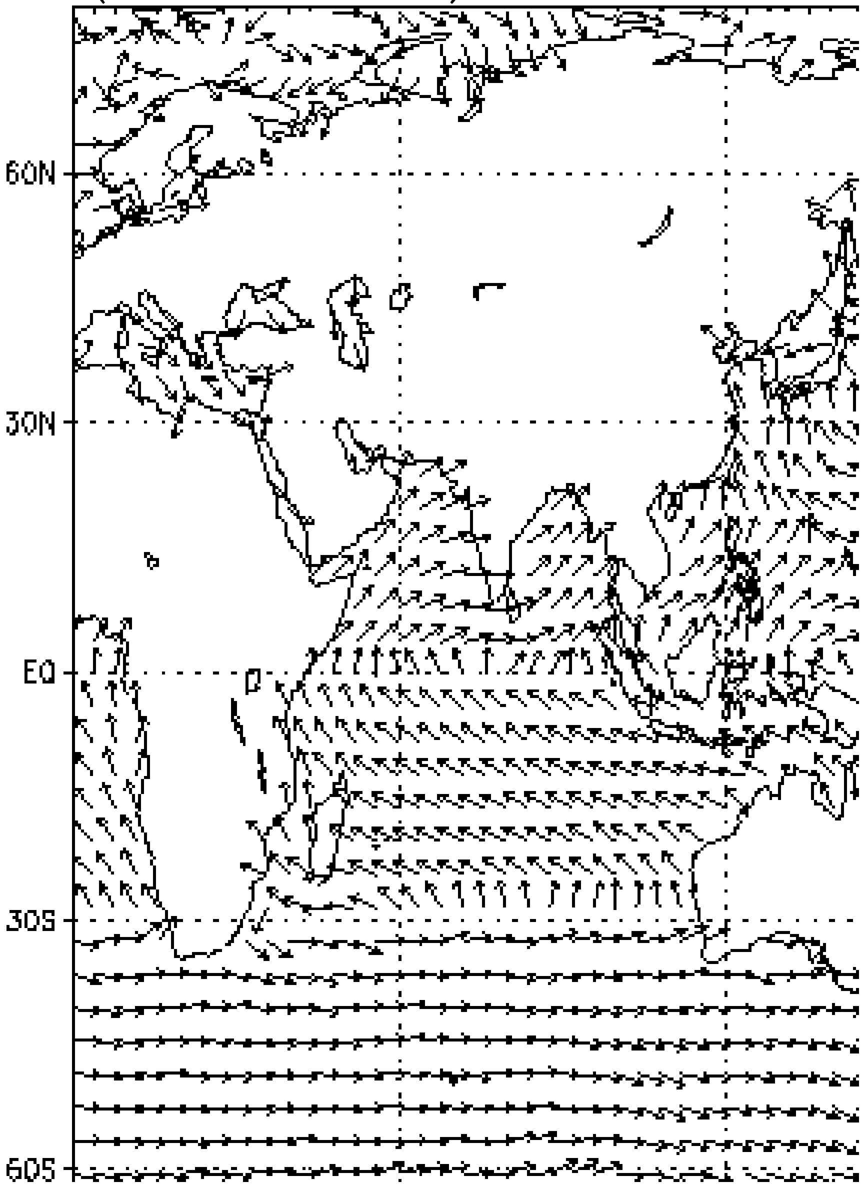
Associated with density changes due to heating/cooling, evaporation/precipitation, diffusion of heat and salt - total ocean (top to bottom), weak flow, but is the most common type of flow in deep ocean.
Space scale: mostly global.
Time scale: climate time scales.

Wind Driven Circulation

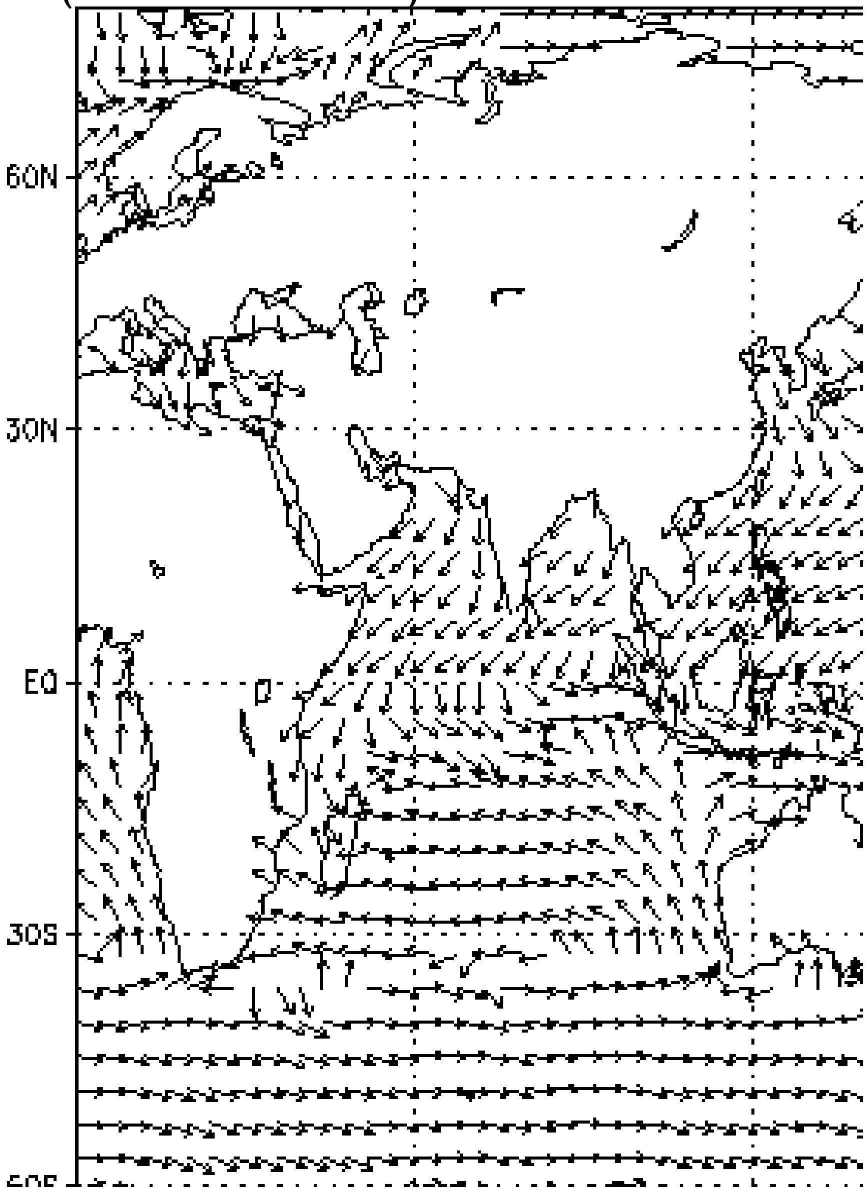


Wind Driven Circulation – Indian Ocean

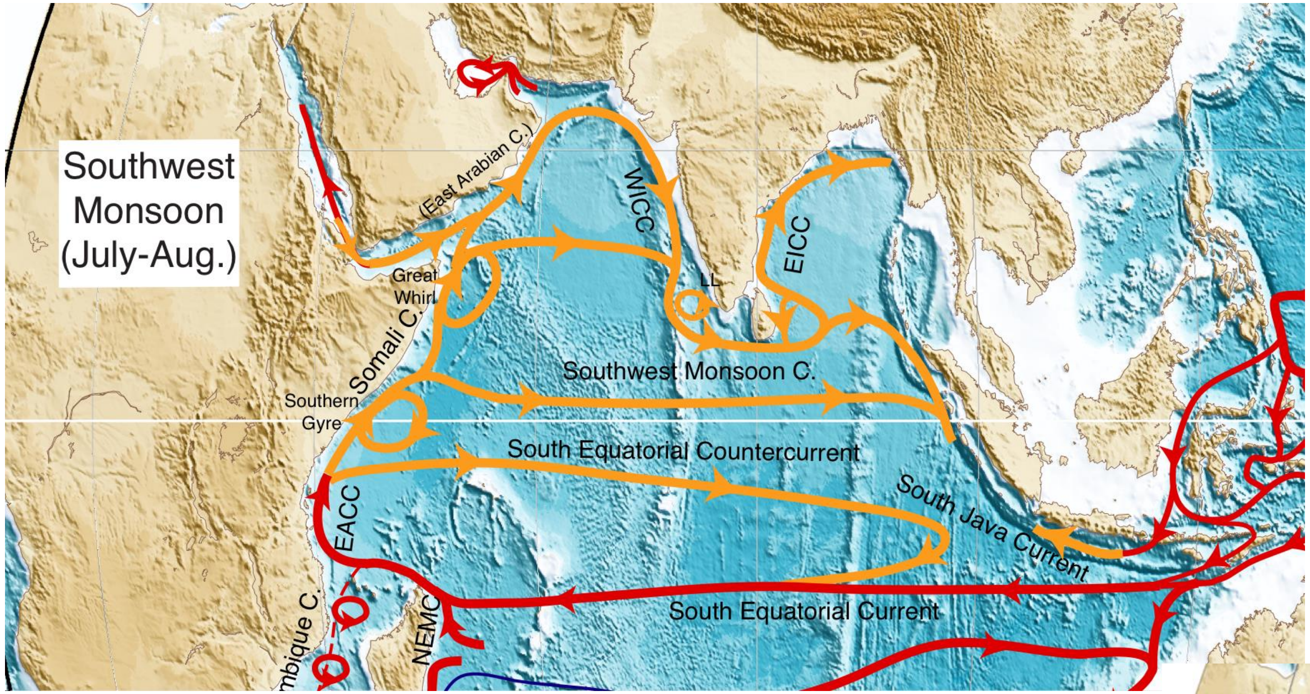
June-August winds
(SW monsoon)

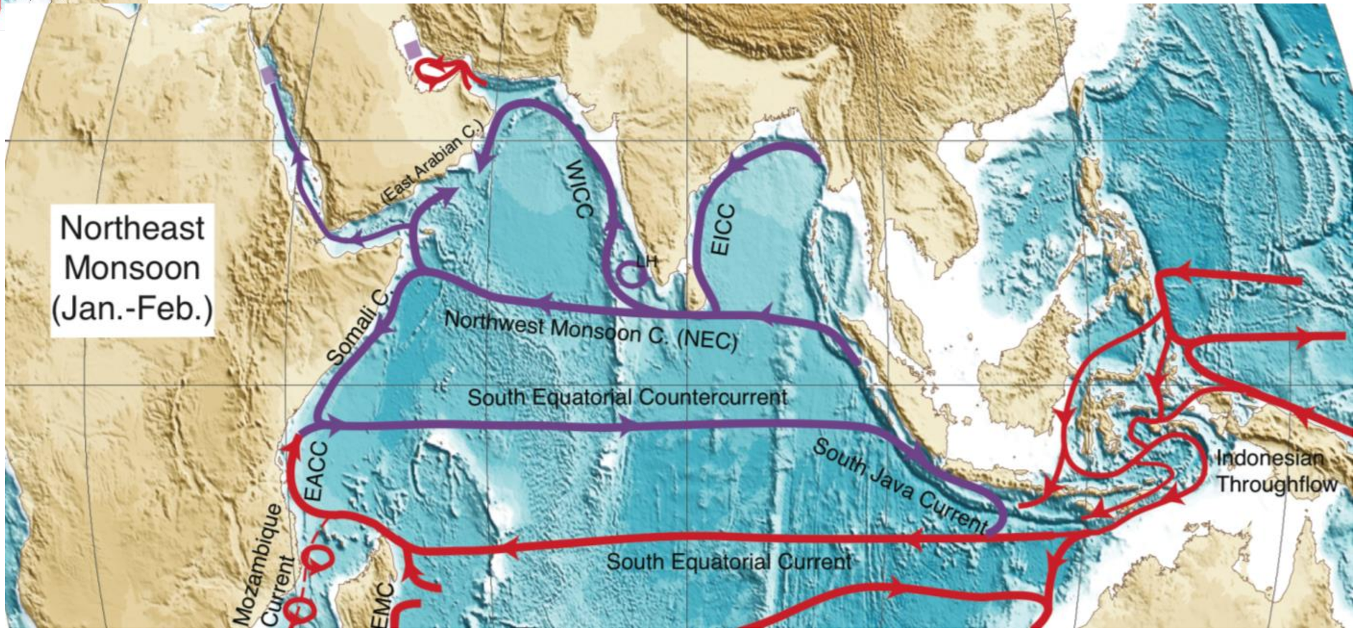


Dec-Feb winds
(NE monsoon)



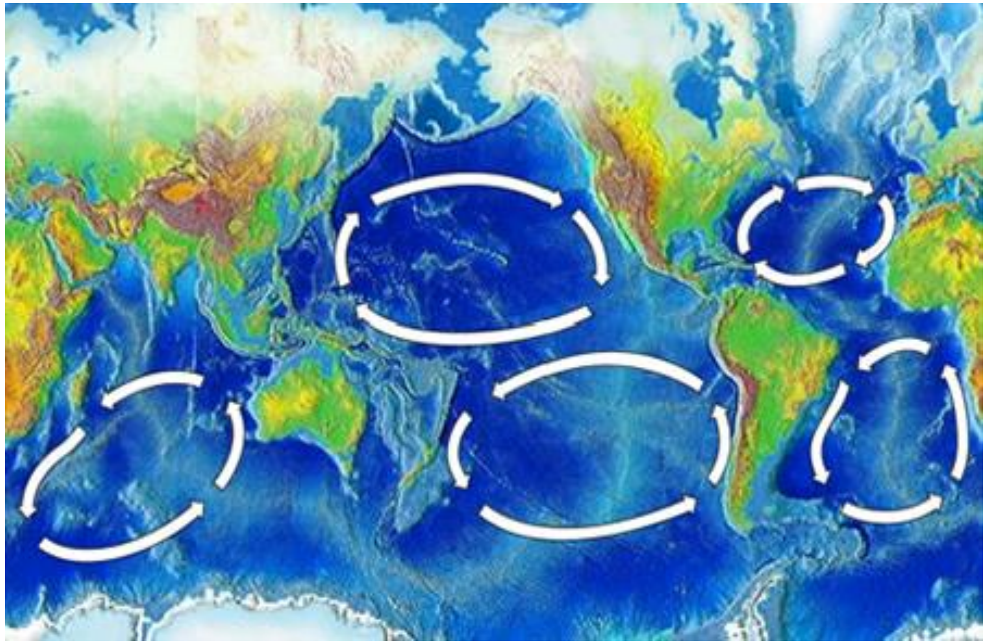
Wind Driven Circulation – Indian Ocean

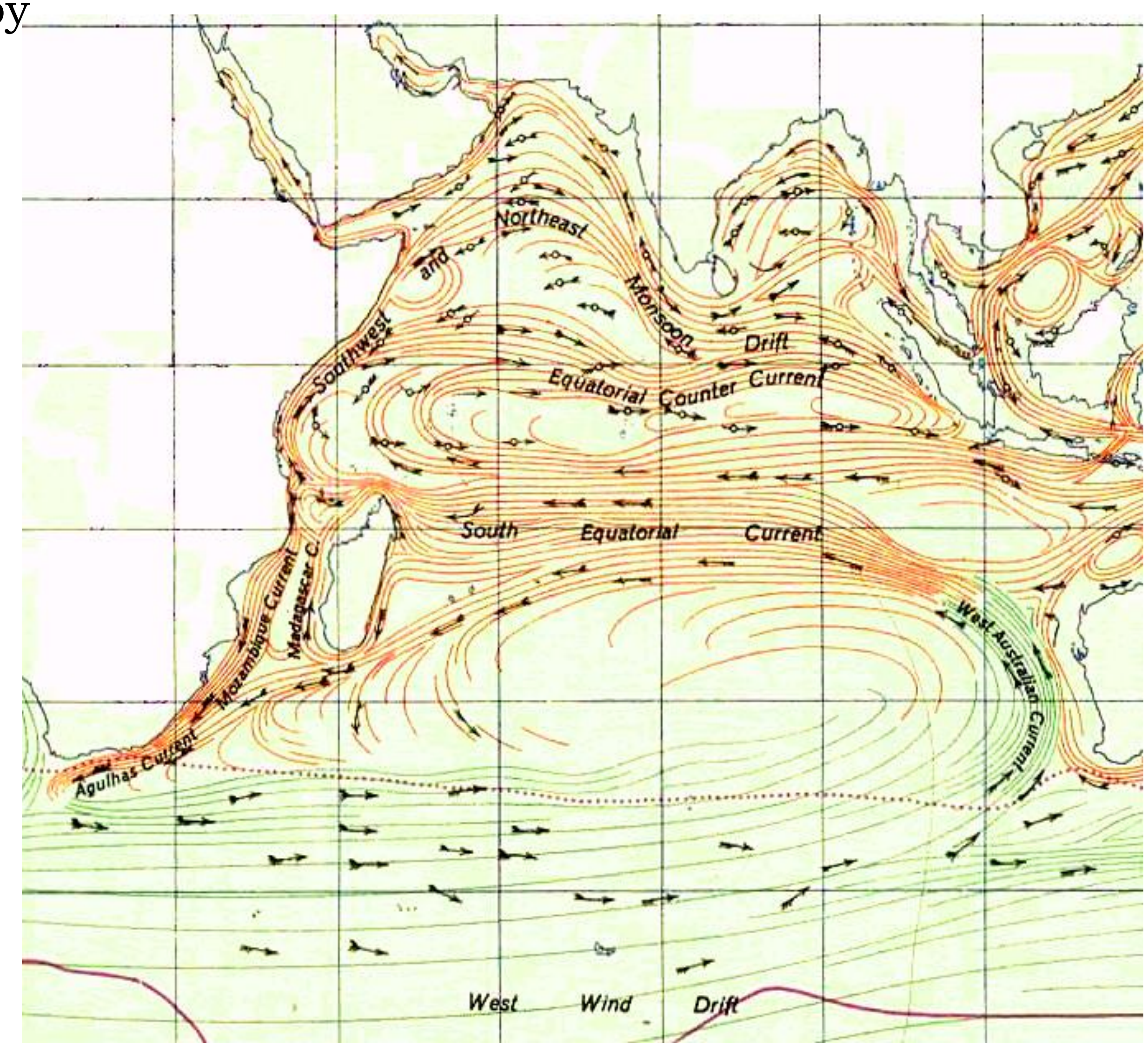




Ocean Gyres

Large system of circular ocean currents formed by global wind patterns and Earth's rotation.





Ocean Gyres as Garbage Patches

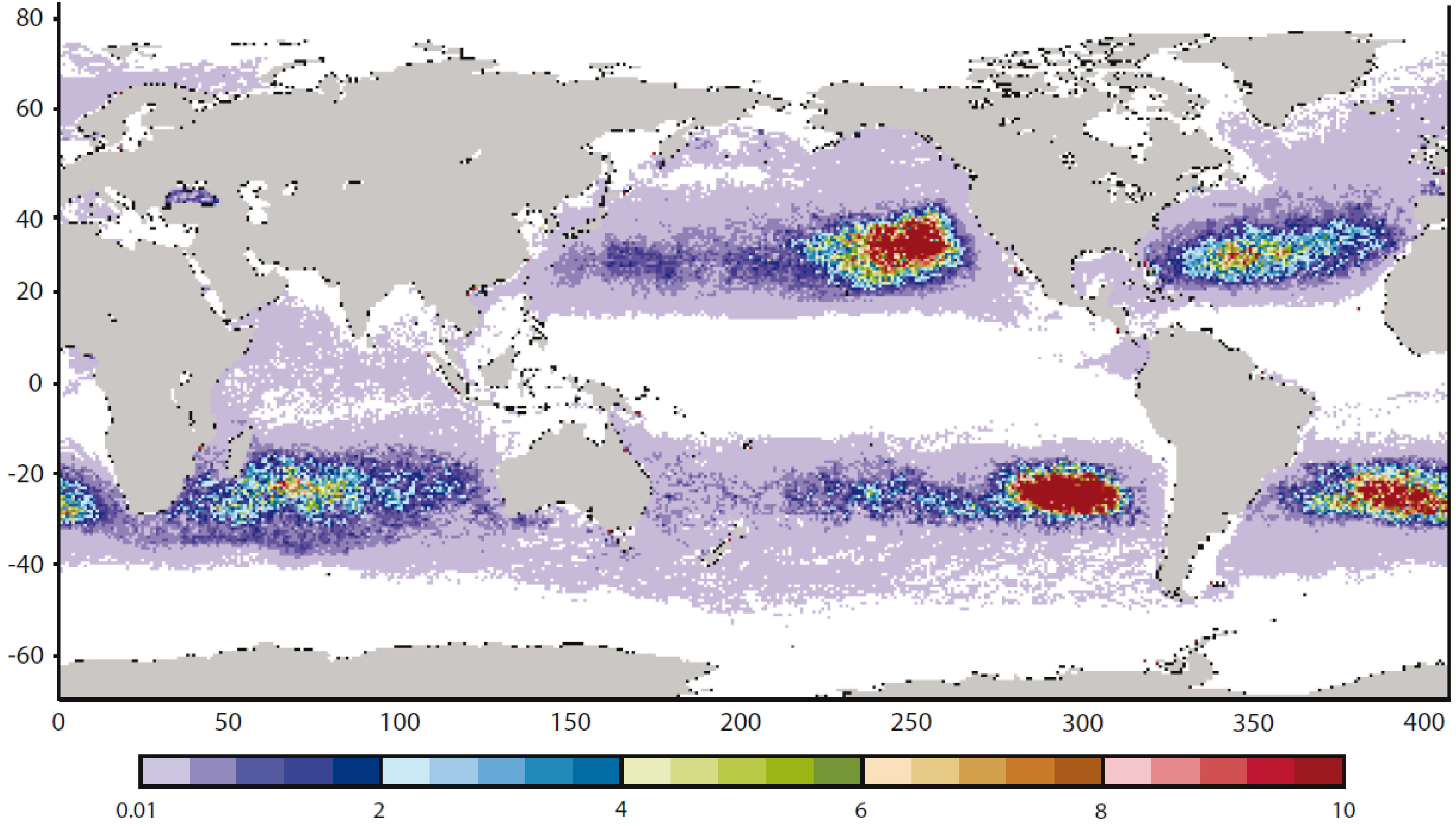
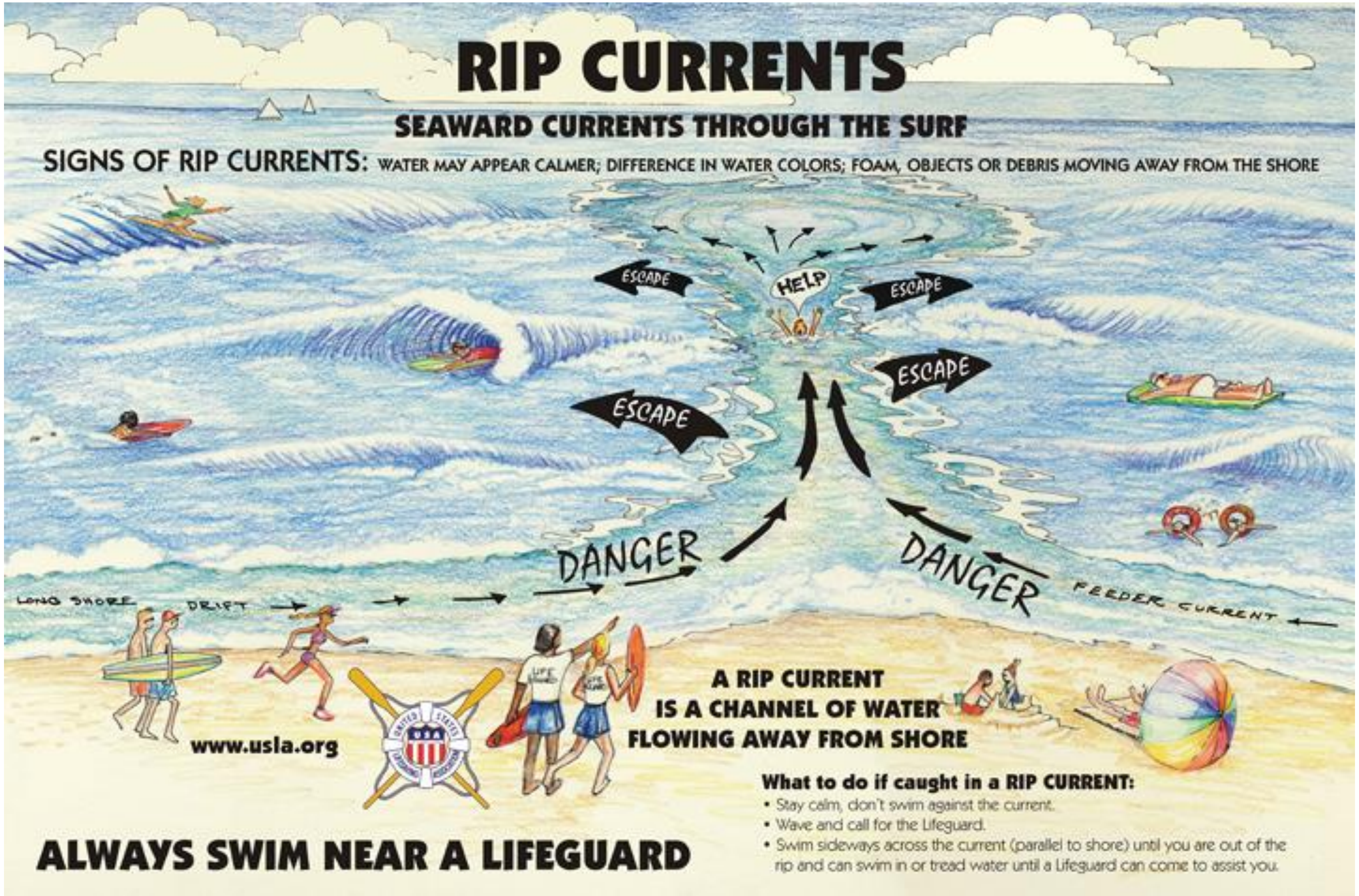


Figure 3: A model simulation of the distribution of marine litter in the ocean after ten years shows plastic converging in the five gyres: the Indian Ocean gyre, the North and South Pacific gyres, and the North and South Atlantic gyres. The simulation, derived from a uniform initial distribution and based on real drifter movements, shows the influence of the five main gyres over time. *Source: IPRC 2008*

Rip Current



RIP CURRENTS
SEAWARD CURRENTS THROUGH THE SURF
SIGNS OF RIP CURRENTS: WATER MAY APPEAR CALMER; DIFFERENCE IN WATER COLORS; FOAM, OBJECTS OR DEBRIS MOVING AWAY FROM THE SHORE
ESCAPE
HELP
ESCAPE
ESCAPE
ESCAPE
DANGER
DANGER
LONG SHORE
DRIFT
FEEDER CURRENT
A RIP CURRENT
IS A CHANNEL OF WATER
FLOWING AWAY FROM SHORE
www.usla.org
What to do if caught in a RIP CURRENT:
• Stay calm, don't swim against the current.
• Wave and call for the Lifeguard.
• Swim sideways across the current (parallel to shore) until you are out of the rip and can swim in or tread water until a Lifeguard can come to assist you.
ALWAYS SWIM NEAR A LIFEGUARD

Rip Current

Thermohaline Circulation - Simplified

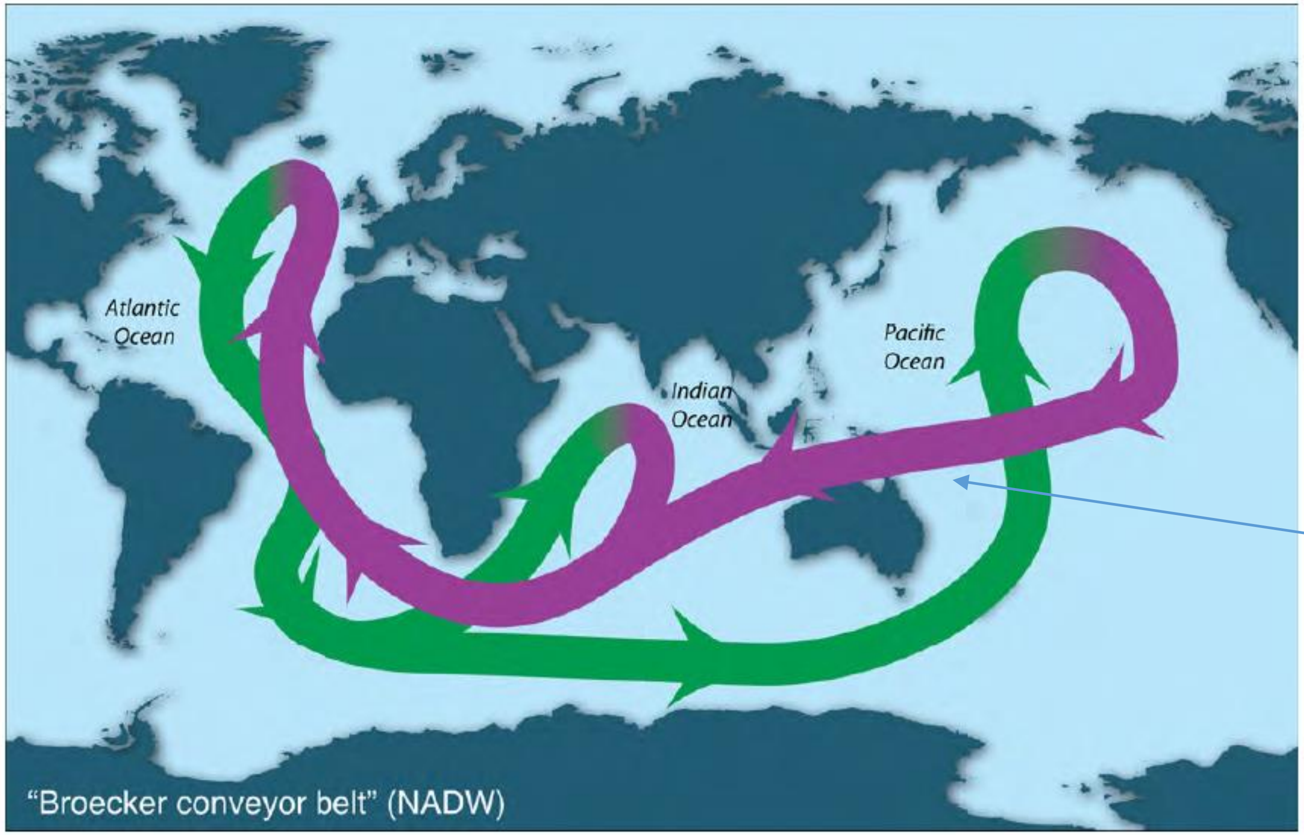


FIGURE 14.10 Simplified global NADW cell, which retains sinking only somewhere adjacent to the northern North Atlantic and upwelling only in the Indian and Pacific Oceans. See text for usefulness of, and also issues with, this popularization of the global circulation, which does not include any Southern Ocean processes. *Source: After Broecker (1987).*

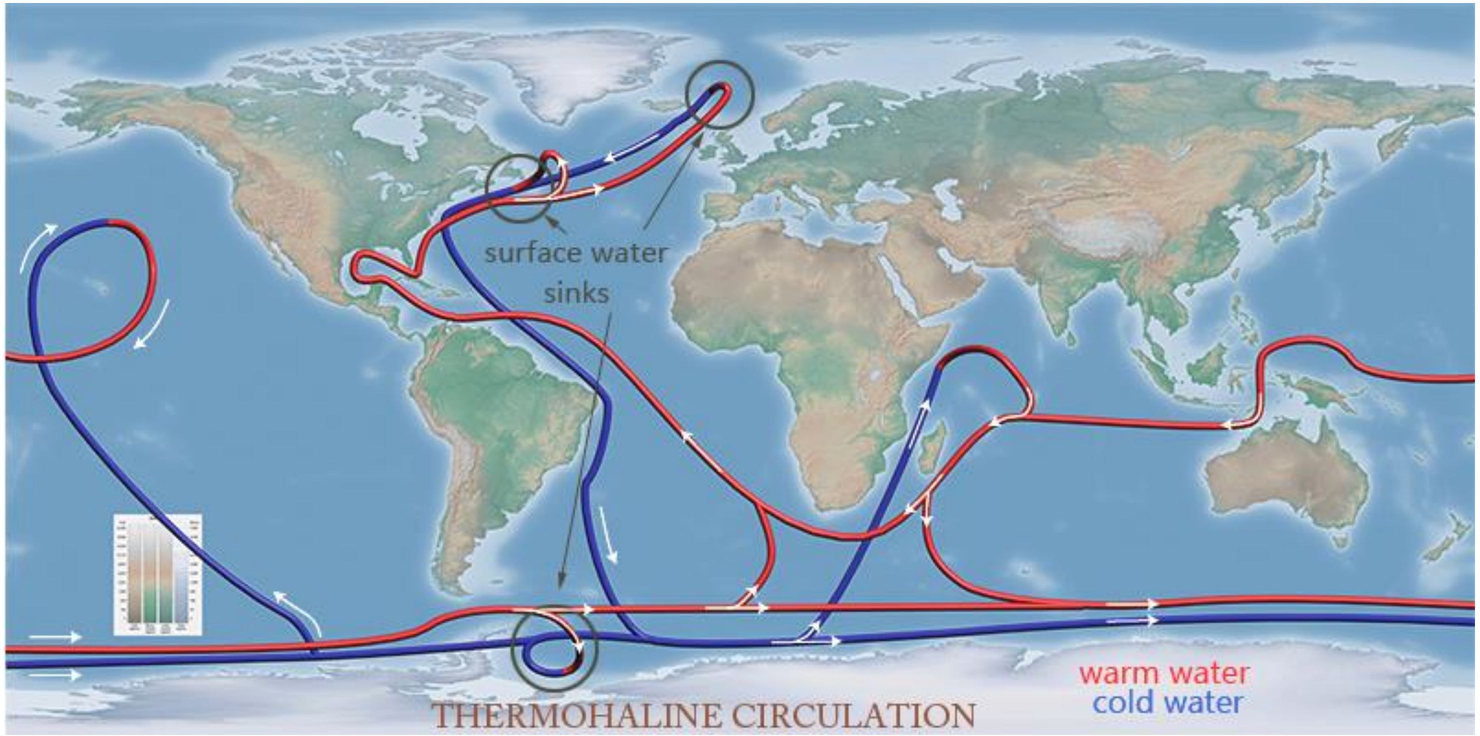
The characters in NEMO followed the EAC, part of the global ocean circulation which goes through east of Australia.

Thermohaline circulation (THC) refers to a part of the large-scale ocean circulation that is driven by global density gradients created by surface heat and freshwater fluxes.

Names: THC, Global Conveyer Belt, MOC

It is estimated that it can take ~**1,000 years** for a "parcel" of water to complete the journey along the global conveyor belt.

Thermohaline Circulation - Detailed



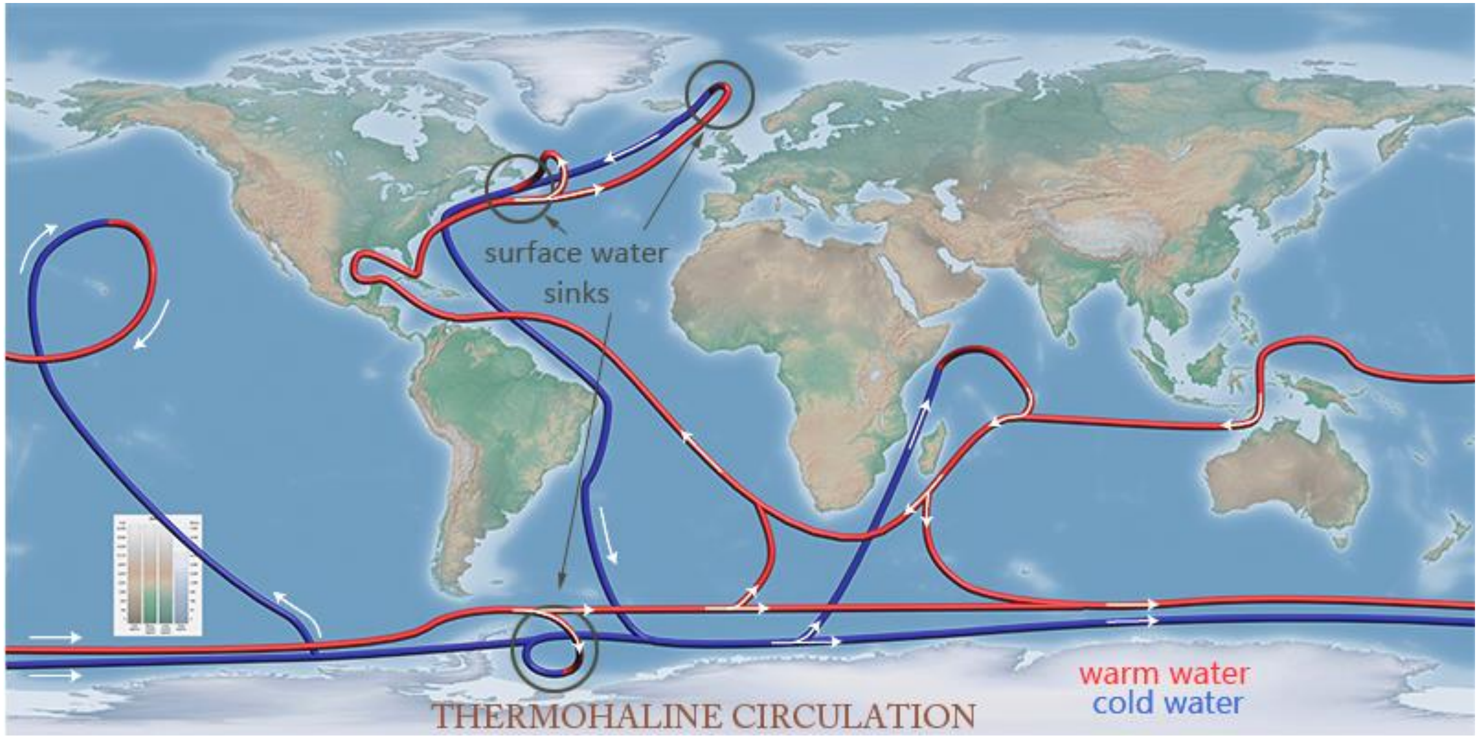
1. Begins on the surface of the ocean near the pole in the North Atlantic. Here, the water is chilled by arctic temperatures, and the winds. It also gets saltier because when sea ice forms, the salt does not freeze and is left behind in the surrounding water. The cold water is now more dense, due to the added salts, and sinks toward the ocean bottom. Surface water moves in to replace the sinking water, thus creating a current.

2. This deep water moves south, between the continents, past the equator, and down to the ends of Africa and South America. The current travels around the edge of Antarctica, where the water cools and sinks again, as it does in the North Atlantic. Thus, the conveyor belt gets "recharged."

3. As it moves around Antarctica, two sections split off the conveyor and turn northward. One section moves into the Indian Ocean, the other into the Pacific Ocean.

4. These two sections that split off warm up and become less dense as they travel northward toward the equator, so that they rise to the surface (upwelling). They then loop back southward and westward to the South Atlantic, eventually returning to the North Atlantic, where the cycle begins again.

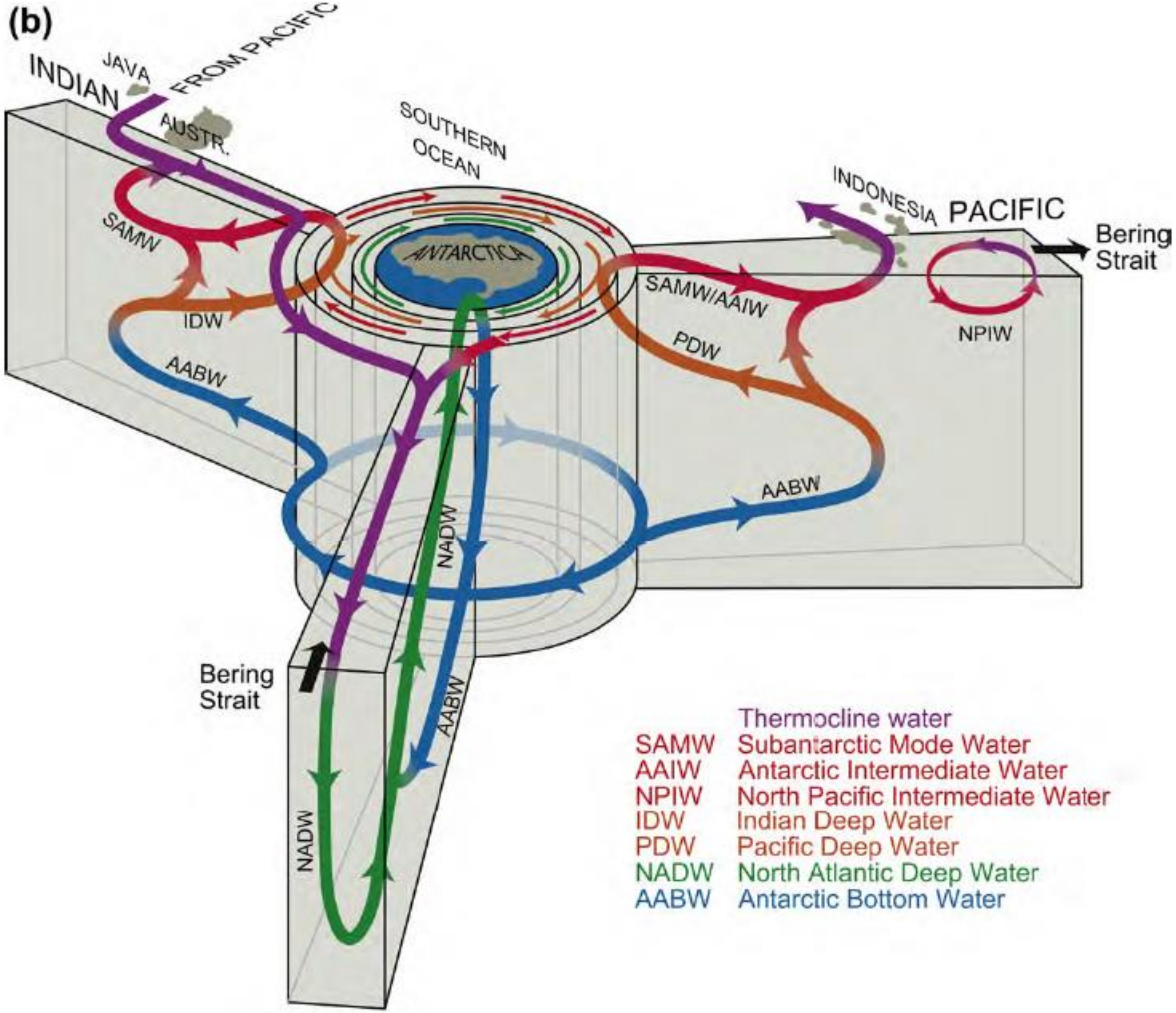
Thermohaline Circulation - Detailed



Cold water in polar zones sink relatively rapidly over a small area, while warm water in temperate and tropical zones rise more gradually across a much larger area. It then slowly returns poleward near the surface to repeat the cycle.

The continual diffuse upwelling of deep water maintains the existence of the permanent thermocline found everywhere at low and mid-latitudes. This slow upward movement is estimated to be about 1 cm/day over most of the ocean.

Thermohaline Circulation - Detailed



Water mass and water types

Two basic circulation systems in the oceans

- the wind-driven surface circulation
- the deepwater density-driven circulation

Only about 10% of the ocean volume is involved in wind-driven surface currents. The other 90% circulates due to density differences in water masses

Water masses are identified by their temperature, salinity, and other properties such as nutrients or oxygen content.

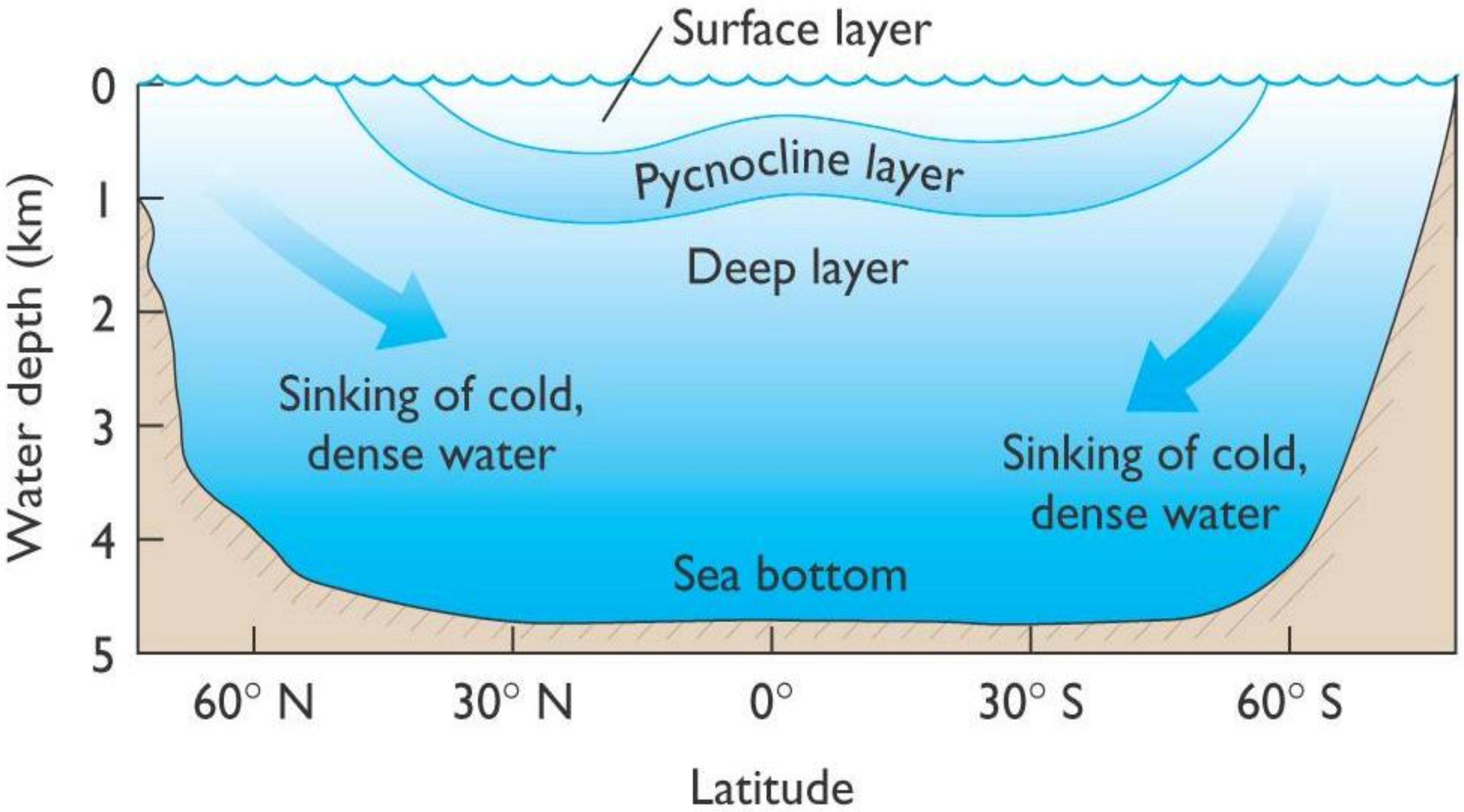
All water masses gain their particular characteristics because of interaction with the surface during their development.
Water properties, such as T and S, are formed only when the water is at the surface or in the mixed layer.

- Different inputs of freshwater
- Patterns of precipitation
- Evaporation
- temperature regimes
- brine rejection

Once water masses sink, their temperature and salinity are modified only by mixing with other water masses (diffusive and turbulent heat exchange).

process is very slow
Thus water from a particular region has a particular temperature associated with a particular salinity, and the relationship changes little as the water moves through the deep ocean.

Deep and bottom waters



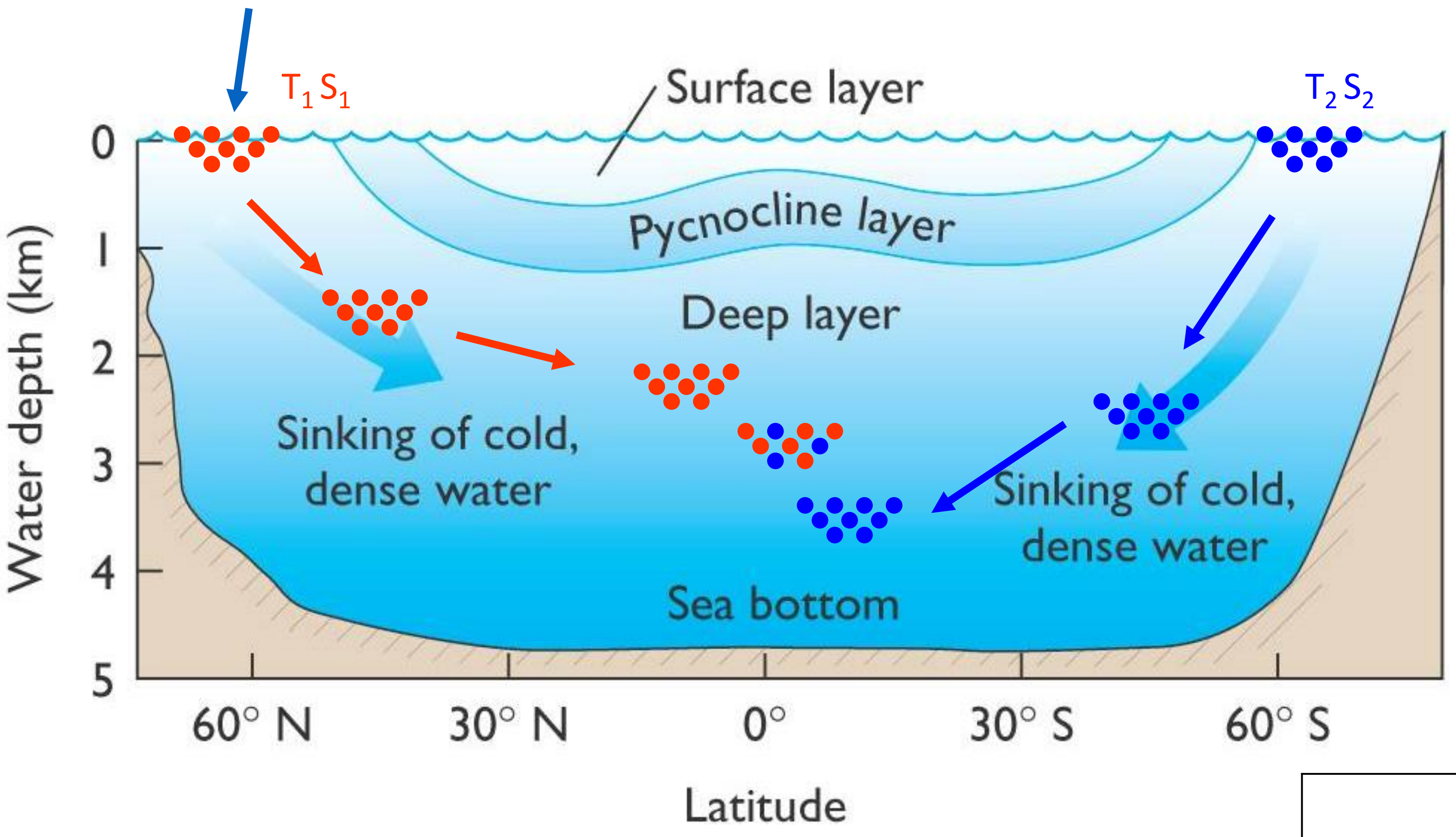
Tropical oceans: pycnocline ≈ thermocline

Mid-latitudes: pycnocline ≈ halocline

High latitudes: no pycnocline formation

Deep and bottom waters

Diabatic exchanges with the atmosphere at the surface



Adiabatic changes and *Mixing* in ocean interior

Season	Mean temperature	
	North Pole	**South Pole**
Summer	0° C	−28.2° C
Winter	−40° C	−60° C

Why is the south pole colder than the north?

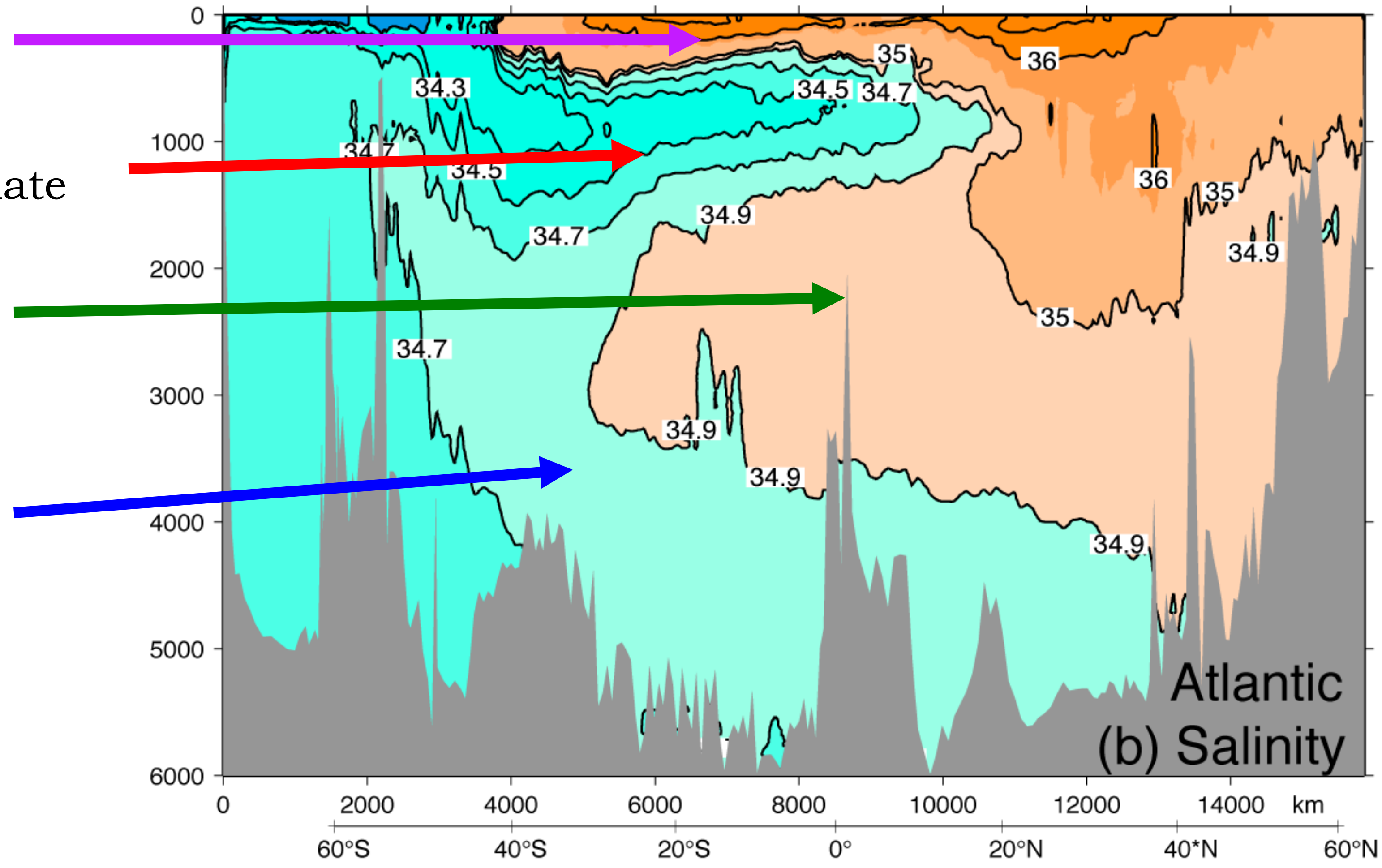
Water masses

Upper

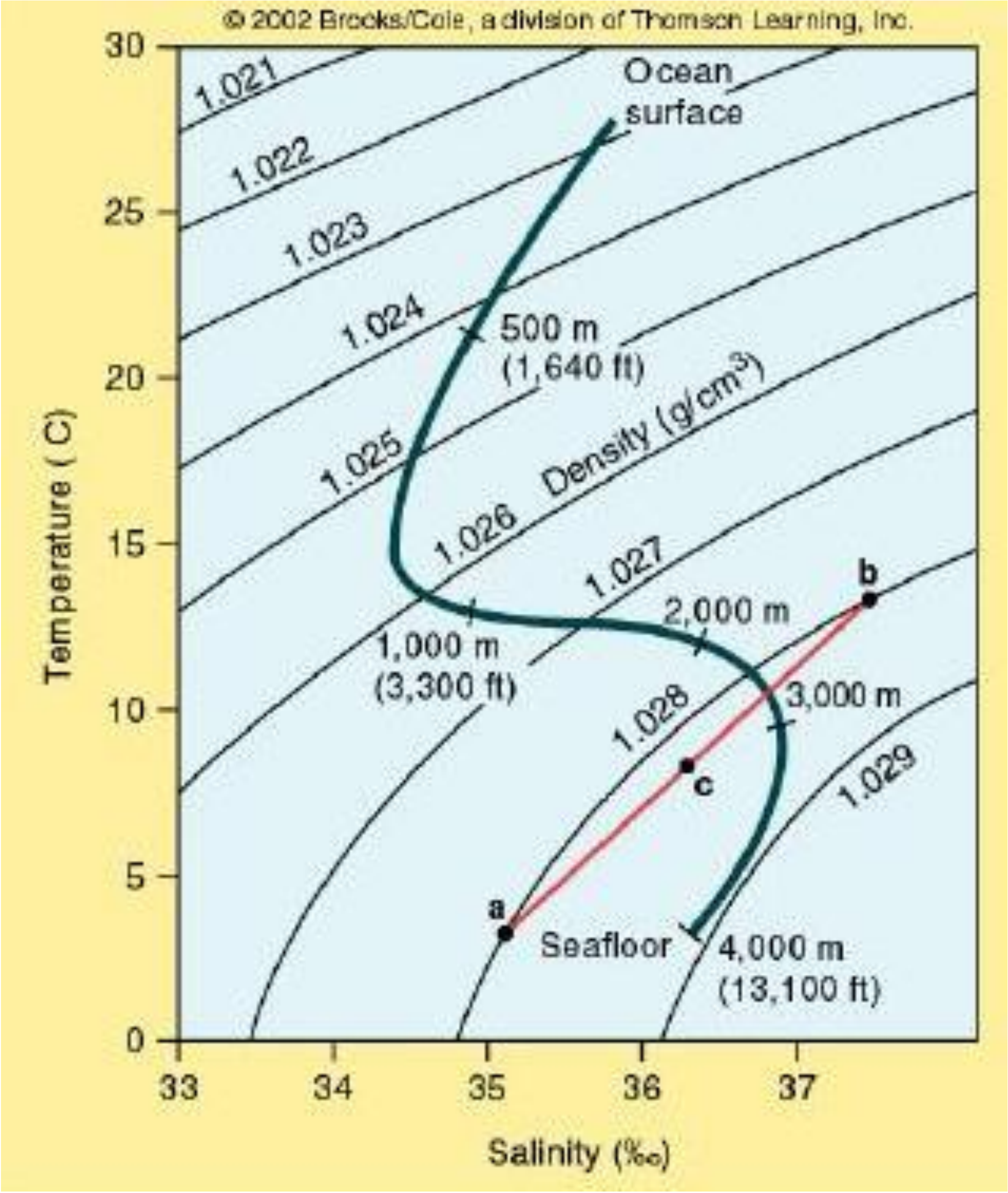
Intermediate

Deep

Bottom



T-S Diagram – relative conservative



Surface WM

Central WM

Intermediate WM

Deep WM

Bottom WM

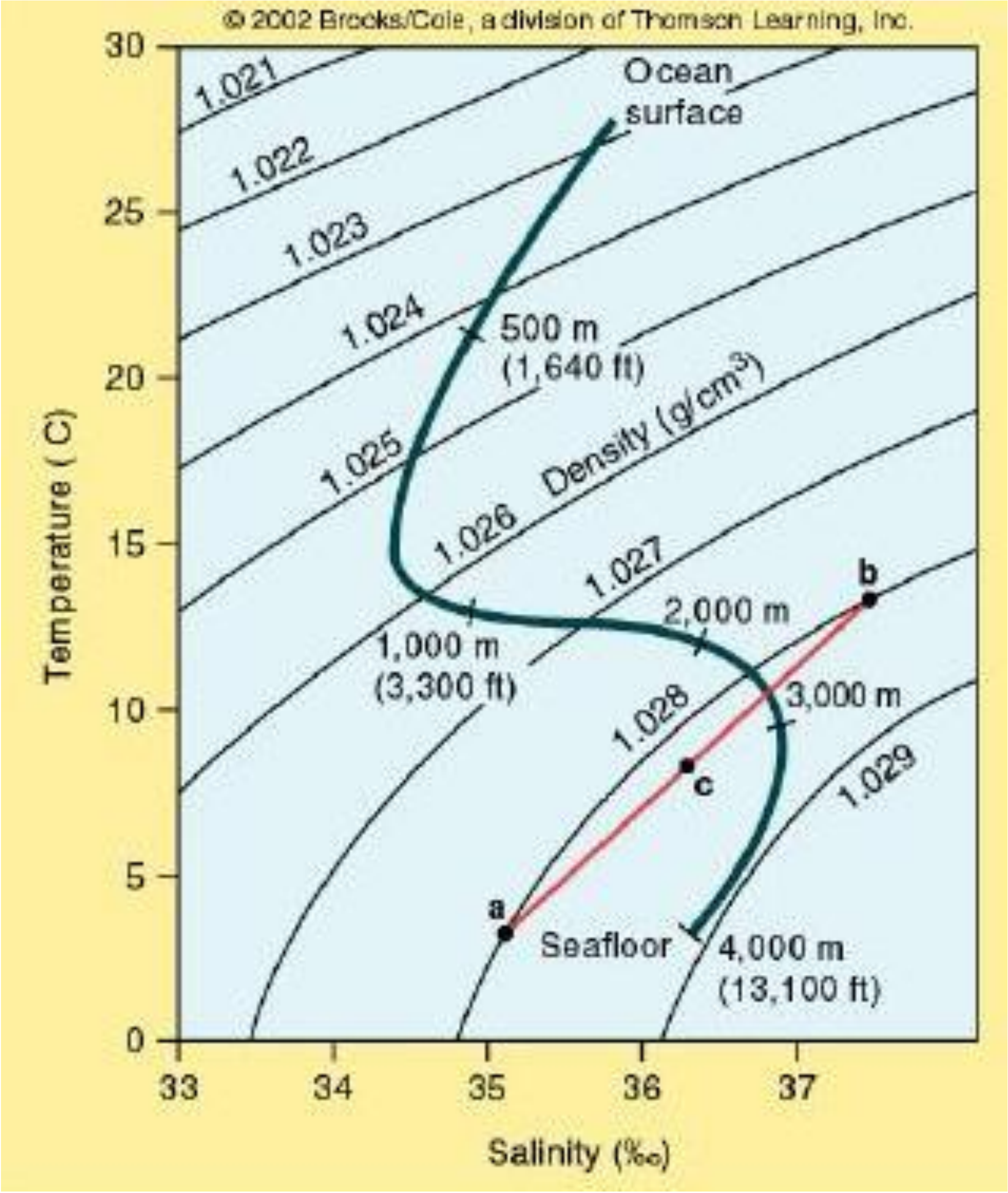
Plots of salinity as a function of temp., called **T-S plots**, are used to delineate water masses and their geographical distribution.

To easily determine the relationship between temperature and salinity for a water mass, scientists use a tool called the temperature-salinity diagram (TS diagram). It works by plotting salinity on the x-axis and temperature on the y-axis.

Temperature and salinity combine to form the water's density, which is represented in the diagram by lines of equal density. The **isopycnals**, as lines of equal density are called, are determined by the interaction of temperature and salinity.

Water masses can be distinguished by their characteristic temperature-salinity profiles, which are determined by their formation conditions at the surface.

T-S Diagram – relative conservative



Surface WM

Central WM

Intermediate WM

Deep WM

Bottom WM

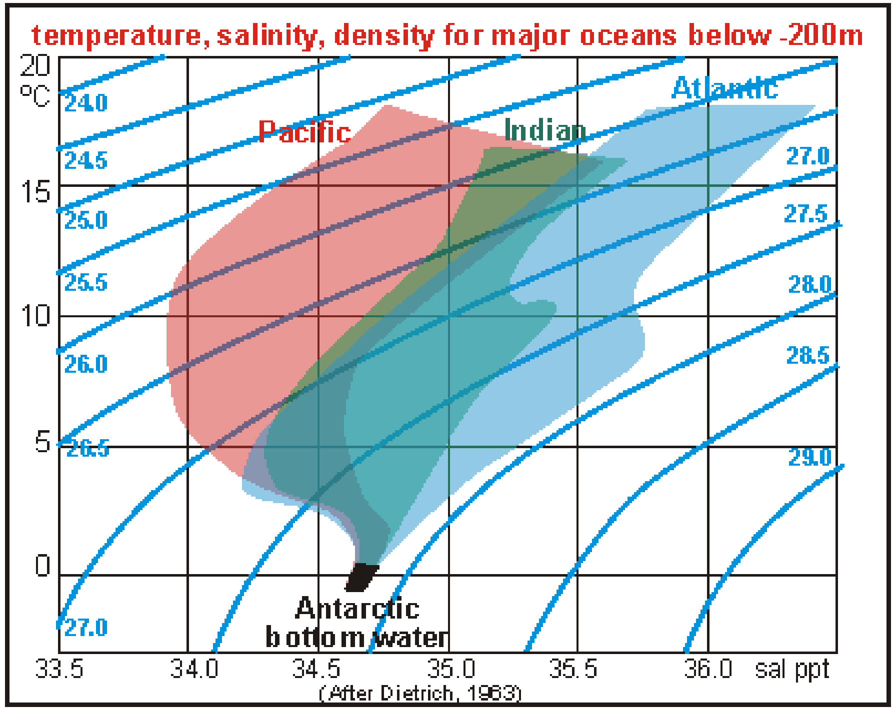
Plots of salinity as a function of temp., called **T-S plots**, are used to delineate water masses and their geographical distribution.

- Single point: A water type
- A smooth curve: a water mass

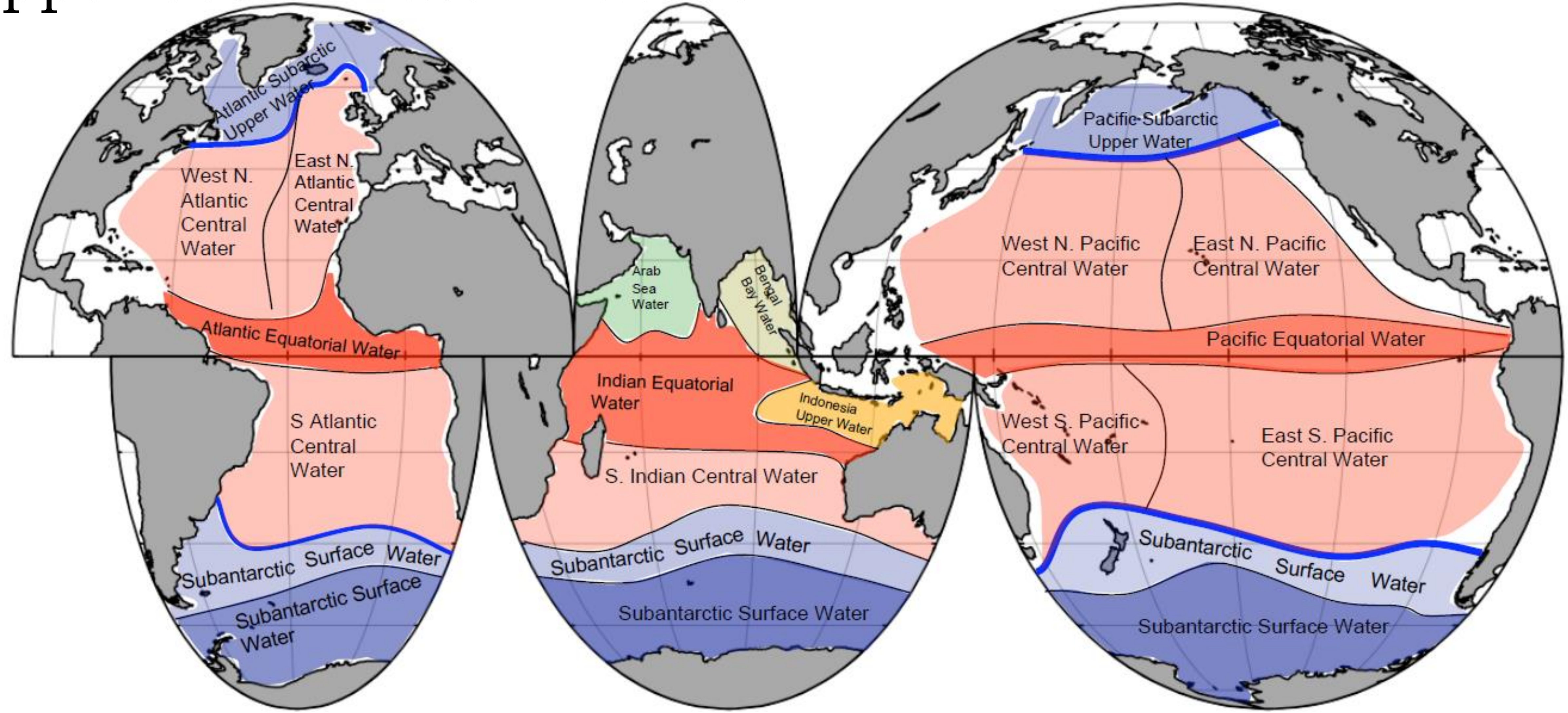
Application:

(a) identify the water masses and its sources

(b) water mixing



Upper ocean water masses



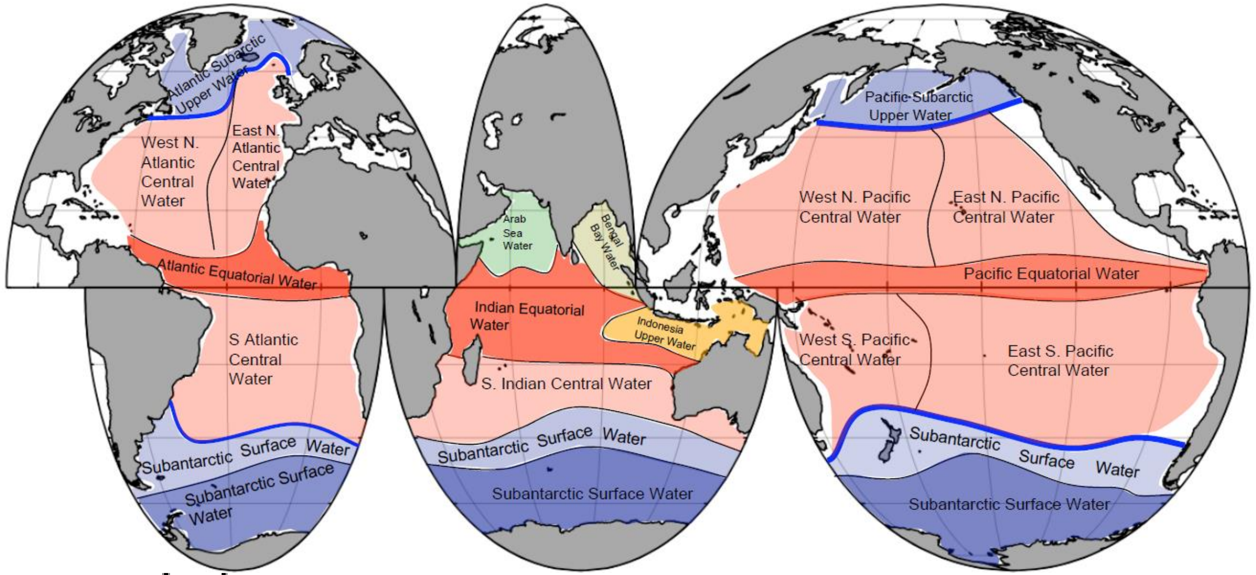
Characterization: Surface mixed layer down through the main pycnocline.
Location: In the tropics and subtropics and into the subpolar regions
Formation mechanisms: Mixed layer properties are set by air-sea fluxes, and depth by wind stirring or buoyancy-driven convection. Late winter mixed layer properties are "subducted" into the ocean interior.

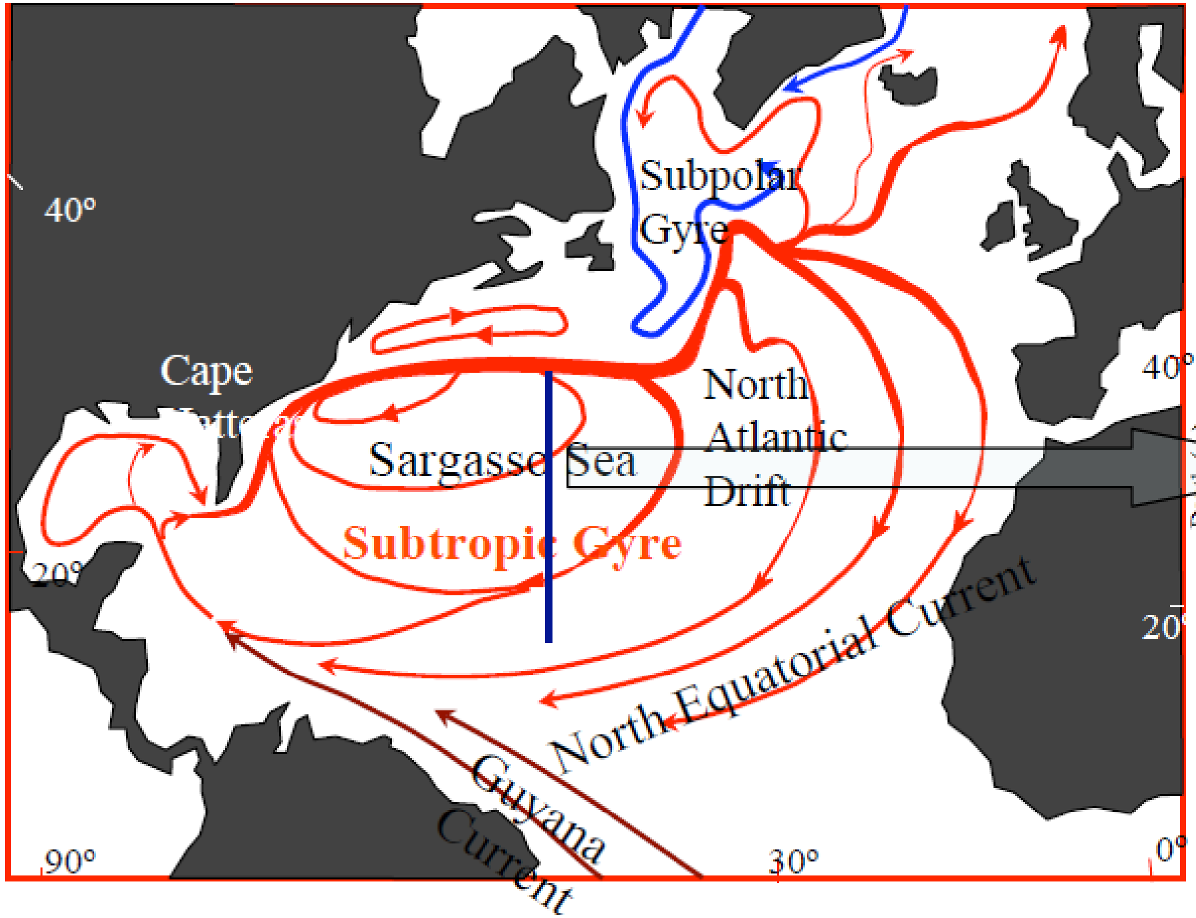
Upper ocean water masses

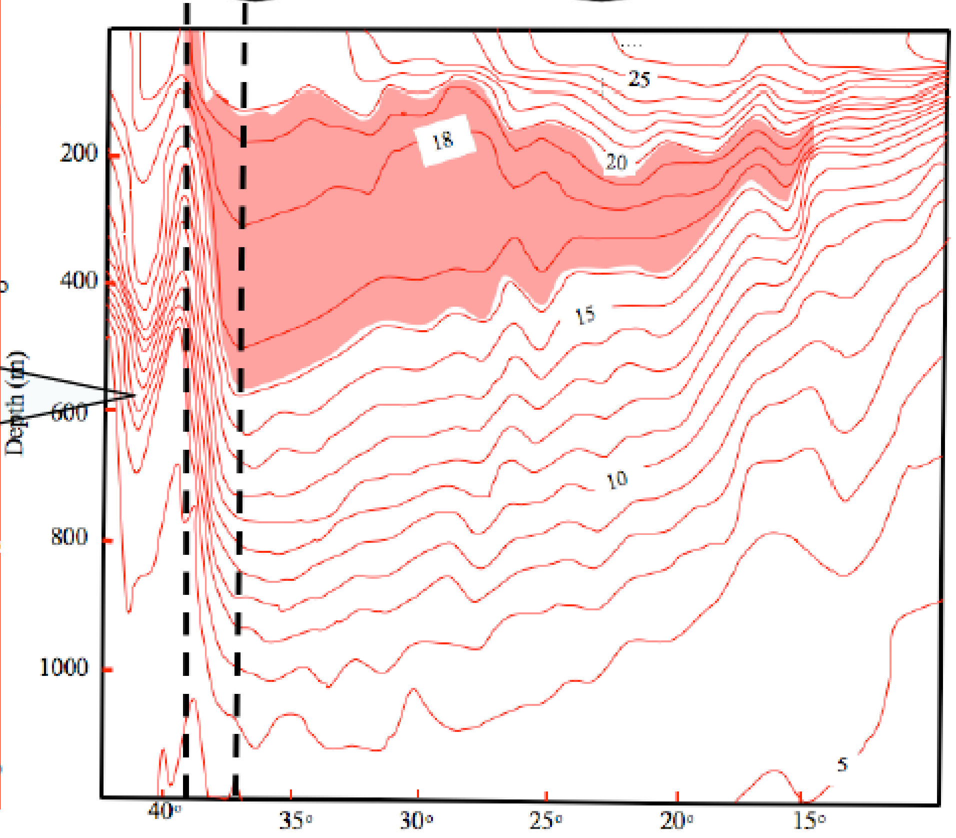
Central/Mode water is characterized by the minimum vertical density gradient.

Mode waters are characterized by homogeneity of water properties in the vertical as well as the horizontal.

~18°C

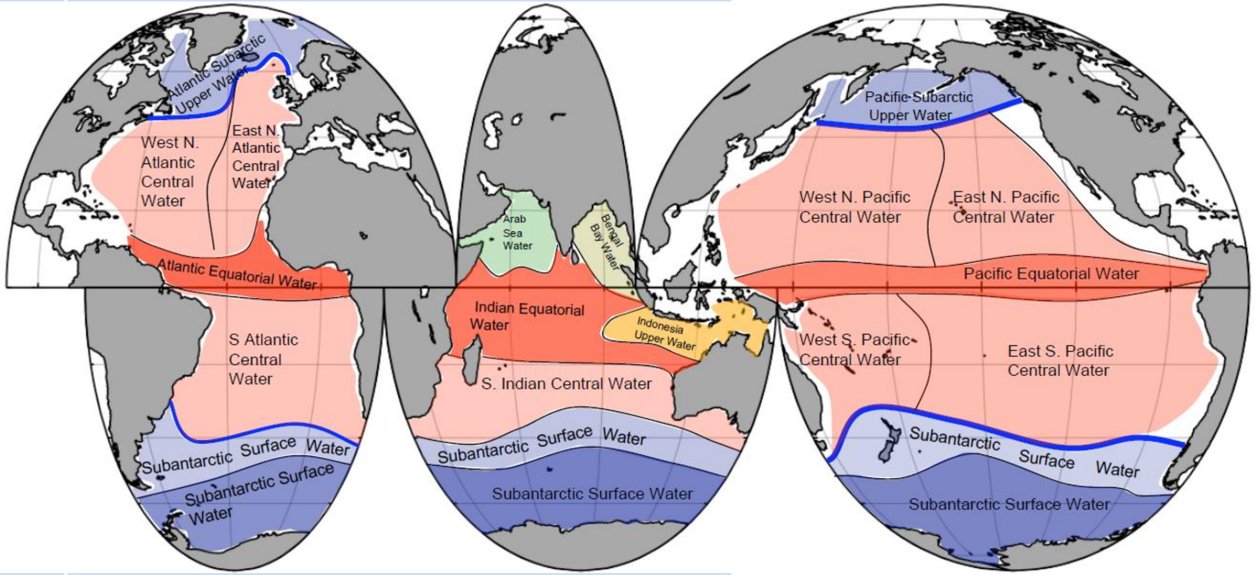




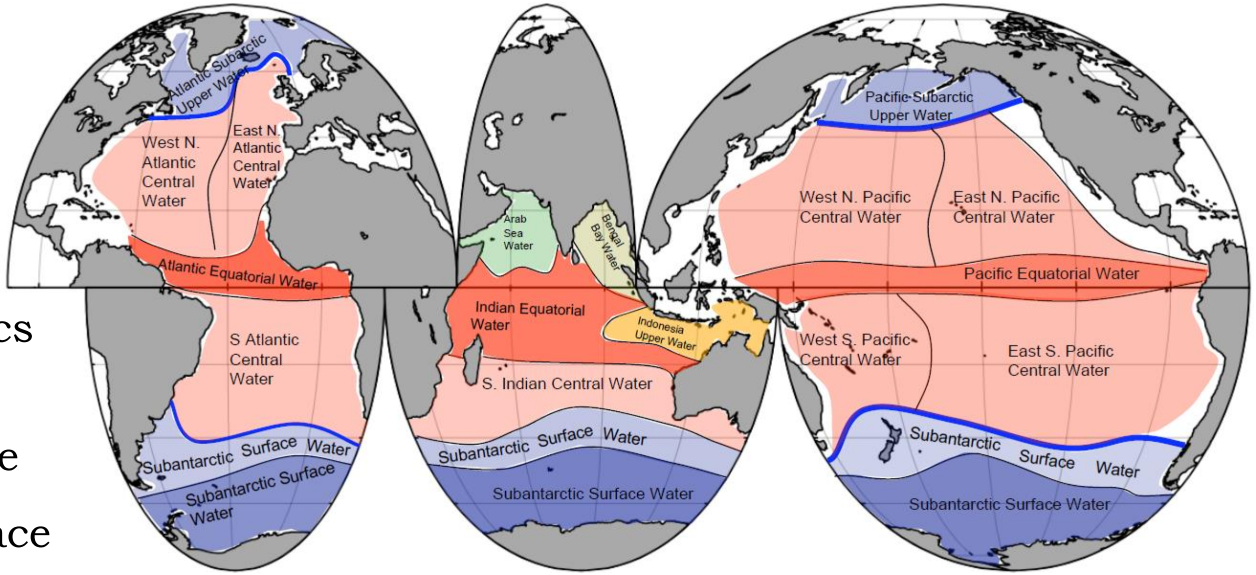


Upper ocean water masses

Atlantic Ocean	**Pacific Ocean**	**Indian Ocean**
West North Atlantic Central Water (**WNACW**) **T: 7.0-20.0°C** **S: 35.0-36.7 PSU**	West North Pacific Central Water (**WNPCW**) **T: 10.0-22.0°C** **S: 34.2-35.2 PSU**	South Indian Central Water (SICW) **T: 8.0-25.0°C** **S: 34.6-35.8 PSU**
East North Atlantic Central Water (**ENACW**) **T: 8.0-18.0°C** **S: 35.2-36.7 PSU**	East North Pacific Central Water (**ENPCW**) **T: 12.0-20.0°C** **S: 34.2-35.2 PSU**	
South Atlantic Central Water (**SACW**): **T: 5.0-18.0°C** **S: 34.3-35.8 PSU**	West South Pacific Central Water (**WSPCW**) **T: 6.0-22.0°C** **S: 34.5-35.8 PSU**	
	East South Pacific Central Water (**ESPCW**) **T: 8.0-24.0°C** **S: 34.4-36.4 PSU**	



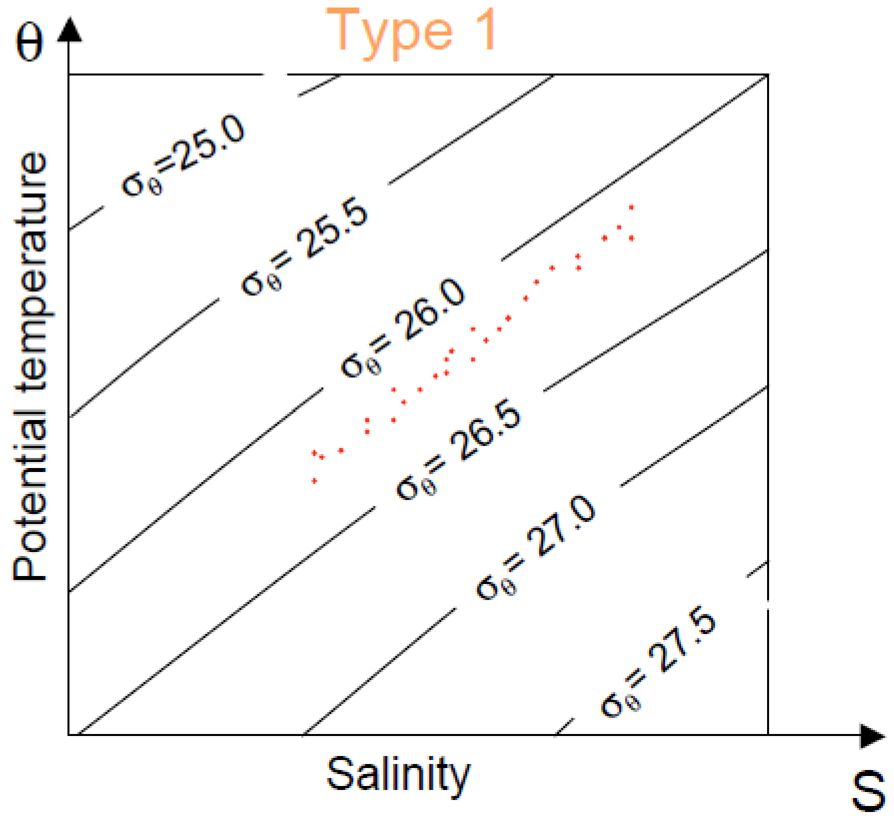
Upper ocean water masses

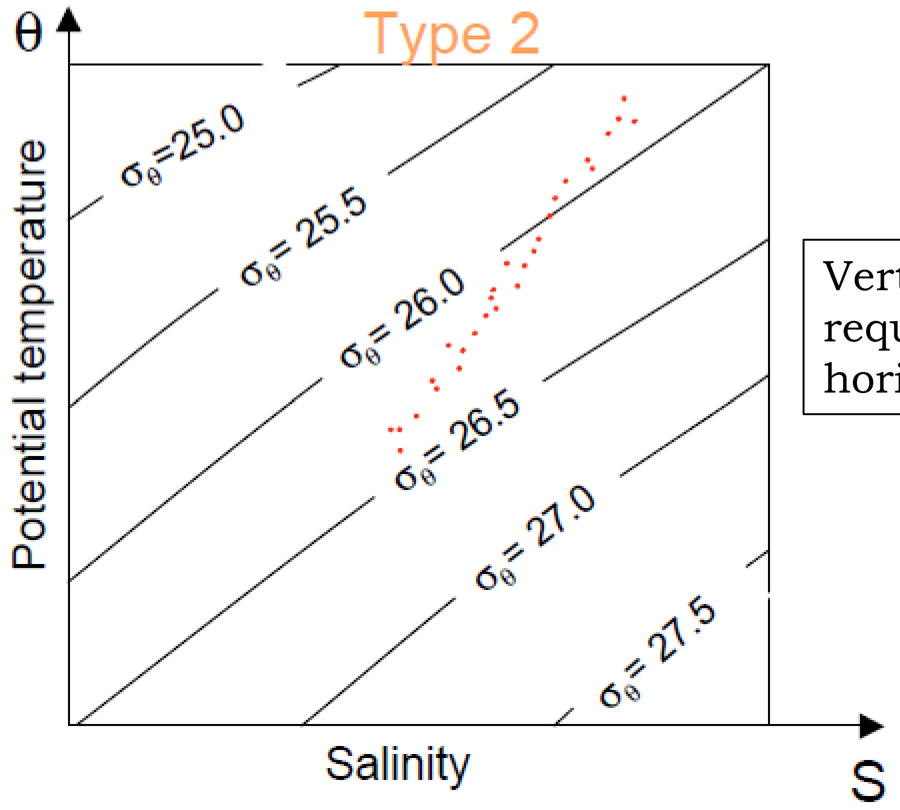


A T-S diagram also can help us determine the characteristics of water mixing in the water column

Type 1: Isopycnal mixing ⇒ mixing along the density surface

Type 2: Diapycnal mixing ⇒ mixing across the density surface

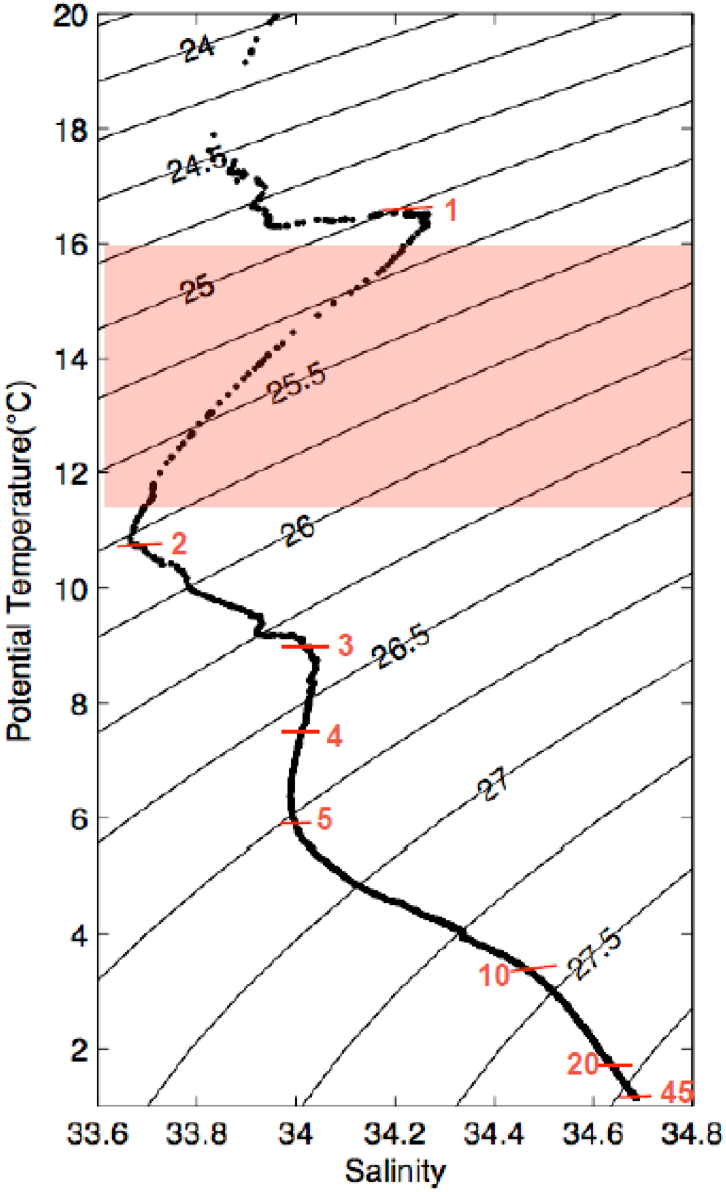




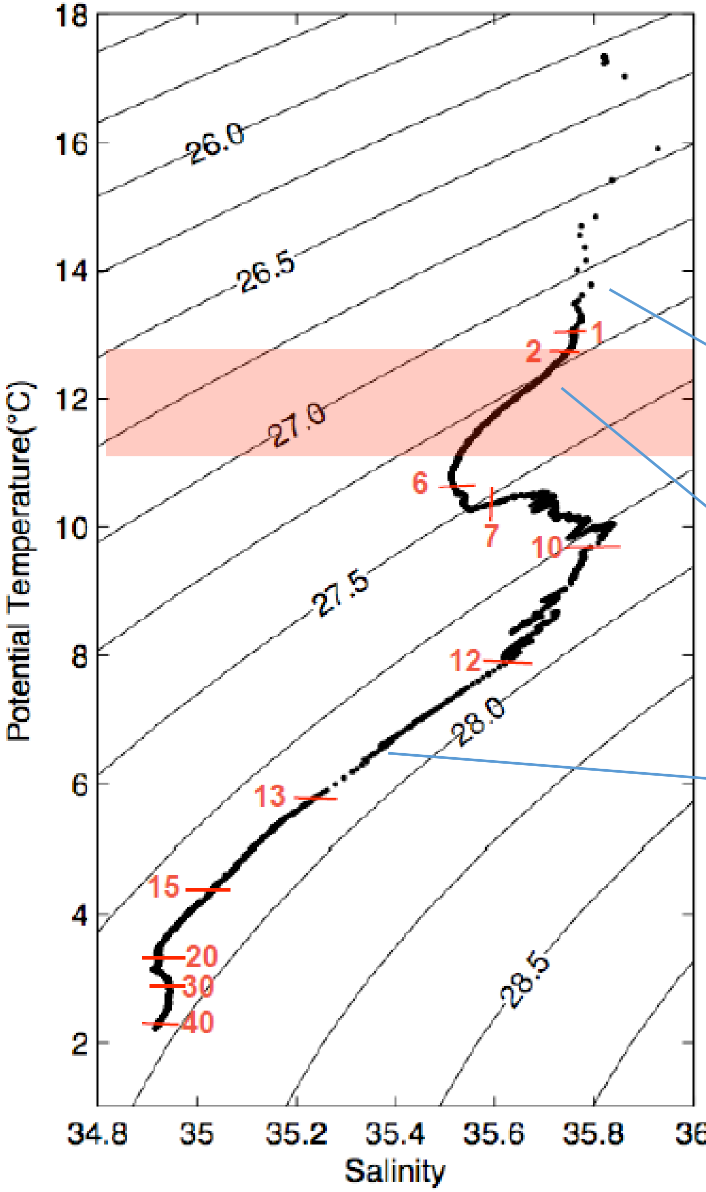
Vertical (diapycnal) mixing requires more energy than horizontal (isopycnal) mixing.

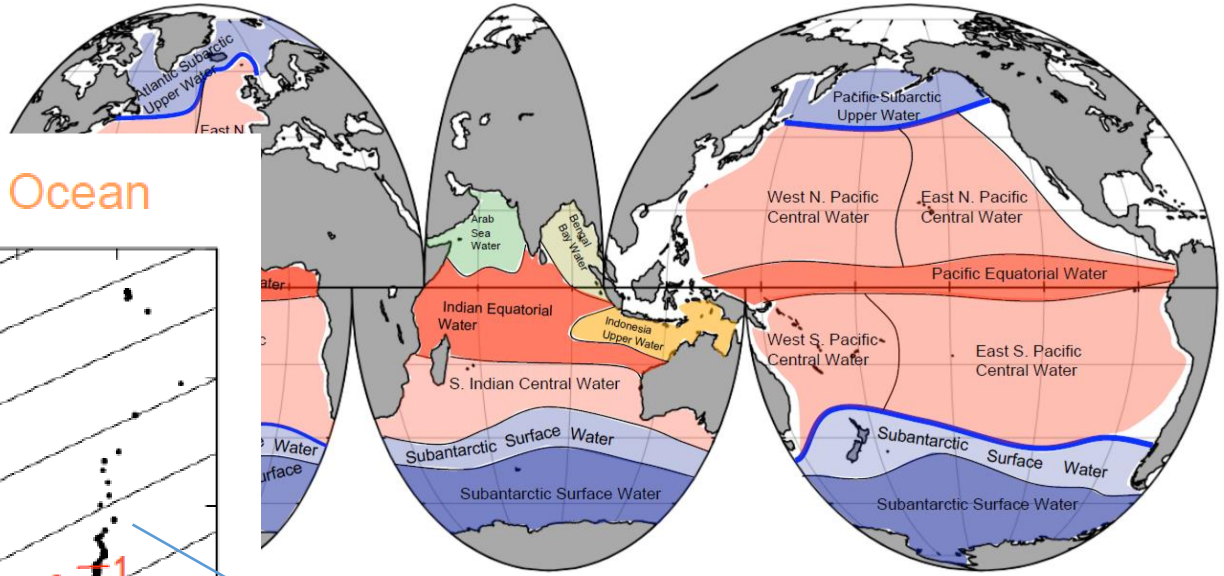
Upper ocean water masses

East Pacific Ocean



East Atlantic Ocean





Surface Waters

Central Waters show the **diapycnal** (and isopycnal) **mixing**, so it does not form only by the along-density surface subduction

Isopycnal mixing

Intermediate waters

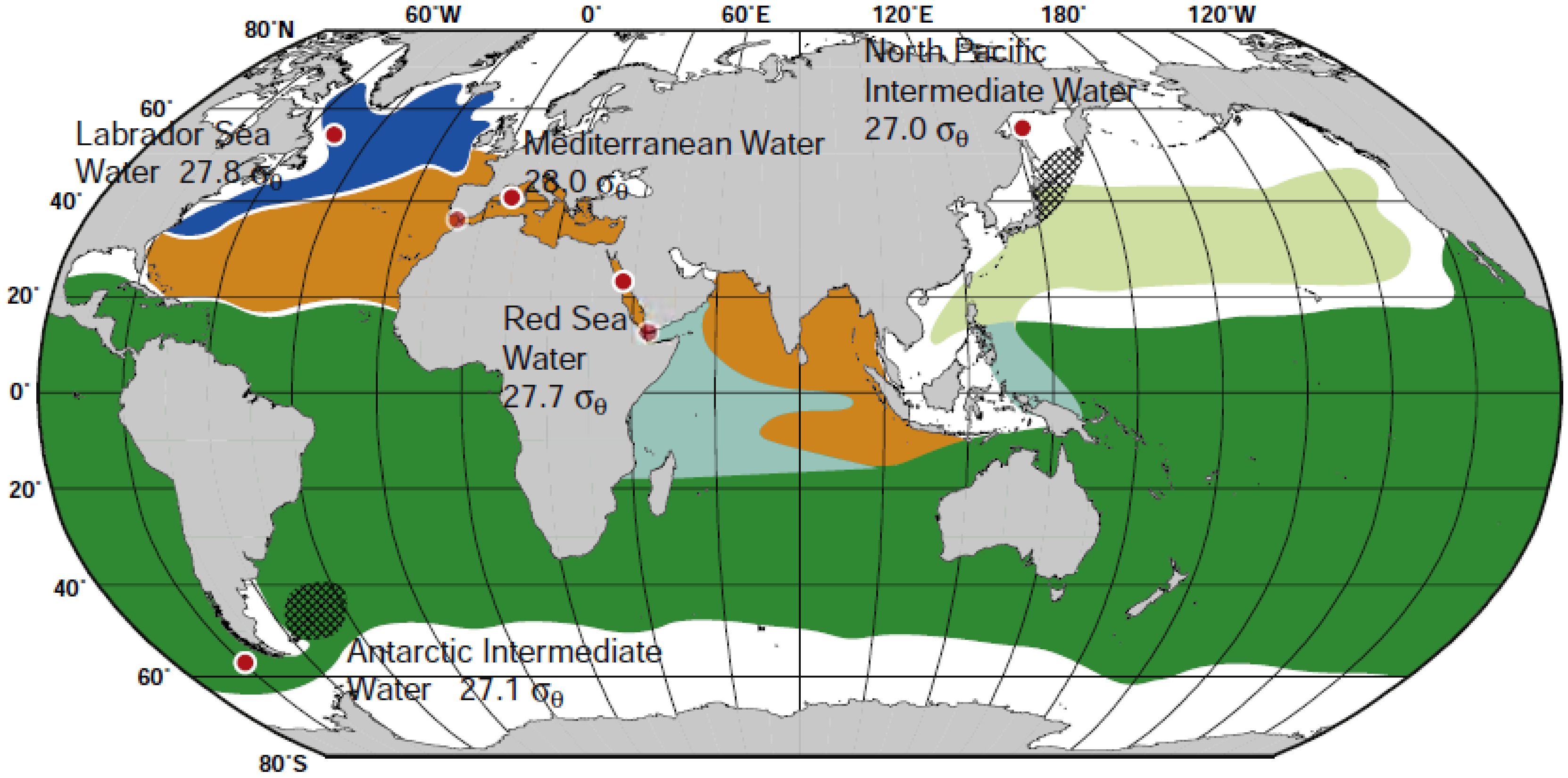


FIGURE 14.13 Low- and high-salinity intermediate waters. AAIW (dark green), NPIW (light green), LSW (dark blue), MW (orange in Atlantic), RSW (orange in Indian). Light blue in Pacific: overlap of AAIW and NPIW. Light blue in Indian: overlap of AAIW and RSW. Cross-hatching: mixing sites that are particularly significant for the water mass. Red dots indicate the primary formation site of each water mass; fainter dots mark the straits connecting the Mediterranean and Red Seas to the open ocean. The approximate potential density of formation is listed. This figure can also be found in the color insert. *Source: After Talley (2008).*

Characterization: Large-scale salinity maximum and minimum layers.
Location: Just below the pycnocline in most of the ocean (tropics-subtropics), ~1000 - 2000 m depth.
Formation mechanisms: Deep convection; brine rejection; vigorous mixing where boundary currents meet; otherwise nearly-isopycnal spreading.

Deep and bottom waters

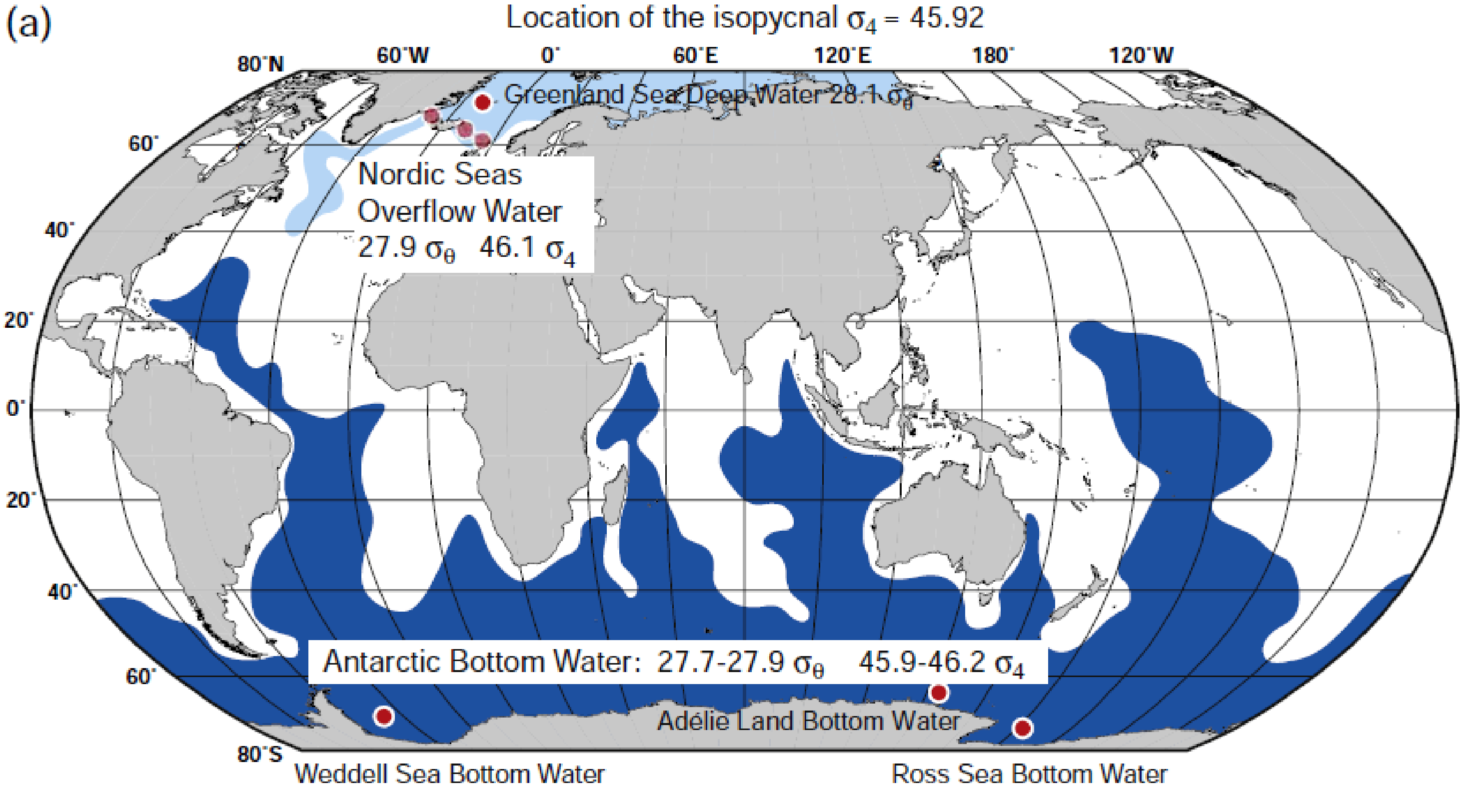


FIGURE 14.14 Deep and bottom waters. (a) Distribution of waters that are denser than $\sigma_4 = 45.92\ kg/m^3$. This is approximately the shallowest isopycnal along which the Nordic Seas dense waters are physically separated from the Antarctic's dense waters. At lower densities, both sources are active, but the waters are intermingled. Large dots indicate the primary formation site of each water mass; fainter dots mark the straits connecting the Nordic Seas to the open ocean. The approximate potential density of formation is listed. *Source: After Talley (1999).* (b) Potential temperature (°C), and (c) salinity at the ocean bottom, for depths greater than 3500 m. *Source: After Mantyla and Reid (1983).*

Deep waters

Characterization: This is a thick layer below the intermediate layer and above the bottom waters, characterized by extrema of salinity, nutrients.
Location: Roughly from 2000 - 4000 m depth.
Formation mechanisms: deep convection; brine rejection; upwelling (ocean-wide); vigorous mixing at specific sites (strait overflows); spreading along isopycnals with minimal mixing.

Deep and bottom waters

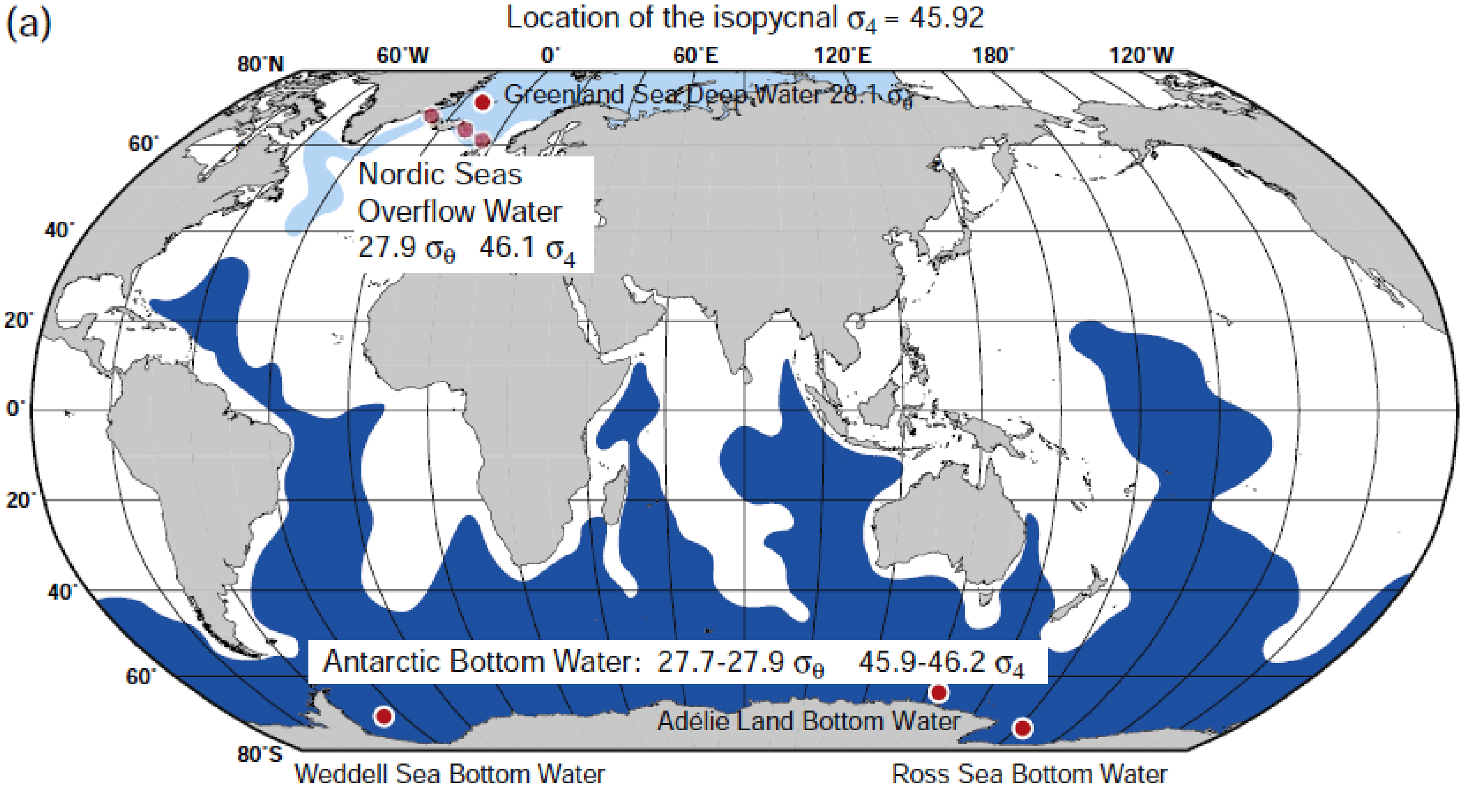
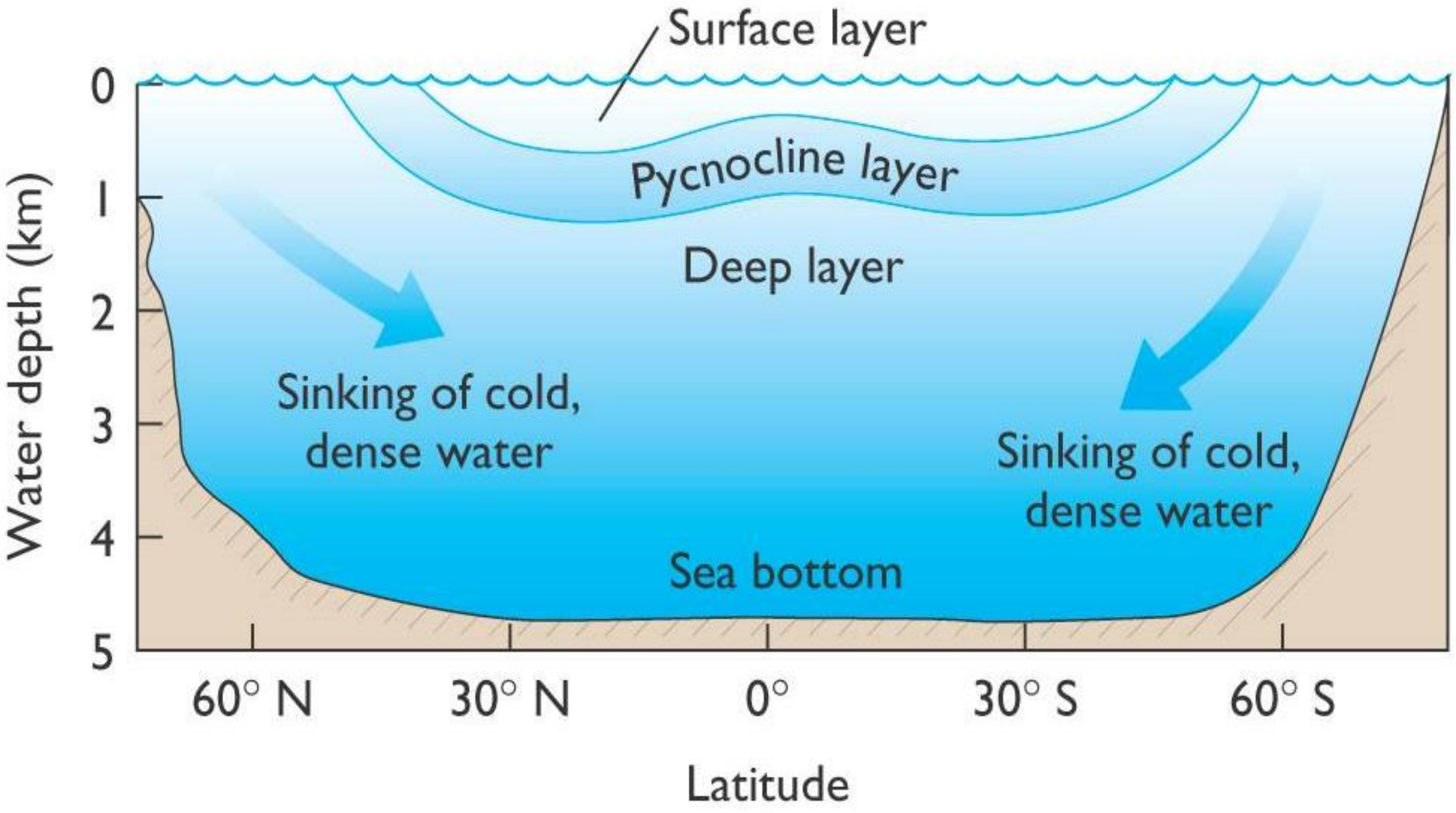


FIGURE 14.14 Deep and bottom waters. (a) Distribution of waters that are denser than $\sigma_4 = 45.92$ kg/m^3. This is approximately the shallowest isopycnal along which the Nordic Seas dense waters are physically separated from the Antarctic's dense waters. At lower densities, both sources are active, but the waters are intermingled. Large dots indicate the primary formation site of each water mass; fainter dots mark the straits connecting the Nordic Seas to the open ocean. The approximate potential density of formation is listed. *Source: After Talley (1999).* (b) Potential temperature (°C), and (c) salinity at the ocean bottom, for depths greater than 3500 m. *Source: After Mantyla and Reid (1983).*

Bottom water

Characterization: Densest, coldest layer.
Location: ocean bottom, usually connotes very dense water from the Antarctic.
Formation mechanisms: brine rejection close to Antarctica.

Deep and bottom waters



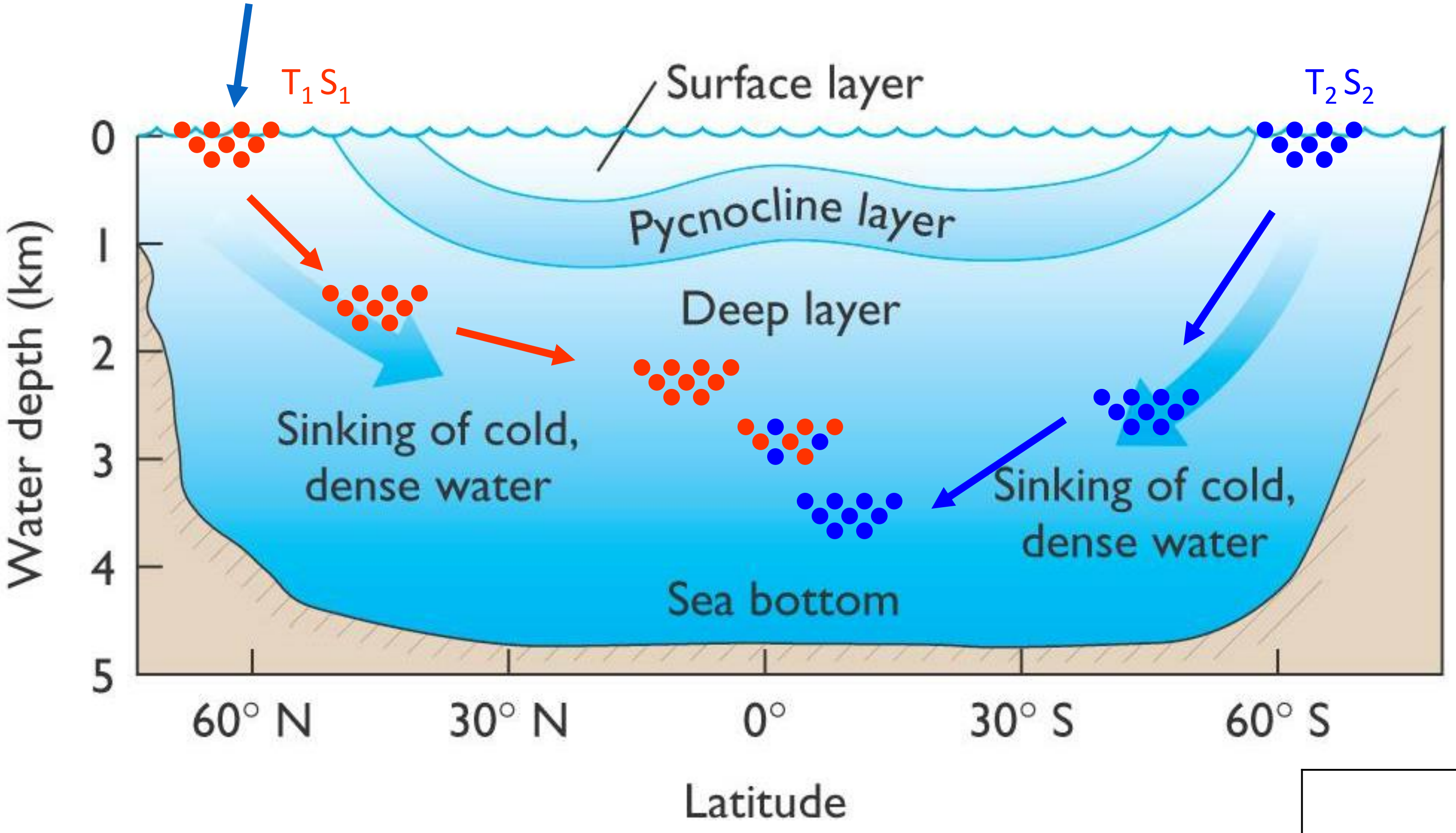
Tropical oceans: pycnocline ≈ thermocline

Mid-latitudes: pycnocline ≈ halocline

High latitudes: no pycnocline formation

Deep and bottom waters

Diabatic exchanges with the atmosphere at the surface



Adiabatic changes and *Mixing* in ocean interior

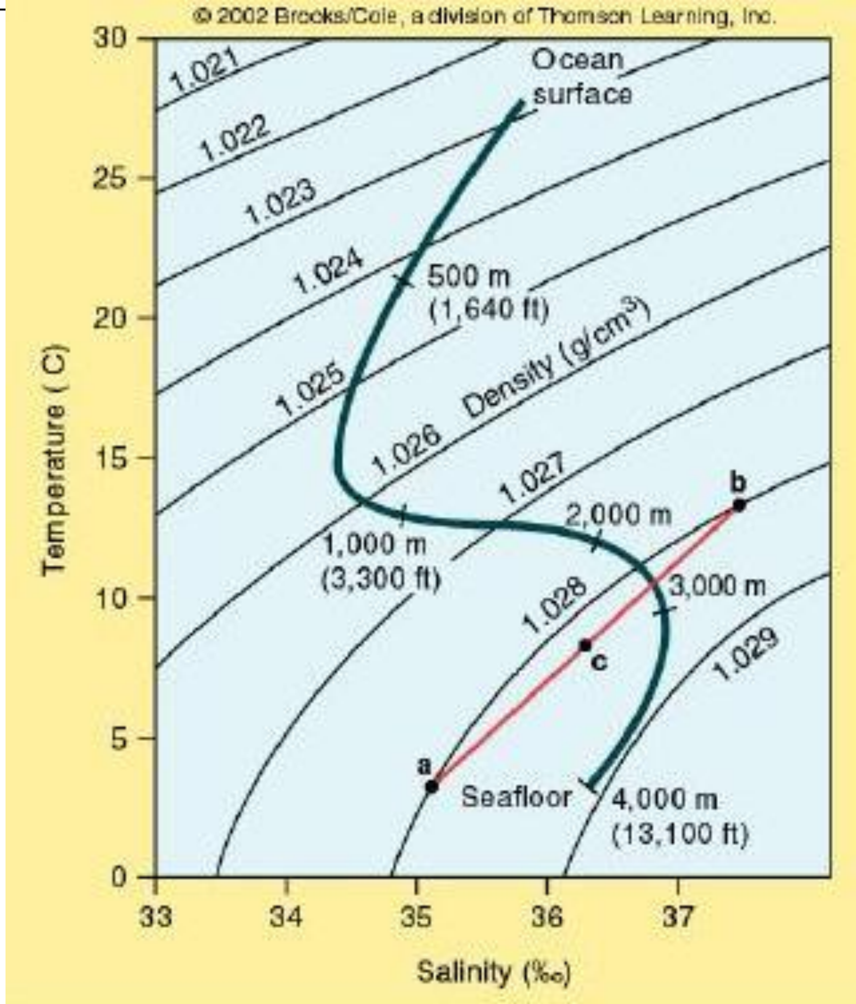
Season	Mean temperature	
	North Pole	**South Pole**
Summer	0° C	−28.2° C
Winter	−40° C	−60° C

Why is the south pole colder than the north?

T-S Diagram - Exercise

The data shown is for a station in the North Atlantic, at about 20° N latitude.

Plot the data on the T-S chart. Below is a sample T-S diagram



Depth (m)	T (°C)	S (‰)
100	16.0	36.1
200	13.0	35.8
400	11.0	35.5
500	9.0	35.3
600	8.0	35.0
850	13.0	37.3
950	12.5	37.1
1200	11.0	36.7
1500	4.8	34.8
2000	4.0	35.0
2200	3.5	34.9
2500	2.0	34.8
3000	0.0	34.7
4000	-1.9	34.6
5000	-2.0	34.6

T-S Diagram - Exercise

Identify the following water masses in the T-S diagram

Water mass	Source	Identifying charecteristic
Surface water masses	Regional	Variable, generally warm
Mediterranean intermediate water (MIW)	Mediterranean Sea off Turkey	High Salinity and temperature tongue at intermediate depths
North Atlantic deep water (NADW)	Area near Greenland	Intermediate salinity maximum, may show intermediate temperature maximum
Antartic bottom water	Weddell Sea	Bottom temperature minimum

Water Types and Water Masses

W J Emery, University of Colorado, Boulder, CO, USA

Introduction

Much of what is known today about the currents of the deep ocean has been inferred from studies of the water properties such as temperature, salinity, dissolved

problems, there is very little research directed at improving our knowledge of water mass distributions and their changes over time.

What is a Water Mass?

The concept of a 'water mass' is borrowed from meteorology, which classifies different atmospheric characteristics as 'air masses'. In the early part of the

Maximum mixed layer depth [mainly late winter in each location]

Surface layer of the ocean is almost always vertically mixed to some degree.

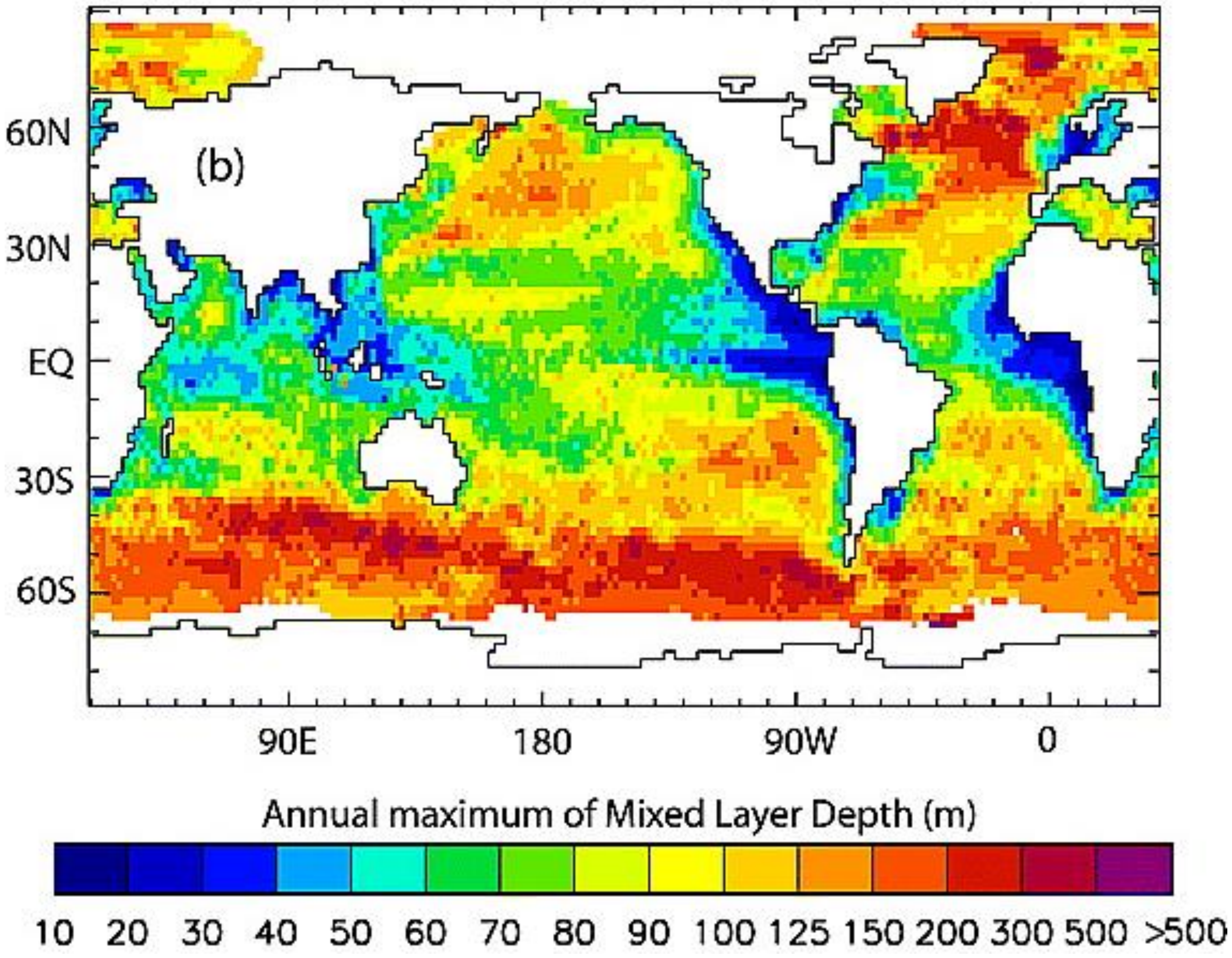
In summer, calm, warm conditions, the mixed layer might be very thin (several meters).

At the end of winter, after the full season of cooling and storms, mixed layers reach their maximum thickness.

Mixed layers are created by –

Wind stirring (max. depth of around 100 m)

Cooling and evaporation (increasing the density of the surface water), which creates vertical convection. Max. depth can range up to about 1000 m, but is mainly 200-300 m.



deBoyerMontegut et al. (JGR, 2004)

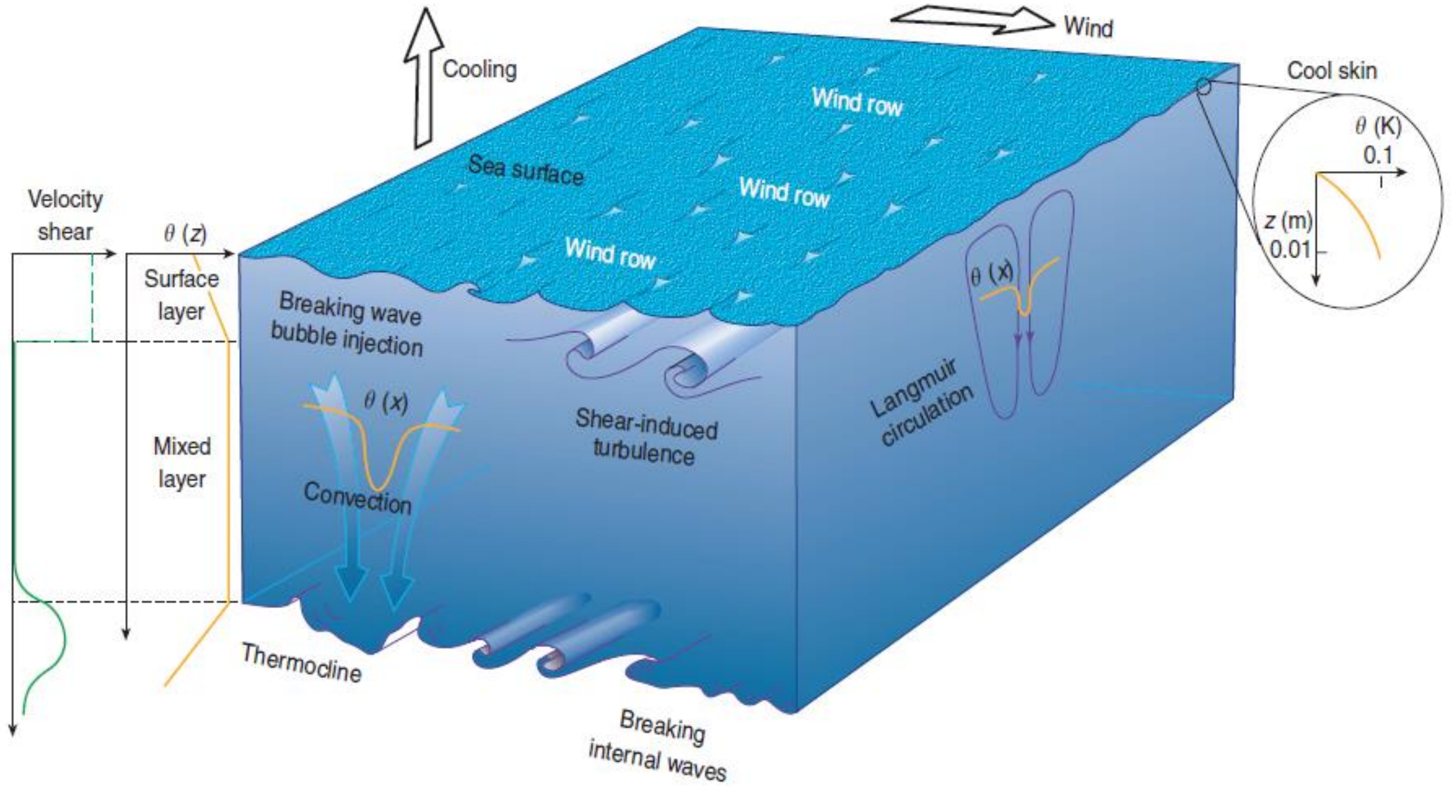
Upper ocean mixing processes

Convection: Cooling at the sea surface creates parcels of cool, dense fluid, which later sink to a depth determined by the local stratification in a process known as convection.

Wind Forcing: Convection is aided by wind forcing, in part because winds help to disrupt the viscous sub-layer at the sea surface, permitting more rapid transport of heat through the surface.

Breaking waves: Turbulent dissipation of energy.

Langmuir circulations are coherent structures within the mixed layer that produce counter rotating vortices with axes aligned parallel to the wind – due to particular type of winds blowing steadily over the sea surface.



Wind driven shear

Induces gradient of horizontal velocity, turbulence, internal gravity waves, etc.

Upper ocean mixing processes

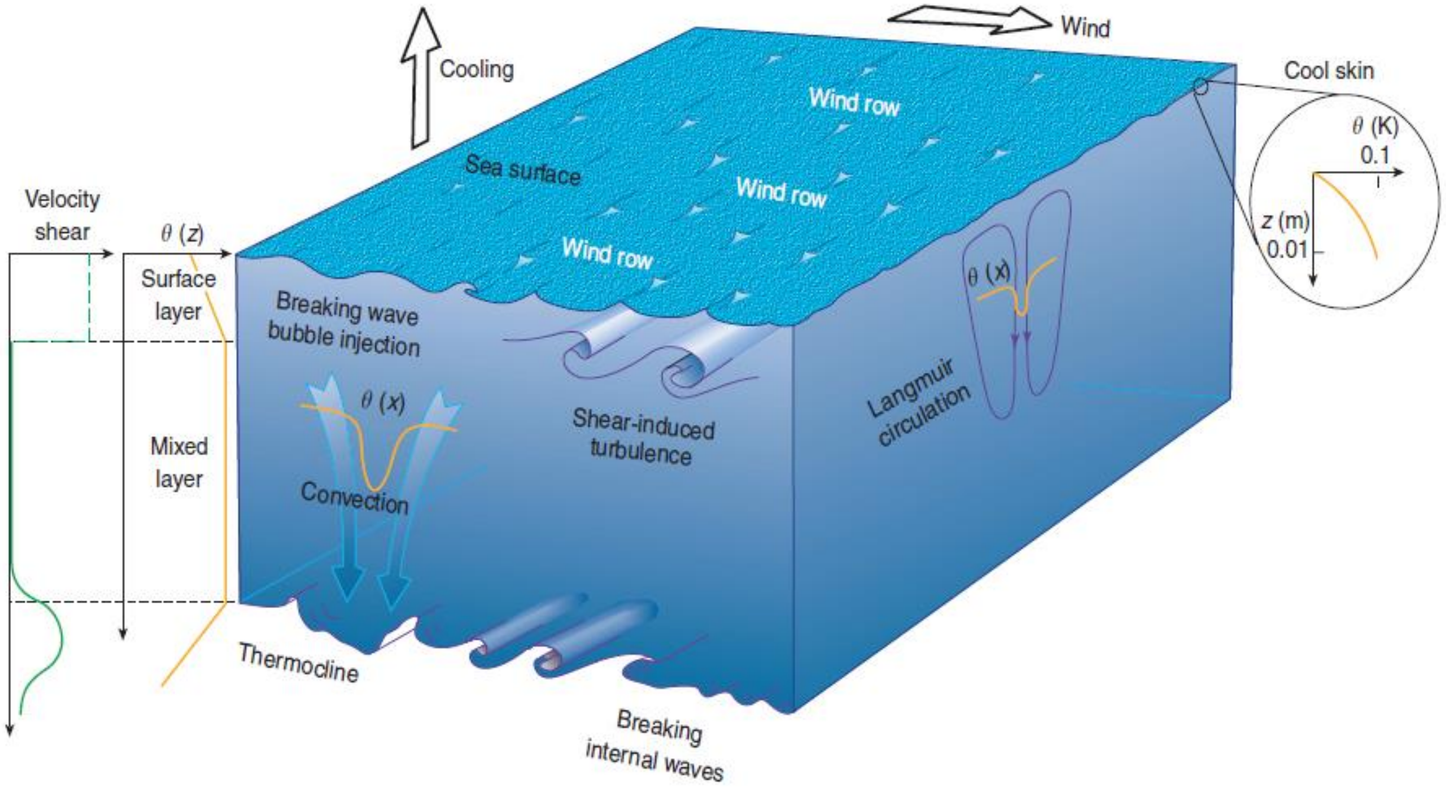
Precipitation: Rainfall on the sea surface can catalyze several important processes that act to both accentuate and reduce upper ocean mixing.

Drops falling on the surface disrupt the viscous boundary layer, and may carry air into the water by forming bubbles.

Smaller waves (20cm wavelength) may be damped by subsurface turbulence as heavy rainfall acts to transport momentum vertically, causing drag on the waves.

Fresh water – eg: Bay of Bengal, changes the thermodynamic properties – stratification etc.

Evaporative cooling – latent heat transfer



UPPER OCEAN MIXING PROCESSES

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doi:10.1006/rwos.2001.0156

Introduction

The ocean's effect on weather and climate is governed largely by processes occurring in the few tens of meters of water bordering the ocean surface. For ocean mixing. An example is shown in **Figure 1**, which illustrates mixed-layer[1] evolution, temperature structure and small-scale turbulence. The small white dots in **Figure 1** indicate the depth above which stratification is neutral or unstable and mixing is intense, and below which stratification is stable and mixing is suppressed. This represents a means of determining the vertical extent of the mixed layer directly forced by local atmospheric conditions. (We will call the mixed-layer depth *D*.) Following the change in sign from negative (surface

Ocean surface heat budget

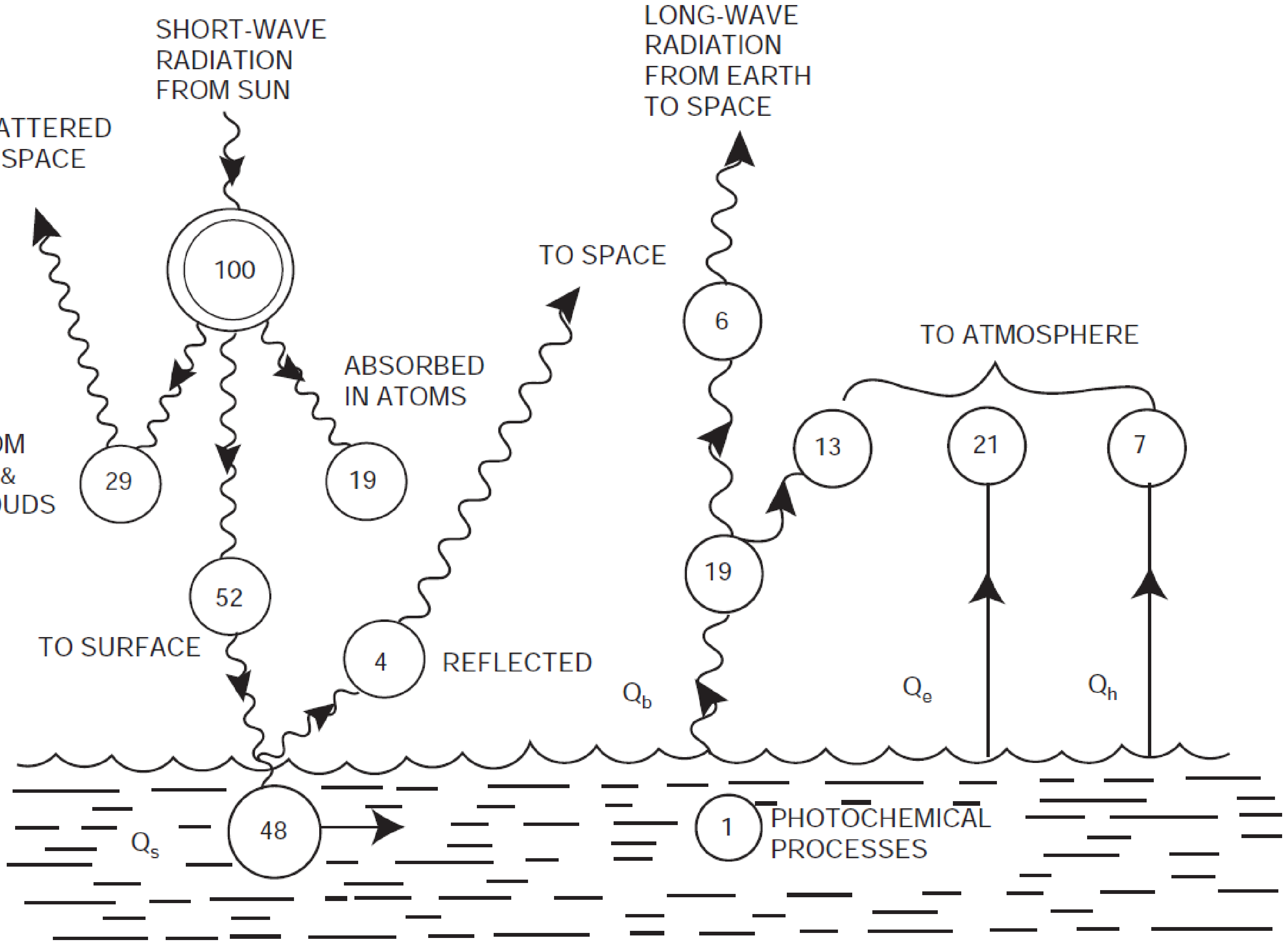
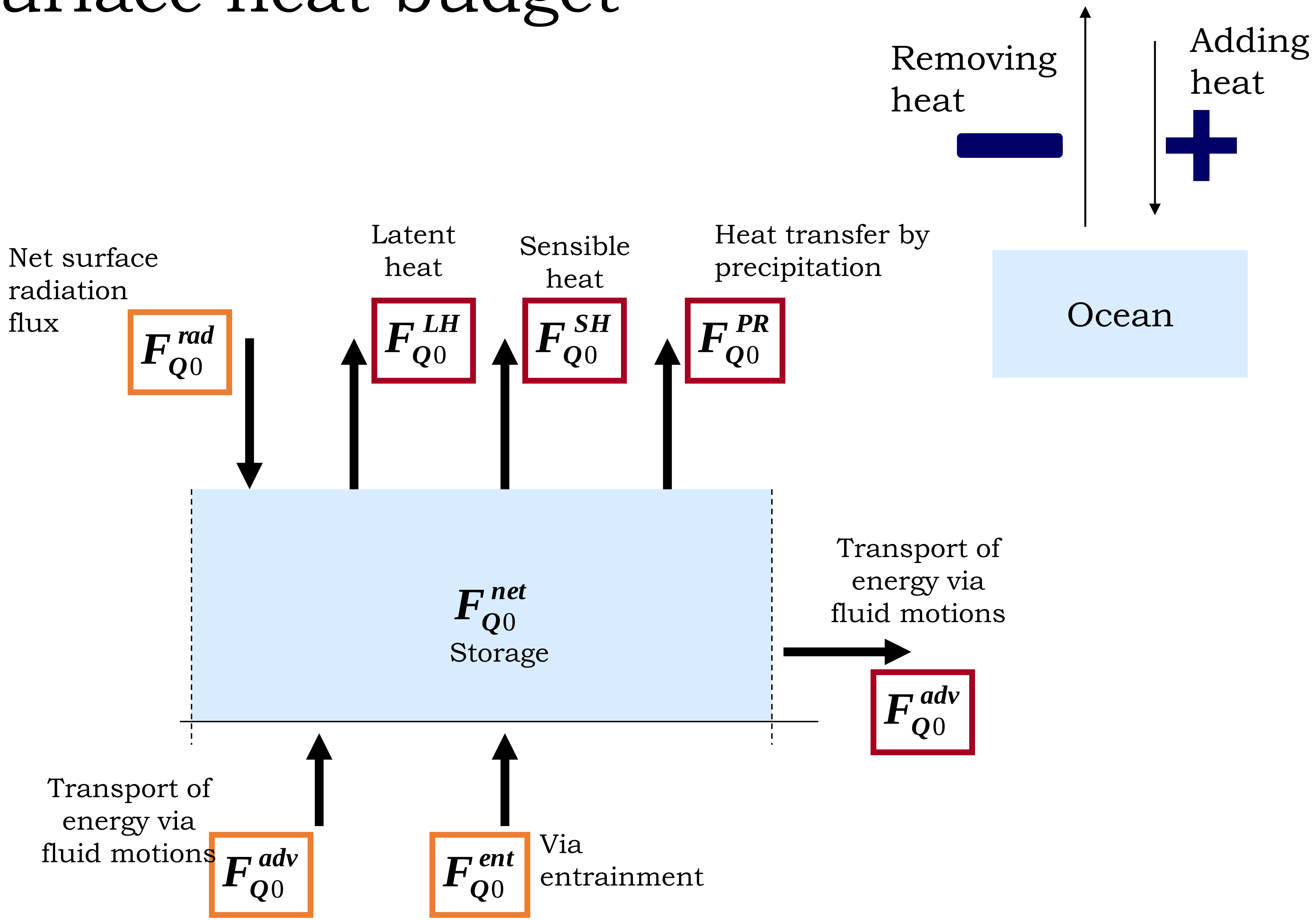


FIGURE 5.5 Distribution of 100 units of incoming shortwave radiation from the sun to Earth's atmosphere and surface: long-term world averages.

Ocean surface heat budget

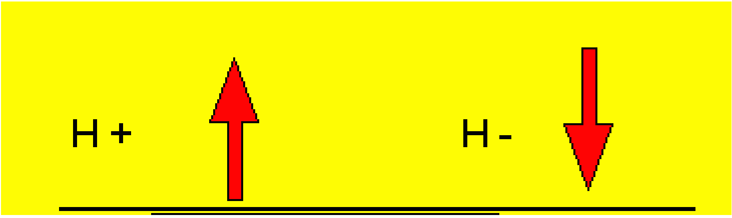


Sensible heat flux

Sensible heat is heat energy transferred between the surface and air when there is a difference in temperature between them. A change in temperature over distance is called a "temperature gradient".

Heat is initially transferred into the air by conduction as air molecules collide with those of the surface. As the air warms it circulates upwards via convection. Thus the transfer of sensible heat is accomplished in a two-step process.

When the surface is warmer than the air above, heat will be transferred upwards into the air as a positive sensible heat transfer. If the air is warmer than the surface, heat is transferred from the air to the surface creating a negative sensible heat transfer.



Latent heat flux

When energy is added to water it will change states or phase. The phase change of a liquid to a gas is called evaporation. The heat added during evaporation breaks the bonds between the clusters of water molecules creating individual molecules that escape the surface as a gas. The heat used in the phase change from a liquid to a gas is called the latent heat of vaporization. We say it is "latent" because it is being stored in the water molecules to later be released during the condensation process. We can't sense or feel latent heat as it does not raise the temperature of the water molecules.

When evaporation is taking place we say there is a positive latent heat flux (transfer). A positive latent heat flux is illustrated with an arrow pointing up away from the surface of the earth. Evaporation is a cooling process for a surface because energy is removed from the water as molecules escape the surface. This causes the surface temperature to decrease. You've probably experienced this cooling when water or sweat evaporates from your skin. Condensation is the phase change from a gas to a liquid (negative latent heat flux).

Surface turbulent heat fluxes

Sensible heat flux → $F_{Q0}^{SH} = \rho c_{pd} \boxed{\left(\overline{w'\theta'}\right)_0}$

Latent heat flux → $F_{Q0}^{LH} = \rho L_{lv} \boxed{\left(\overline{w' q_v'}\right)_0}$

Covariances

High-frequency measurements
↓
Rarely available
↓
Estimate in terms of other parameters

Bulk aerodynamic formulae

Near-surface turbulence arises from the mean wind shear over the surface

Turbulent fluxes of heat and moisture are proportional to their gradients just above the ocean surface

Surface turbulent heat fluxes

Bulk aerodynamic formulae

$$F_{Q0}^{SH} = \rho c_p \boxed{C_{DH}} \left(u_a - u_0\right)\left(\theta_a - \theta_0\right)$$

$$F_{Q0}^{LH} = \rho L_{lv} \boxed{C_{DE}} \left(u_a - u_0\right)\left(q_{va} - q_{v0}\right)$$

Aerodynamic transfer coefficients Under Ordinary conditions

$$\boxed{C_{DH} = C_{DE}} = \frac{k^2}{\left(\ln \frac{z_a}{z_0}\right)^2} f\boxed{\left(Ri_B\right)}$$

$Ri_B > 0 \longrightarrow$ Stable

$Ri_B = 0 \longrightarrow$ Neutral

$Ri_B < 0 \longrightarrow$ unstable

Just above the surface

Richardson number

Why is LH much larger than SH over the tropics?

Clausius-Clapeyron equation shows that the amount of water vapour in the air at saturation is strongly dependent on the temperature.

i.e. **When temperature increases, moisture holding capacity of air increases**.

- Cool air cannot hold much water vapor, so is typically dry
- Warm air can hold more water vapor, so is typically moist

Eg: the amount of water vapour that can be present in the atmosphere at a temperature of 20°C is more than three times higher than at 0°C.

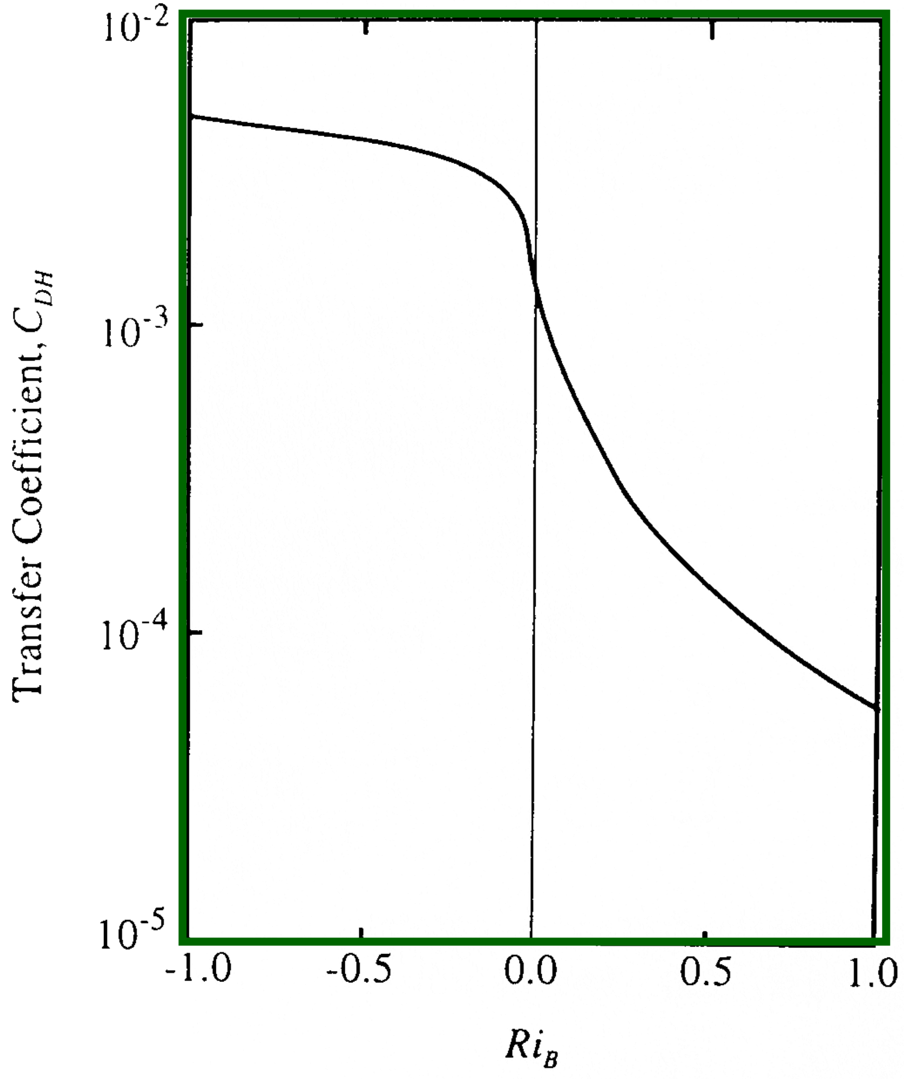
As a consequence, the evaporation and the latent heat flux are much larger at low latitudes than at high ones. The latent heat flux is thus larger than the sensible heat flux at low latitudes, while the two fluxes are generally of the same order of magnitude over the ocean at high latitudes.

$$\boxed{B_0 = \frac{F_{Qo}^{SH}}{F_{Qo}^{LH}}}$$

Bowen Ratio: is used to describe the type of heat transfer in a water body:

If B <1, a greater proportion of the available energy at the surface is passed to the atmosphere as latent heat than as sensible heat

Aerodynamic transfer coefficients



$\boldsymbol{Ri_B} > 0 \longrightarrow$ Stable

$\boldsymbol{Ri_B} = 0 \longrightarrow$ Neutral

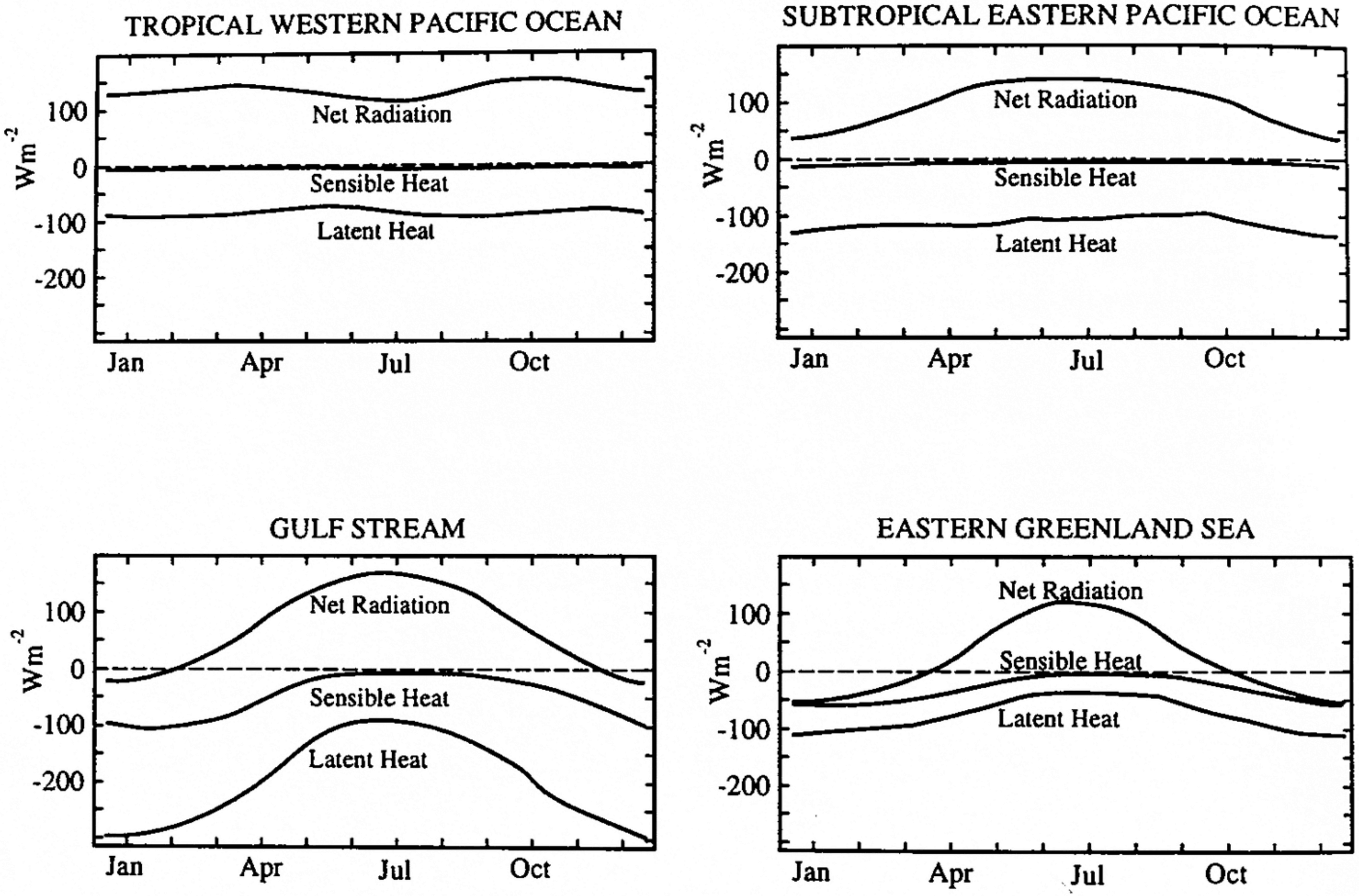
$\boldsymbol{Ri_B} < 0 \longrightarrow$ unstable

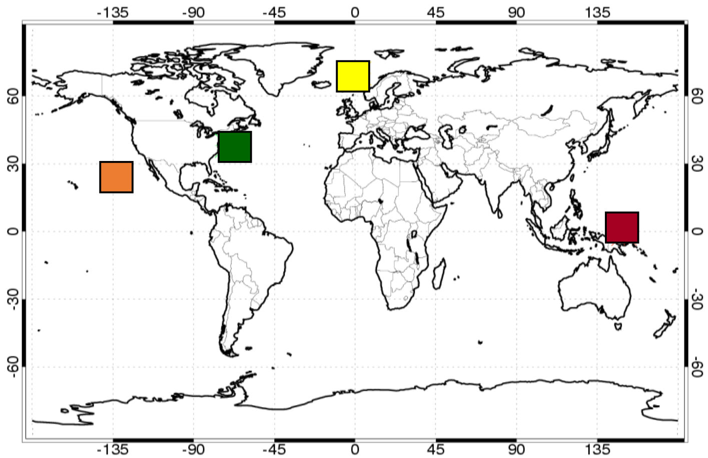
Small for statically stable conditions

Large for unstable conditions

The magnitude of the heat transfer is inversely proportional to the degree of stability

Variation of surface energy budget components





Bowen Ratio

$$\boxed{B_0 = \frac{F_{Qo}^{SH}}{F_{Qo}^{LH}}}$$

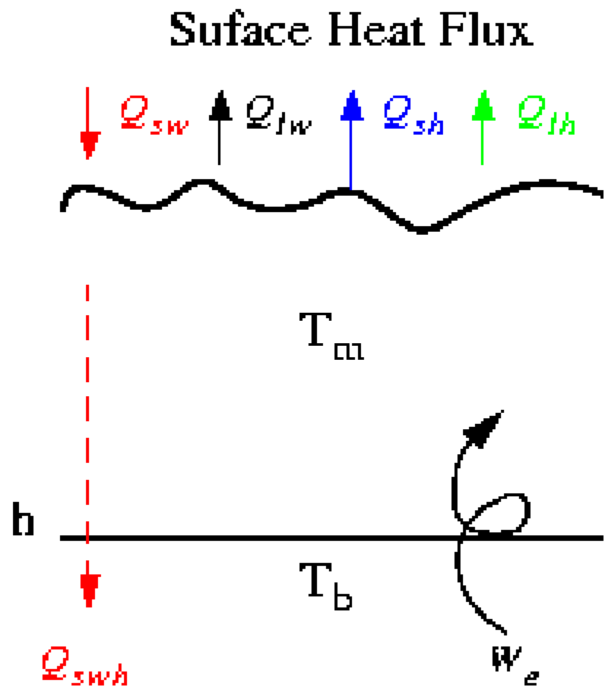
SST Tendency Equation

Integrated heat budget over the mixed layer

$$\frac{\partial T_m}{\partial t} = \frac{Q_{net} - Q_{swh}}{\rho c h} + \left(\frac{w + w_e}{h}\right)(T_b - T_m) - \vec{v} \cdot \nabla T_m + A \nabla^2 T_m$$

Variables

- v – velocity (current in ML)
- T_m – mixed layer temp (SST)
- T_b – temp just beneath ML
- h – mixed layer depth
- w – mean vertical velocity
- w_e – entrainment velocity
- Q_{net} – net surface heat flux
- Q_{swh} – penetrating shortwave radiation
- A – horizontal eddy viscosity coefficient
- ρ – density of sea water
- C – Specific heat of sea water



Temperature change due to the surface heat flux

Over March through August a location in the North Pacific typically receives 150 Wm^{-2} flux through the surface. Assuming a constant mixed layer depth of 50 m, and no other changes in the ocean how much will the SST change over that time?

SST Tendency

$dT_m/dt = d(SST)/dt = Q_{net}/\rho ch$

$\Delta SST = (Q_{net}/\rho ch) \times \Delta t$

$\rho = 1025\ kg/m^3$; $c = 3850$ Joules/(kg °C)

What is the winter to summer change in SST ?

Temperature change due to the surface heat flux

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What is the winter to summer change in SST ?

$\Delta SST = (150\ Wm^{-2}\ /\ (1025\ kg\ m^{-3} \times 3850\ Joules\ kg^{-1}\ °C^{-1} \times 50\ m)) * (184\ days * 86400\ s\ day^{-1})$

Importance of the Ocean in Earth's Heat Budget

Land-Ocean comparison:

Heat capacity of soil/rocks and water,

$C_{p(rock)}$ = 800 J kg^{-1} °C^{-1} $C_{p(water)}$ = 4000 J kg^{-1} °C^{-1}

The volume of water which exchanges heat with the atmosphere on a seasonal cycle is 100 m^3 per square meter of surface, i.e. that mass from the surface to a depth of 100m. The density of water is 1000 kg/m^3, and the mass in contact with the atmosphere is density × volume = m_{water} = 100,000 kg. The volume of land which exchanges heat with the atmosphere on a seasonal cycle is 1 m^3. Because the density of rock is 3,000 kg/m^3, the mass of the soil and rock in contact with the atmosphere is 3,000 kg.

The seasonal heat storage values for the ocean and land are therefore:

$\Delta E_{oceans} = C_{p(water)}\, m_{water}\, \Delta T$ (ΔT is the typical change in temperature from summer to winter)

ΔT = 10°C = (4000)(10^5)(10°) Joules = 4.0 X 10^9 Joules

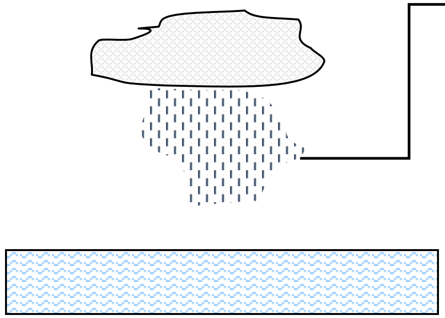
$\Delta E_{land} = C_{p(rock)}\, m_{rock}\, \Delta T$

ΔT = 20°C = (800)(3000)(20°) Joules = 4.8 X 10^7 Joules

$\Delta E_{oceans} / \Delta E_{land} = 100$

Heat flux for precipitation

heat transfer occurs if the precipitation is at different temperature than the surface !!!



Temperature of the rain drop

If thermal equilibrium

T_{rain}= wet bulb T of the atmosphere

$$\boxed{F_{Qo}^{PR}} = \rho_l c_{pl} \dot{P}_r \left(T_{Wa} - T_o\right)$$

Greatest for large rainfall rates and large differences in temperature

Usually → $\boxed{T_{Wa} < T_o}$ → **Heat flux from rain cools the ocean**

Long term contribution to surface energy budget small → Commonly Neglected

Snow?? → Latent heat Melt Snow →

$$F_{Qo}^{PR} = \rho_s c_{ps} \dot{P}_s \left(T_{la} - T_o\right) - \rho_s L_{il} \dot{P}_s$$

$$\boxed{\frac{c_{ps}\left(T_{Ia} - T_0\right)}{L_{il}} = 0.0063\left(T_{Ia} - T_0\right)}$$

The latent heat is an order of magnitude larger than sensible heat term

Ocean surface salinity budget

Precipitation

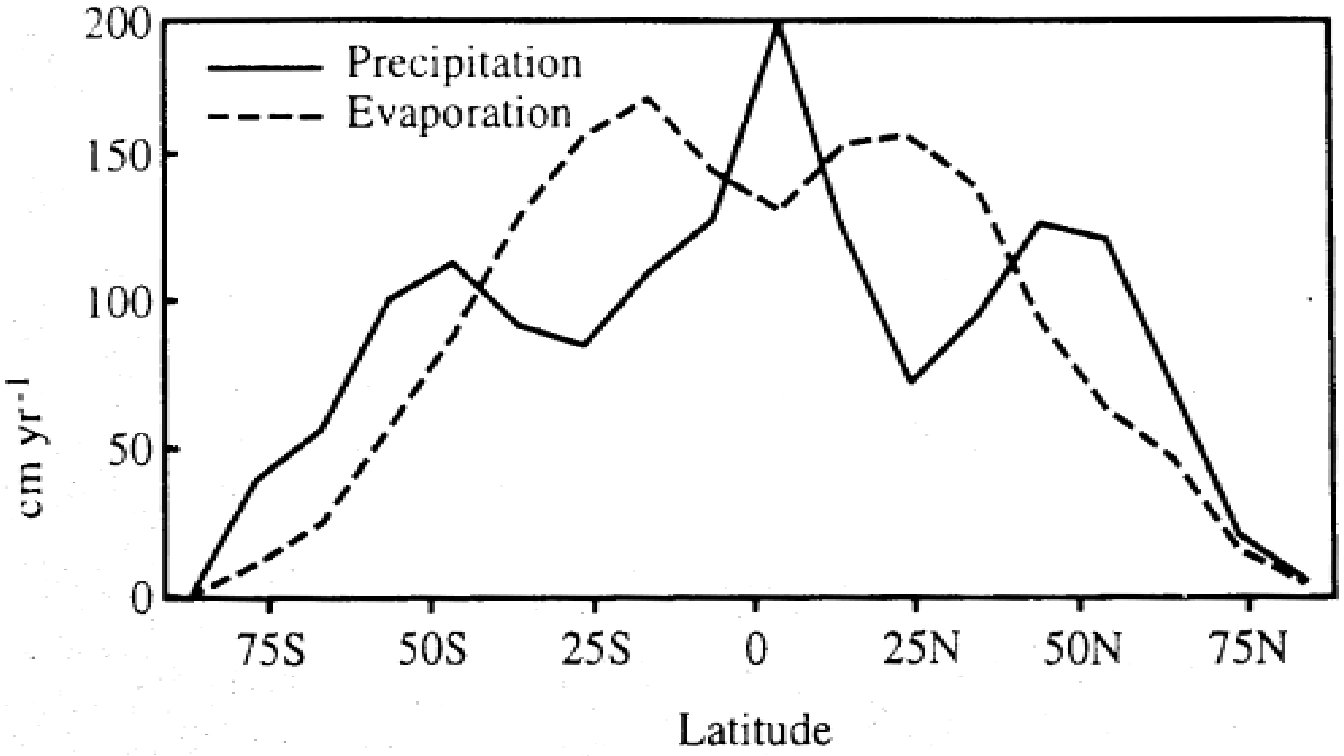
Evaporation

Formation of sea ice

Melting of sea ice

River runoff

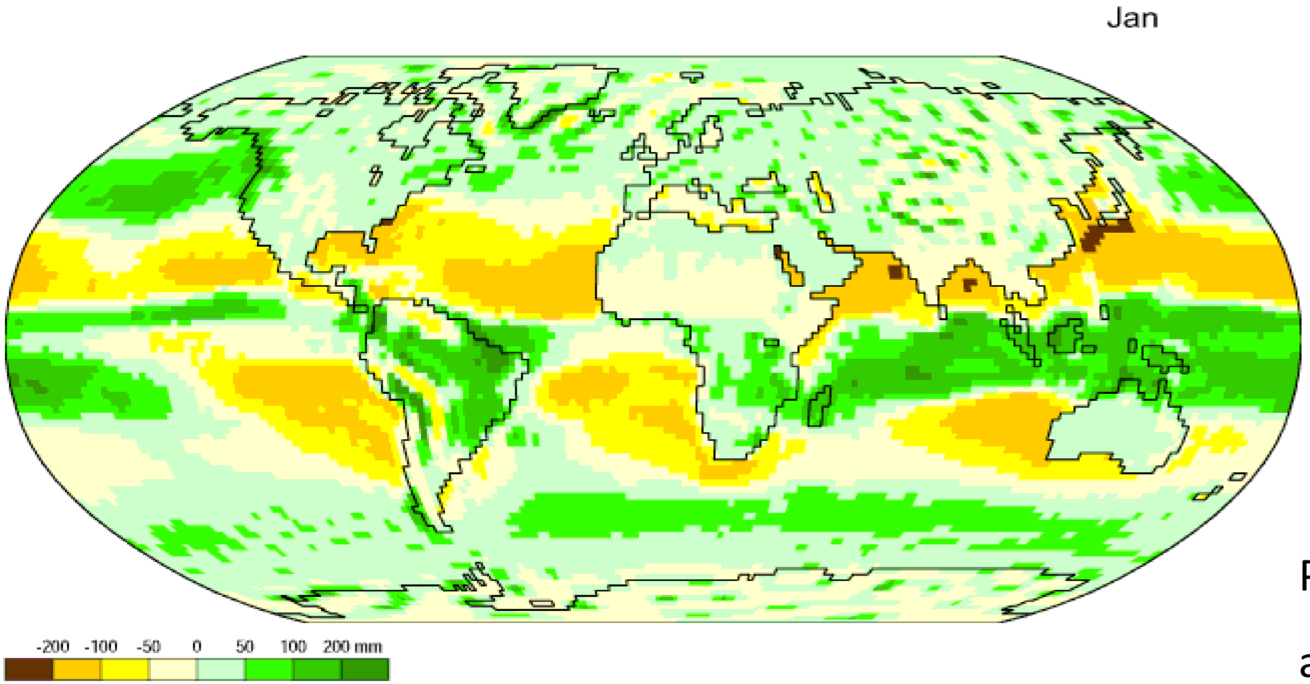
Storage transport below the ocean surface



Important regional differences

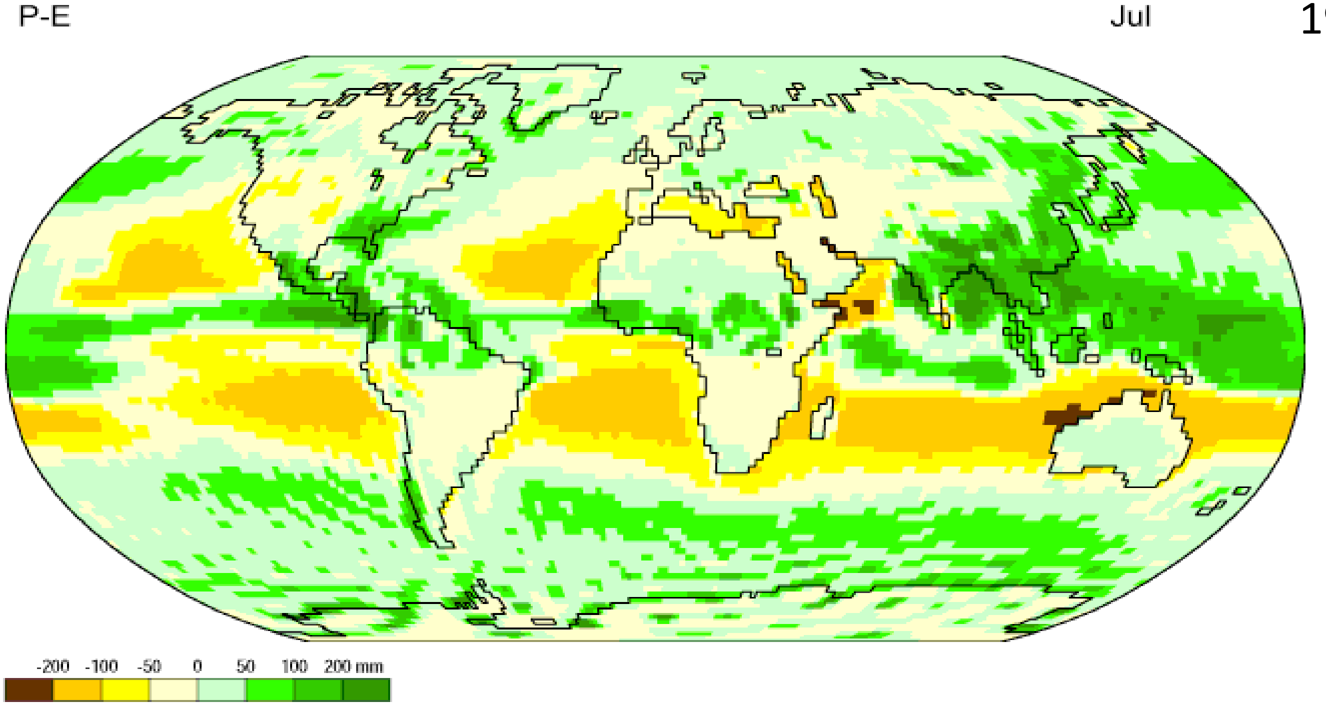
mm/yr	$\dot{P}$	$\dot{E}_0$
Artic Ocean	97	53
Atlantic Ocean	761	1133
Indian Ocean	1043	1294
Pacific Ocean	1292	1202
All Oceans	1066	1176

P-E



P-E

average
1959-1997



Ocean Surface Buoyancy flux

$$F_{B0} = g\left(\frac{\alpha}{c_{p0}} F_{Q0}^{net} - \beta F_{s0}^{net}\right)$$

Negative value meets the instability criterion → Sinking motion in the ocean

Evaporation → Increases the buoyancy flux

Ratio of the cooling term to the salinity term of evaporation → $\boxed{\dfrac{\alpha L_{lv}}{\beta c_p s_0}}$

- Tropics: T=30 C; s=35 psu → $\boxed{8.0}$
- High latitudes: T=0 C; s=35 psu → $\boxed{0.6}$

Precipitation → decreases and increases the buoyancy flux

Freshening effects of rain dominate the cooling effects of rain at ***all latitudes***

Snow → Freshening dominates the effect on the buoyancy flux

Ice/Ocean

Heat flux terms that influence the surface

Penetration of solar radiation beneath the ice

Latent heat associated with freezing or melting ice

Sea Ice grows

Increase salinity

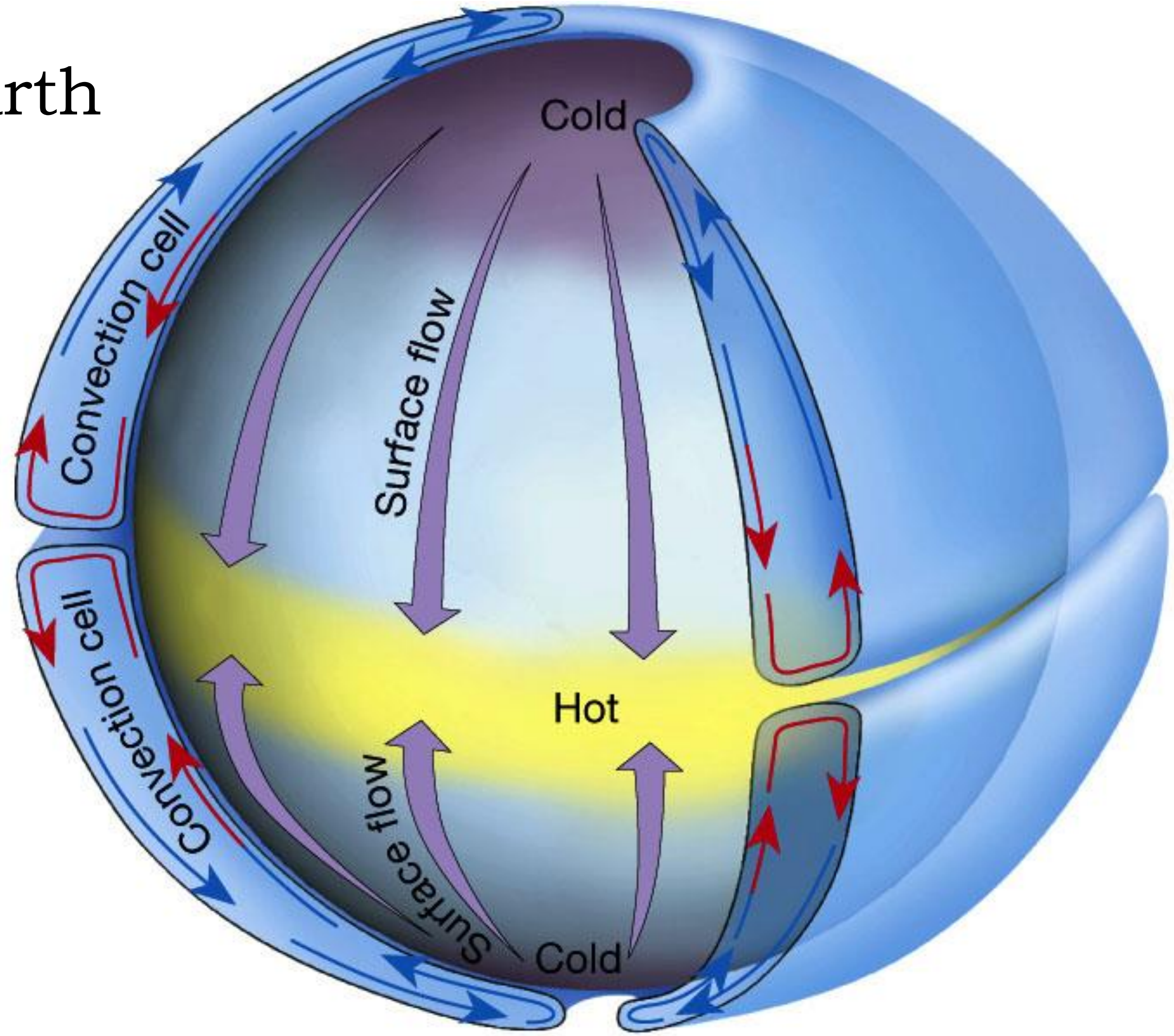
releases latent heat

Typical polar conditions

Salinity term dominates in determining ocean surface buoyancy flux

Coriolis Effect

Non-rotating Earth

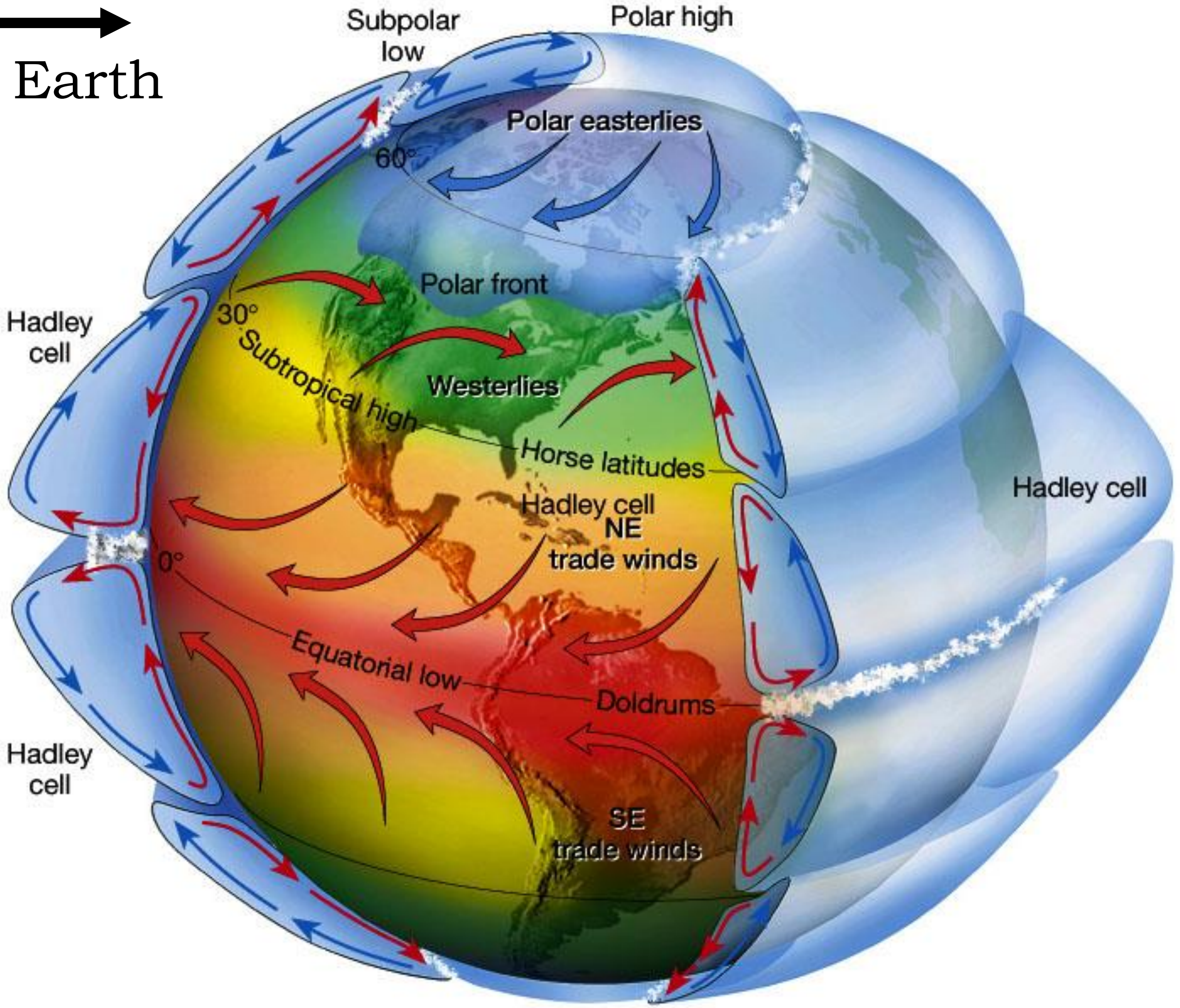


General circulation of the oceans - wind driven

Coriolis Effect

Add rotation and add landmasses ⟶ unequal heating and cooling of the Earth

- The Coriolis effect
  - Is a result of Earth's rotation
  - Causes moving objects to follow curved paths:
    - In Northern Hemisphere, curvature is to right
    - In Southern Hemisphere, curvature is to left
  - Changes with latitude:
    - No Coriolis effect at Equator
    - Maximum Coriolis effect at poles

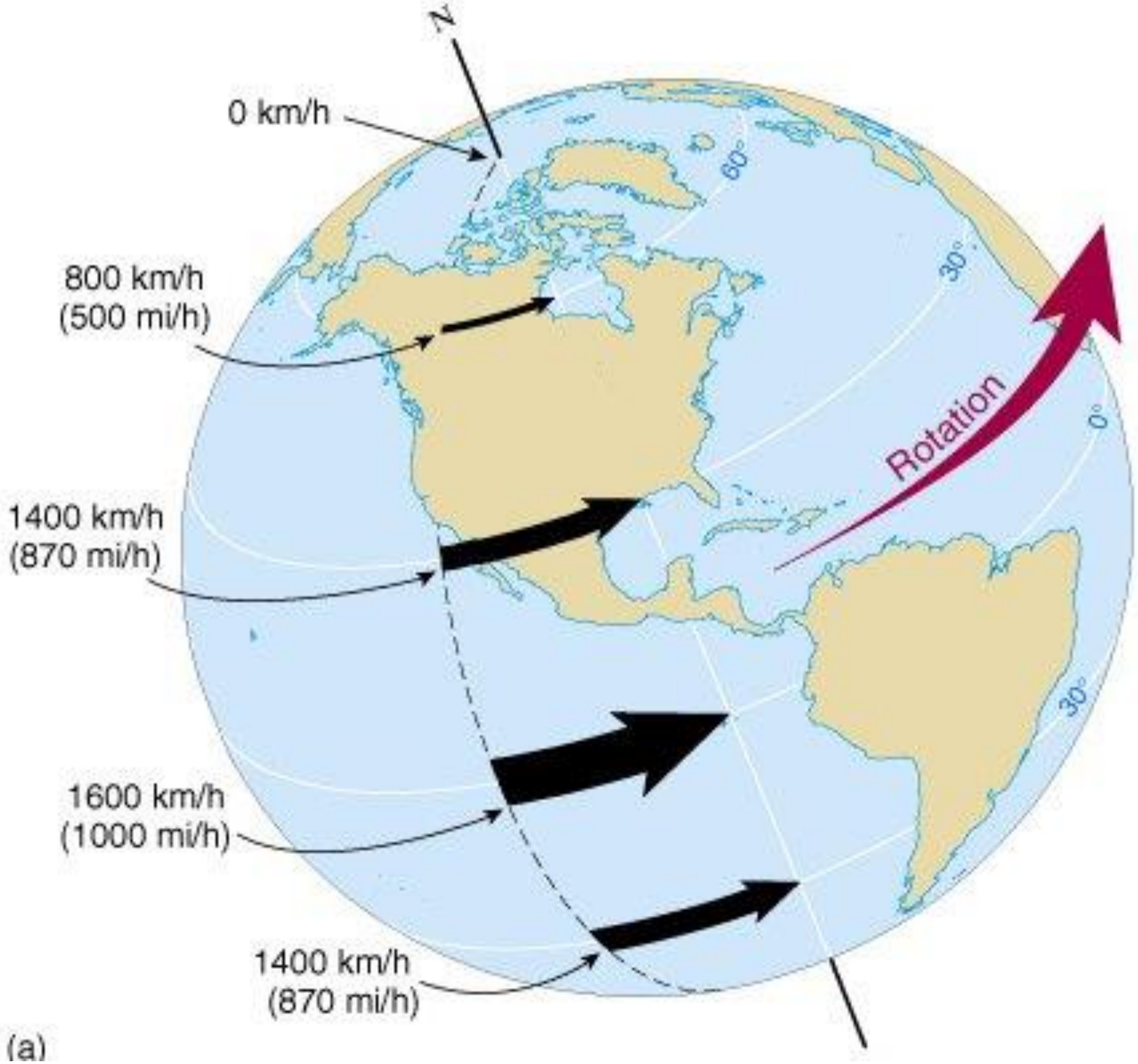


General circulation of the oceans - wind driven

Coriolis Effect

- As Earth rotates, different latitudes travel at different speeds.

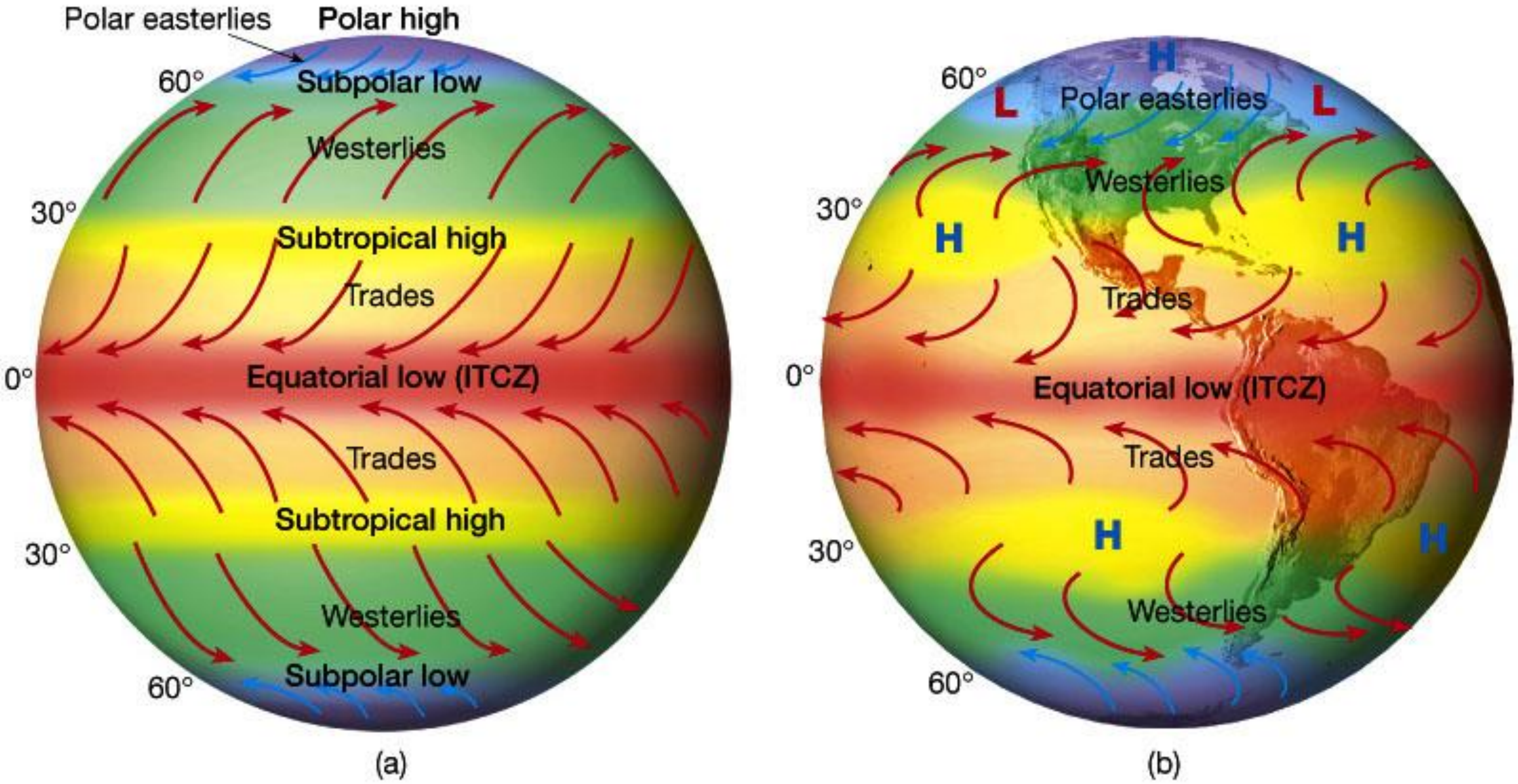
- The change in speed with latitude causes the Coriolis effect.



(a)

General circulation of the oceans - wind driven

Coriolis Effect



A) Idealized winds generated by pressure gradient and Coriolis Force.

B) Actual wind patterns owing to land mass distribution..

General circulation of the oceans - wind driven

Coriolis Effect

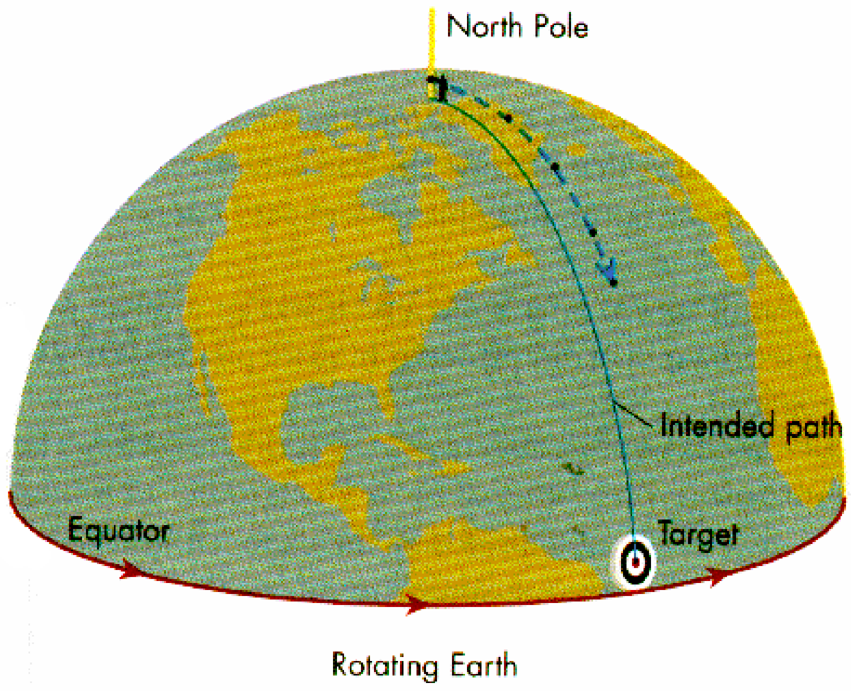
Inertial motion:
motion in a straight line relative to the fixed stars

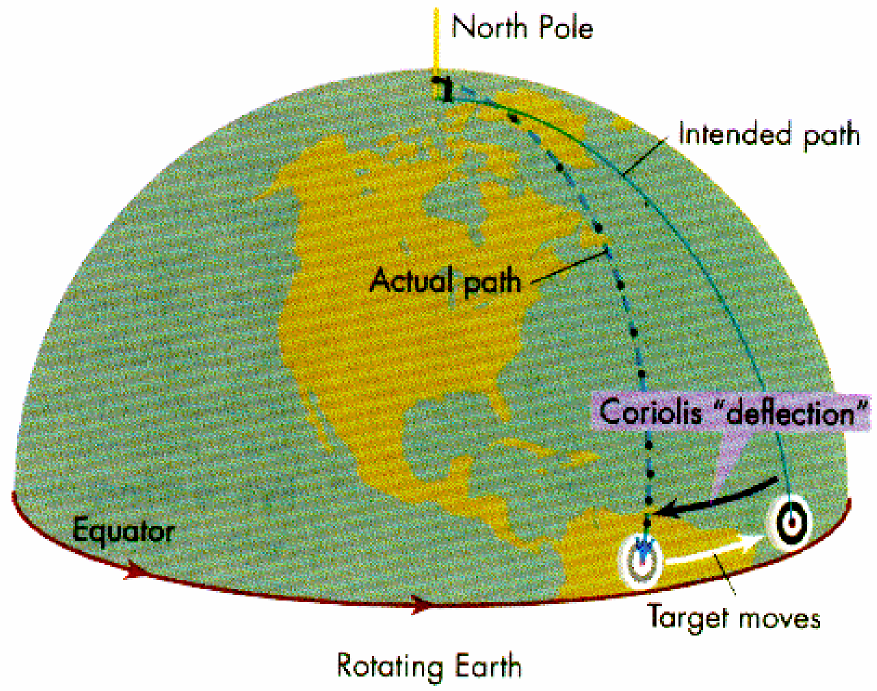
Coriolis effect:
apparent deflection of that inertially moving body just due to the rotation of you, the observer.

Coriolis effect deflects bodies (water parcels, air parcels) to the right in the northern hemisphere and to the left in the southern hemisphere

$f = 2\Omega \sin\varphi$ the "Coriolis parameter".

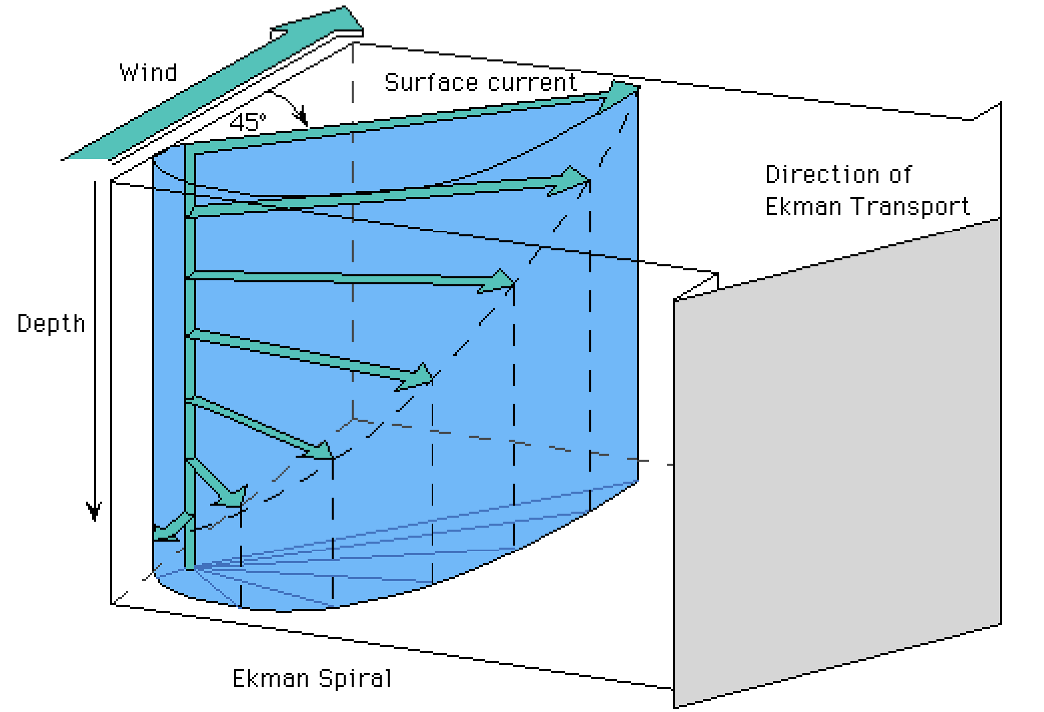
- depends on latitude (φ)
- Ω angular velocity/ rotational speed





Ekman velocity spiral

- Surface velocity to the right of the wind (northern hemisphere, due to Coriolis)
- Surface layer pushes next layer down slightly the right, and slightly weaker current
- Next layer pushes next layer, slightly to right and slightly weaker current
- Producing a “spiral” of the current vectors, to right in northern hemisphere, decreasing speed with increasing depth
- Details of the spiral depend on the vertical viscosity (how frictional the flow is, and also whether “friction” depends on depth)

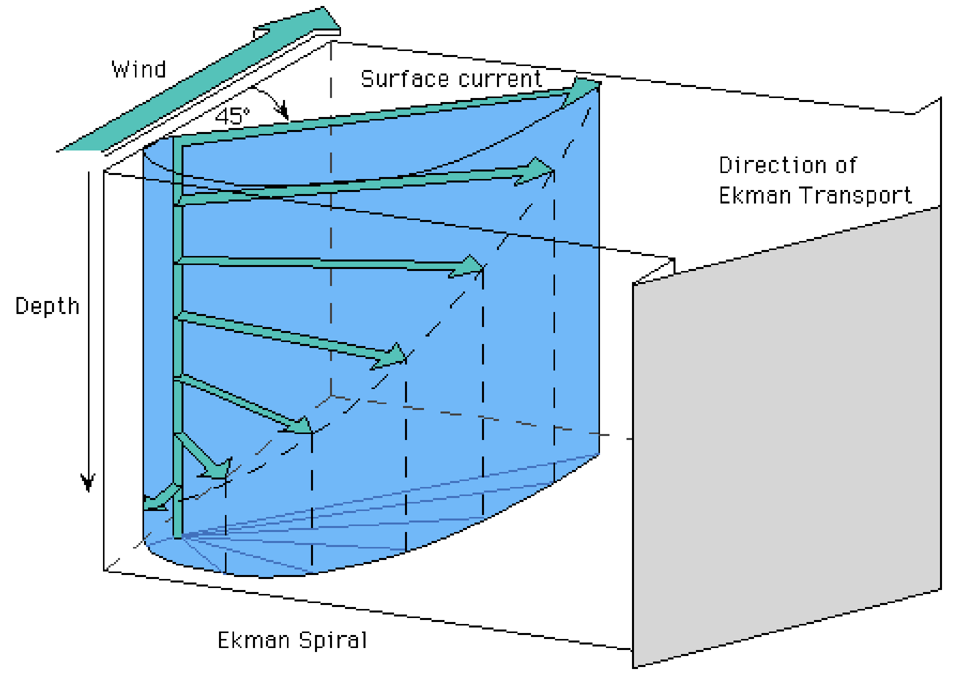


Adapted from Thurman, Harold V. **Essentials of Oceanography, 5th ed.** Prentice-Hall, Inc., 1996.

Ekman layer transport

- This phenomenon was first noted by Fridtjof Nansen, who recorded that ice transport appeared to occur at an angle to the wind direction during his Arctic expedition during the 1890.
- "Transport": 90° to wind, to right in northern hemisphere
- $U_{Ek}= \tau/\rho f$ (units are m^2/s, not m^3/s so technically this is not a transport; need to sum horizontally along a section to get a transport).

  $\tau = C_D \, \rho_{air} \, U^2$



Adapted from Thurman, Harold V. **Essentials of Oceanography, 5th ed.** Prentice-Hall, Inc., 1996.

- Typical size: for wind stress 0.1 N/m^2, U_{Ek}= 1 m^2/s. Integrate over width of ocean, say 5000 km, get total transport of 5 x 106 m^3/sec = 5 Sv.

Ekman layer transport

- $U_{Ek} = \tau/\rho f$ (units are m^2/s, not m^3/s so technically this is not a transport; need to sum horizontally along a section to get a transport).

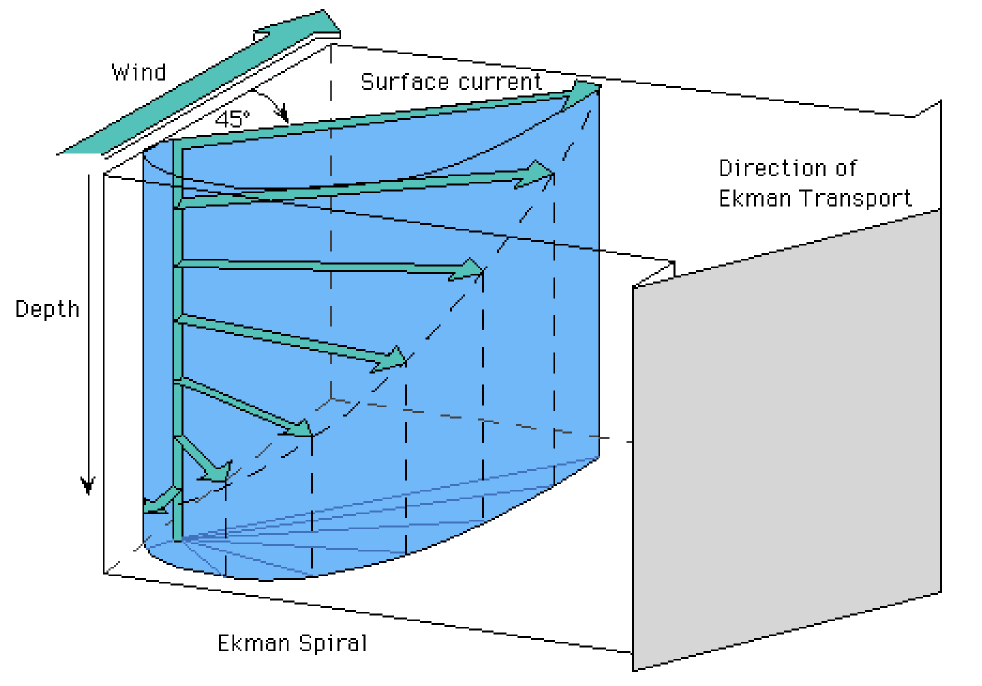
  $\tau = C_D \rho_{air} U^2$ = N/m² or Pascal (Pa)

- Estimate wind stress

  $C_{D} = 1.2 \times 10^{-3}$

  $U = 8$ m/s

  $\rho_{air} = 1.22$ kg/m³



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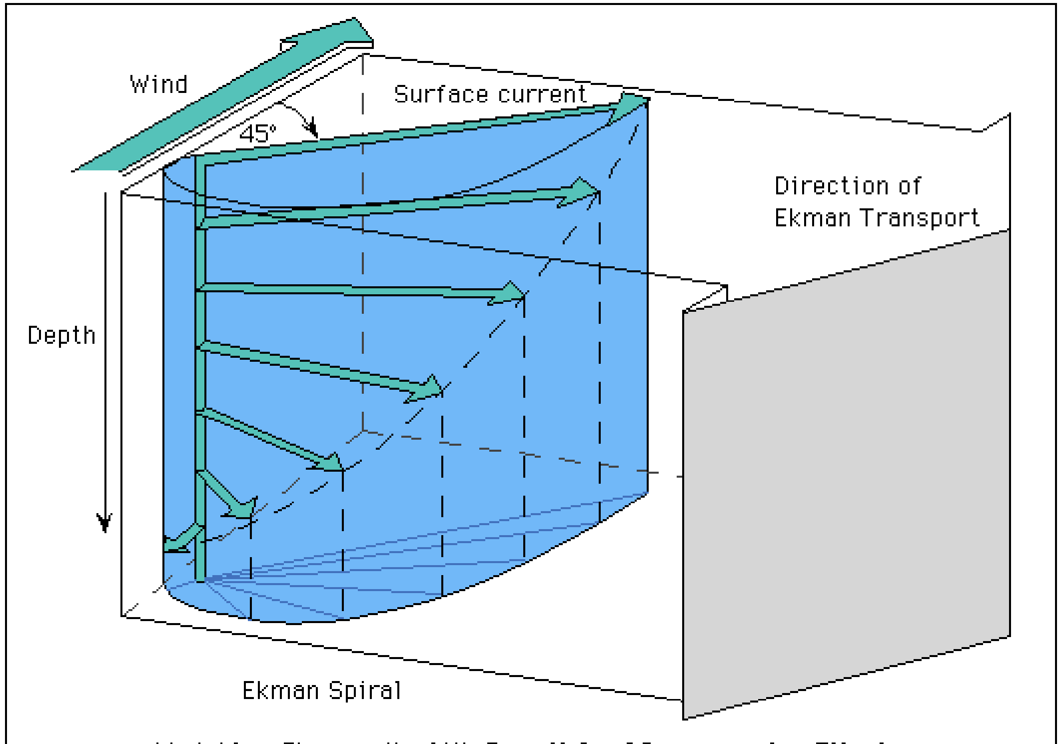
- Typical size: for wind stress 0.1 N/m², U_{Ek} = 1 m²/s.
- Integrate over width of ocean, say 5000 km, get total transport of 5 x 106 m³/sec = 5 Sv.

Ekman layer depth

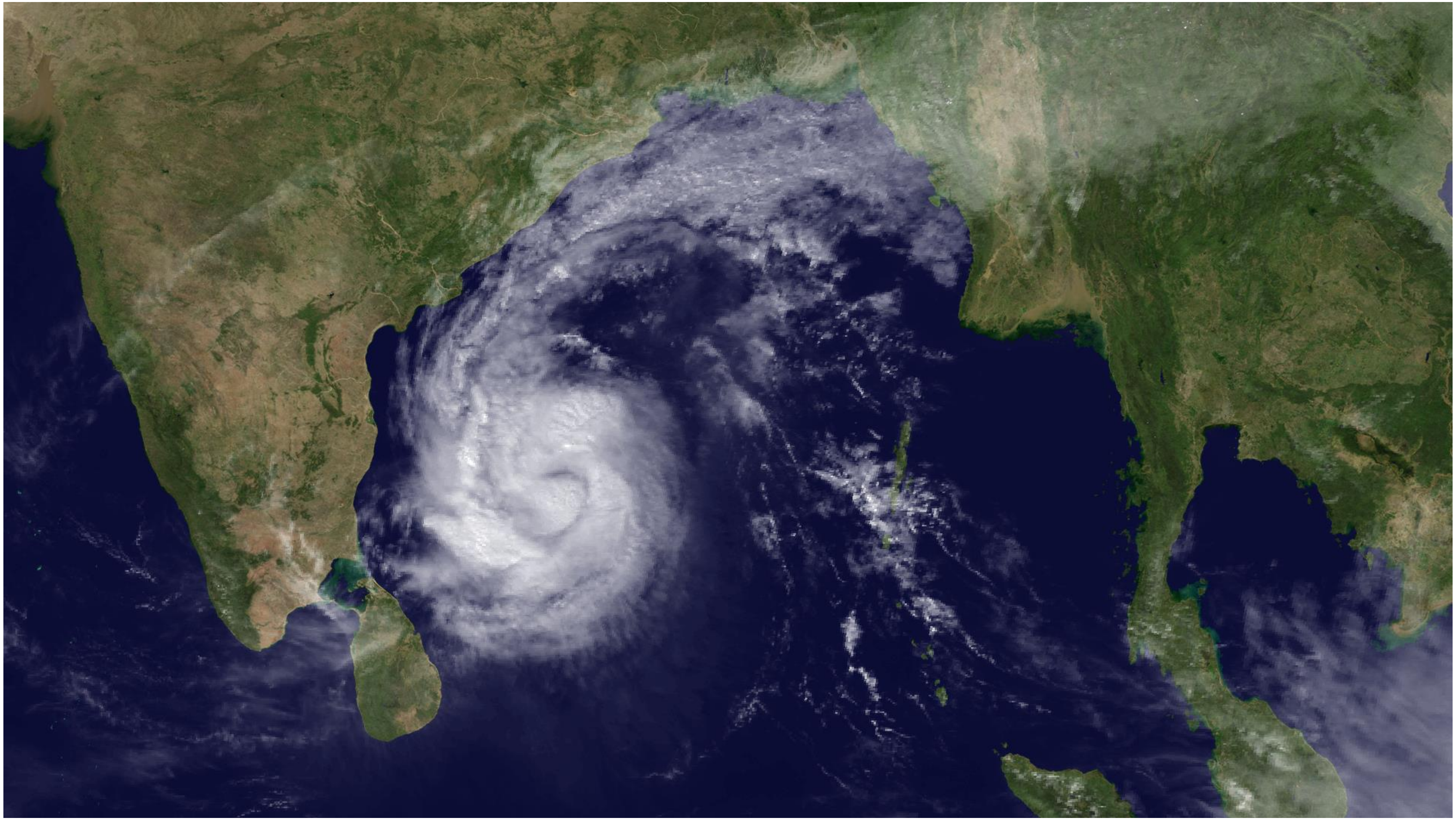
- Depth: depends on eddy viscosity A_V (why?)
  $D_{ek} = (2A_V/f)^{1/2}$



- Eddy viscosity: The turbulent transfer of momentum by eddies (circular currents of water) giving rise to an internal fluid friction.
- Eddy viscosity is about 0.05 m^2/sec in turbulent surface layer, so Ekman layer depth is 20 to 60 m for latitudes 80° to 10°.

Coriolis Effect

Why do cyclones rotate anti-clockwise?



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