An introduction to ocean-atmosphere dynamics and variability

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Topics covered in this lecture

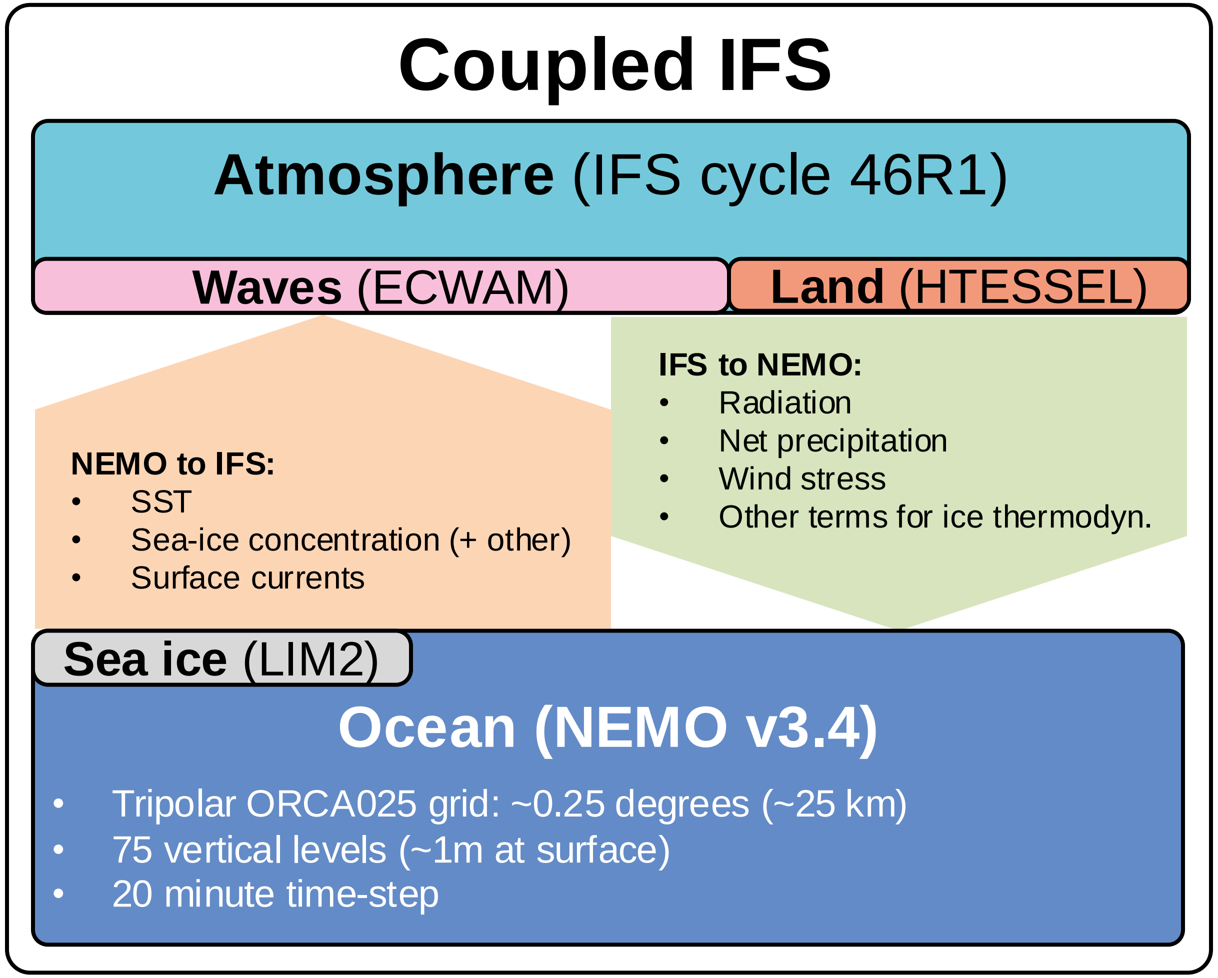
- Why couple to a dynamic ocean?
  - *Ocean coupling in the ECMWF Integrated Forecasting System (IFS)*
  - *Role of the oceans in Earth System Predictability*
  - *Timescales of ocean-atmosphere interaction*
- Oceanography fundamentals (extra slides at the end of the lecture)
  - *Features of the global ocean circulation*
  - *Governing equations and important balances*
- Examples of air-sea interaction and its scale-dependence
  - *Tropics vs mid-latitudes*
- Challenges for coupled ocean-atmosphere forecasting
  - *Horizontal resolution*
  - *Initializing the ocean state*

Things that are *not* covered in this lecture

- Discussion of the MJO – see lecture by Frederic Vitart
- Detailed discussion of ENSO – see lecture by Magdalena Balmaseda

Why use a coupled model?

The ECMWF Integrated Forecasting System (2020)



- All operational configurations of ECMWF IFS include dynamic representations of the ocean, atmosphere, sea ice, land surface, and ocean waves.
- Sub-models exchange information hourly
- The IFS single-executable coupling strategy is described by Mogensen et al. (2012), *EC Tech Memo 673*.
- IFS cycles are documented on the ECMWF website: https://www.ecmwf.int/en/publications/ifs-documentation.
- An overview of the IFS coupled model is available in Roberts et al. (2018), *Geoscientific Model Development*, 11, 3681-3712.

How does the ocean contribute to Earth System Predictability?

1) Coupling. Inclusion of a dynamic ocean allows ocean-atmosphere feedbacks.

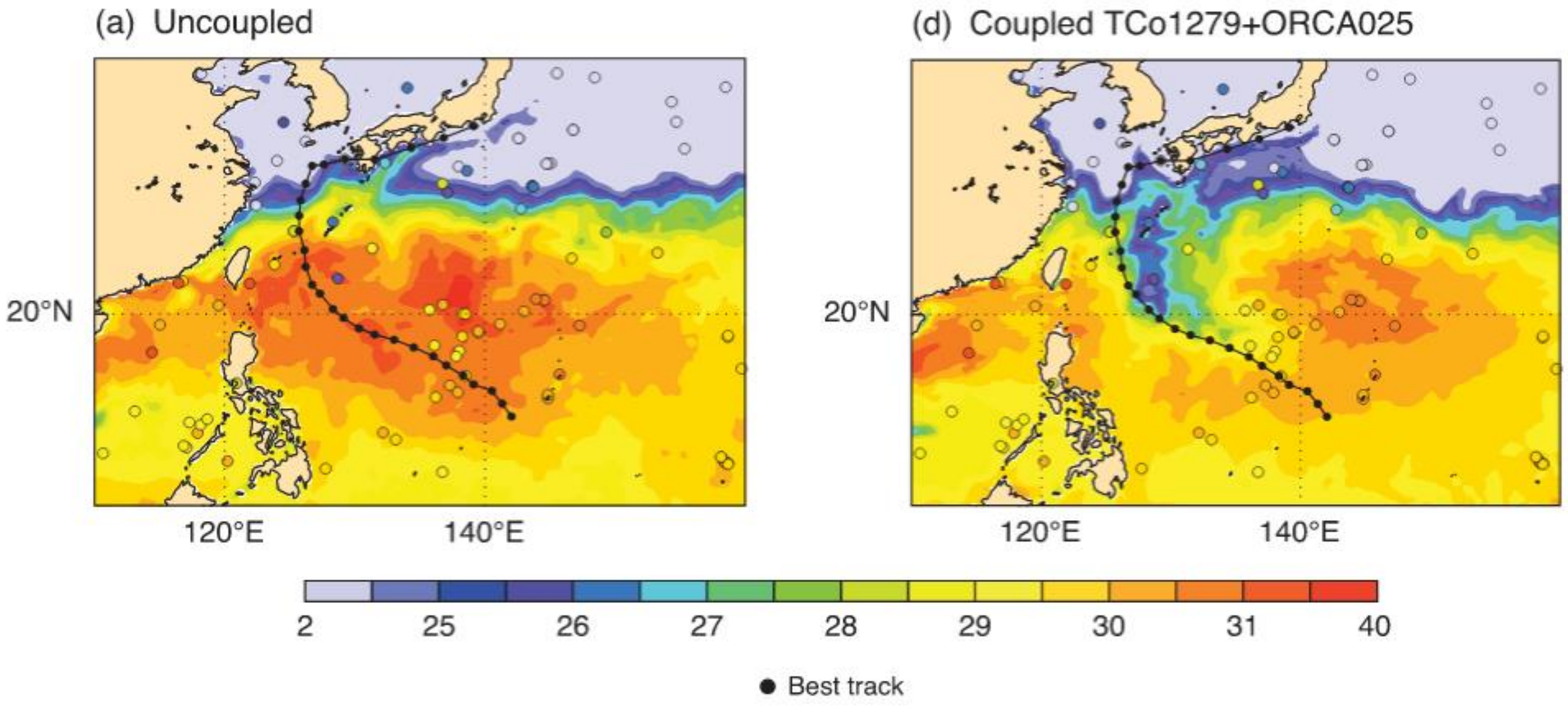


Figure 9. Five day SST forecast (shades) from 5 July 2014 and SST observations 9 July 2014 00 UTC (symbols). All SST's are in °C.

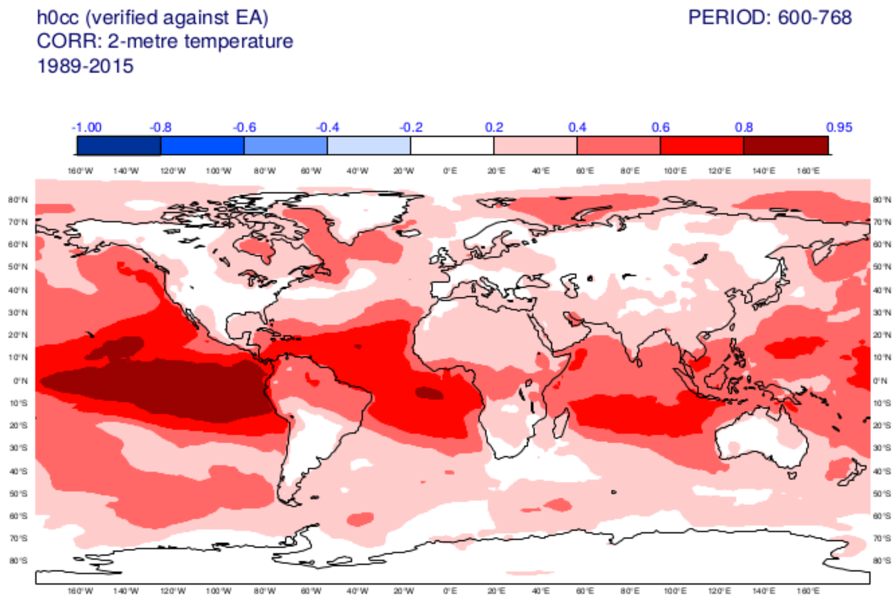
- Case study of tropical cyclone Neoguri (2014).
- 5-day SST forecasts with/without ocean coupling.
- Inclusion of a dynamic ocean allows simulation of a strong cold wake, which in turn impacts the evolution of the overlying cyclone.

Citation:

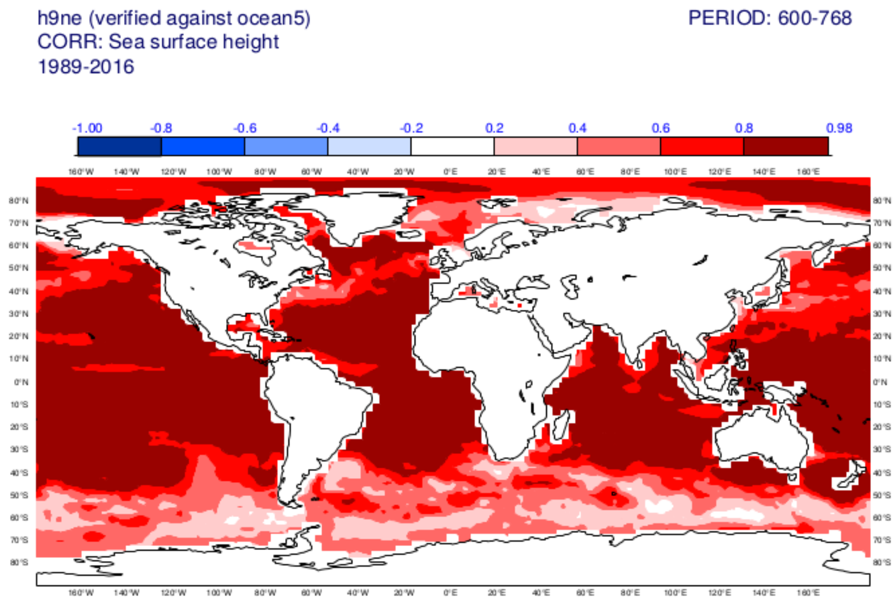
Mogensen, K. S., L. Magnusson, and J.-R. Bidlot (2017), Tropical cyclone sensitivity to ocean coupling in the ECMWF coupled model, *J. Geophys. Res. Oceans*, *122*, 4392–4412, doi:10.1002/2017JC012753.

How does the ocean contribute to Earth System Predictability?

2) The ocean has a long "memory". Relative to the atmosphere, the ocean has a large heat capacity and much slower adjustment time-scales, which become important at longer lead times.



Week 4 anomaly correlation for near-surface (2m) atmospheric temperature.



Week 4 anomaly correlation for sea surface height (largely determined by ocean heat content).

How does the ocean contribute to Earth System Predictability?

3) Predictable ocean dynamics.

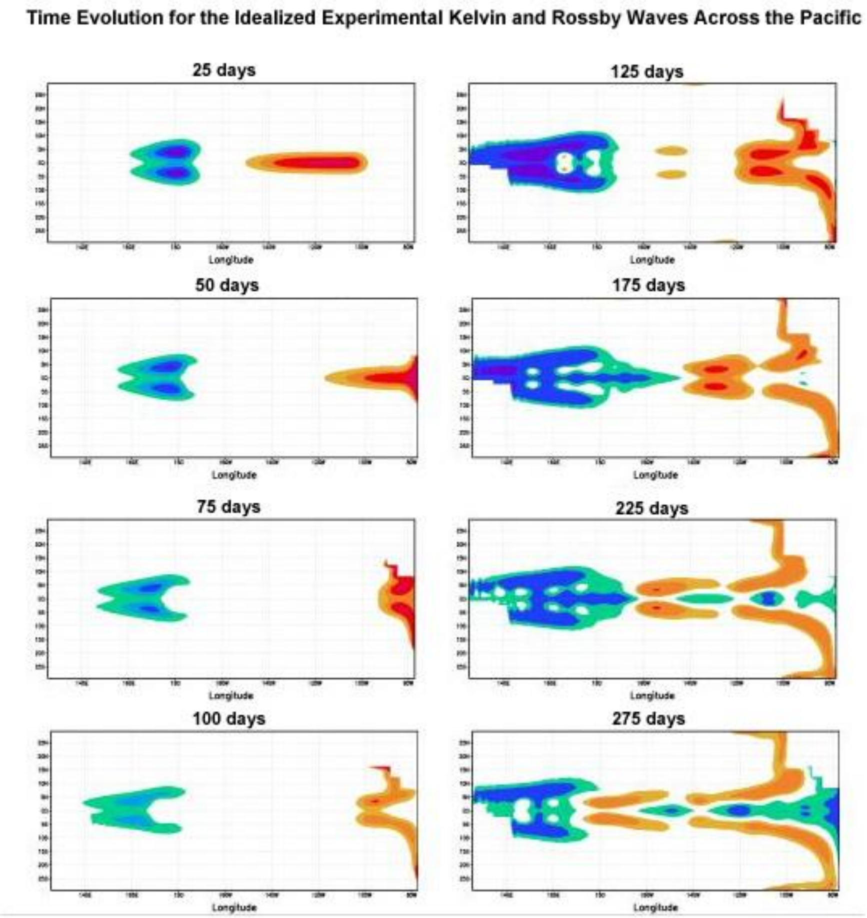


Image credit: *Introduction to Tropical Meteorology, 2nd Edition.*

Predictable ocean dynamics play an important role in modes of coupled ocean-atmosphere variability (e.g. ENSO).

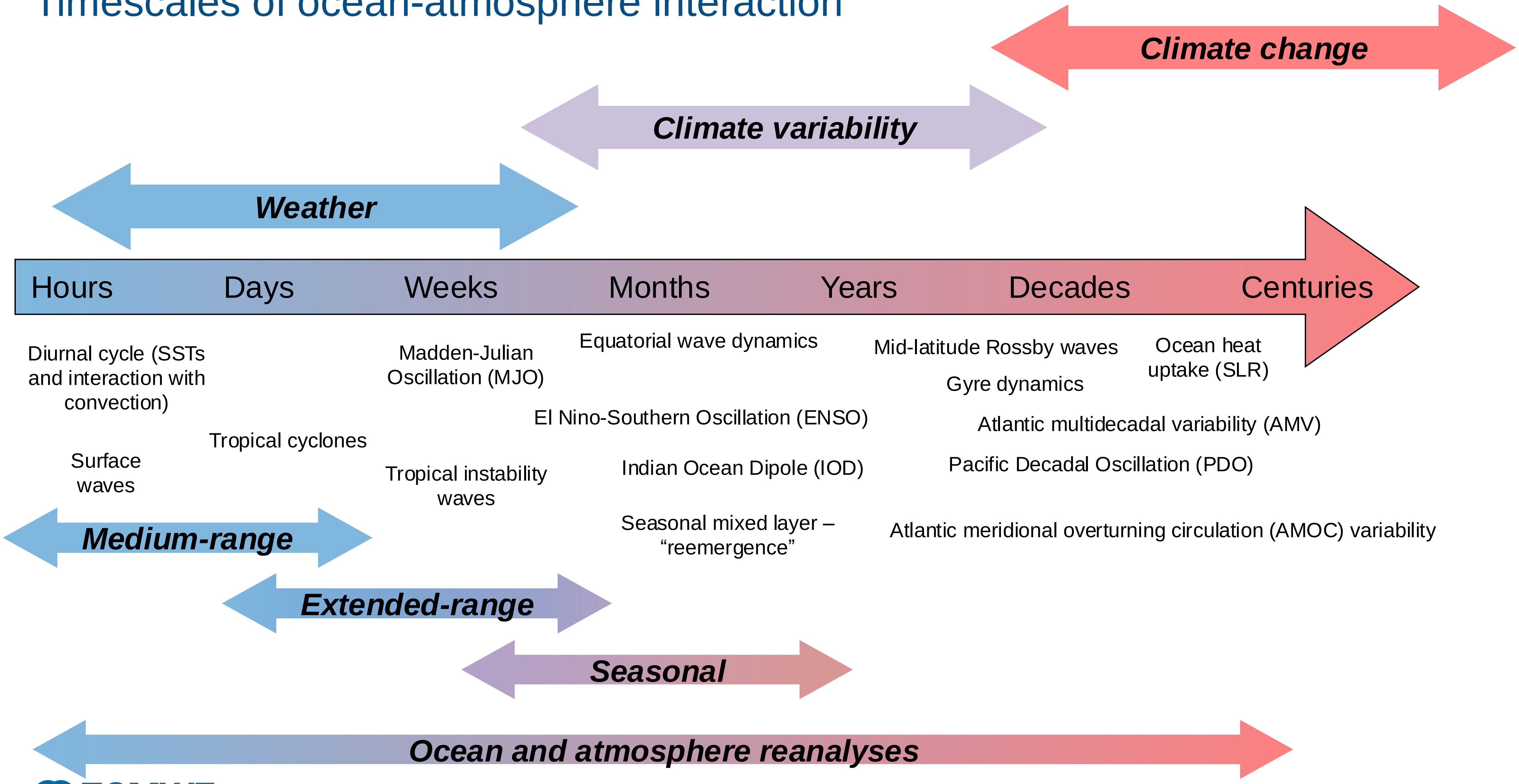
This example shows an idealized model response to a wind stress anomaly associated with El Nino conditions.

The panels show an initial eastward-moving Kelvin wave and westward propagating Rossby waves followed by reflection/propagation at the boundaries.

See “delayed oscillator theory” (Battisti and Hirst, 1989; Suarez and Schopf, 1988).

This type of large-scale equatorial wave dynamics is well-resolved by global ocean models.

Timescales of ocean-atmosphere interaction



Oceanography fundamentals (1):
Features of the global ocean circulation

Ocean vs atmosphere

In many ways the ocean and atmosphere are very similar and can be understood using the same underlying equations of motion. However, there are some key differences:

1. The oceans have coastlines, which impose boundary conditions to the flow.

2. The heat capacity of the ocean is ~1000 times larger than that of the atmosphere.

3. Ocean is strongly stratified in the vertical, and deep convection is limited to the high latitudes.

4. The ocean is forced from above by the winds and heat/freshwater fluxes. There is no real equivalent to the diabatic and radiative processes that can interact with the entire atmospheric column.

5. The main challenge of ocean modelling is the representation of small-scale dynamical processes, which may not be explicitly resolved by the model grid.

6. In contrast, a key challenge in atmospheric modelling is the representation of the complex coupling between dynamic and thermodynamic processes (e.g. interactions between the resolved flow, radiation, clouds, convection, and other parameterized processes).

The wind-driven ocean circulation

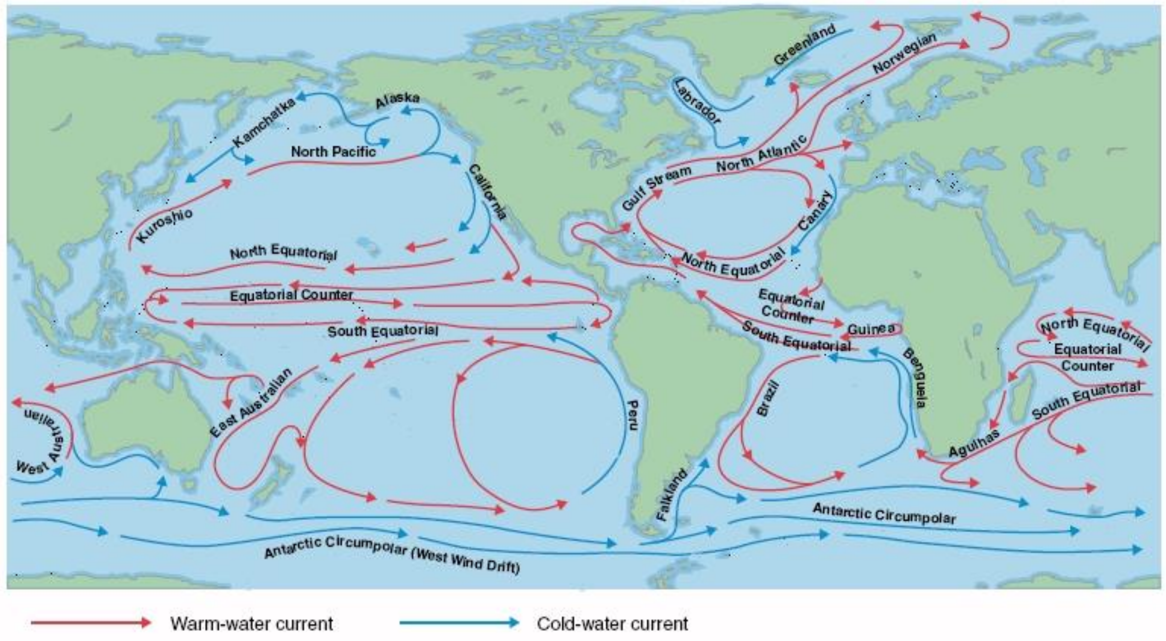
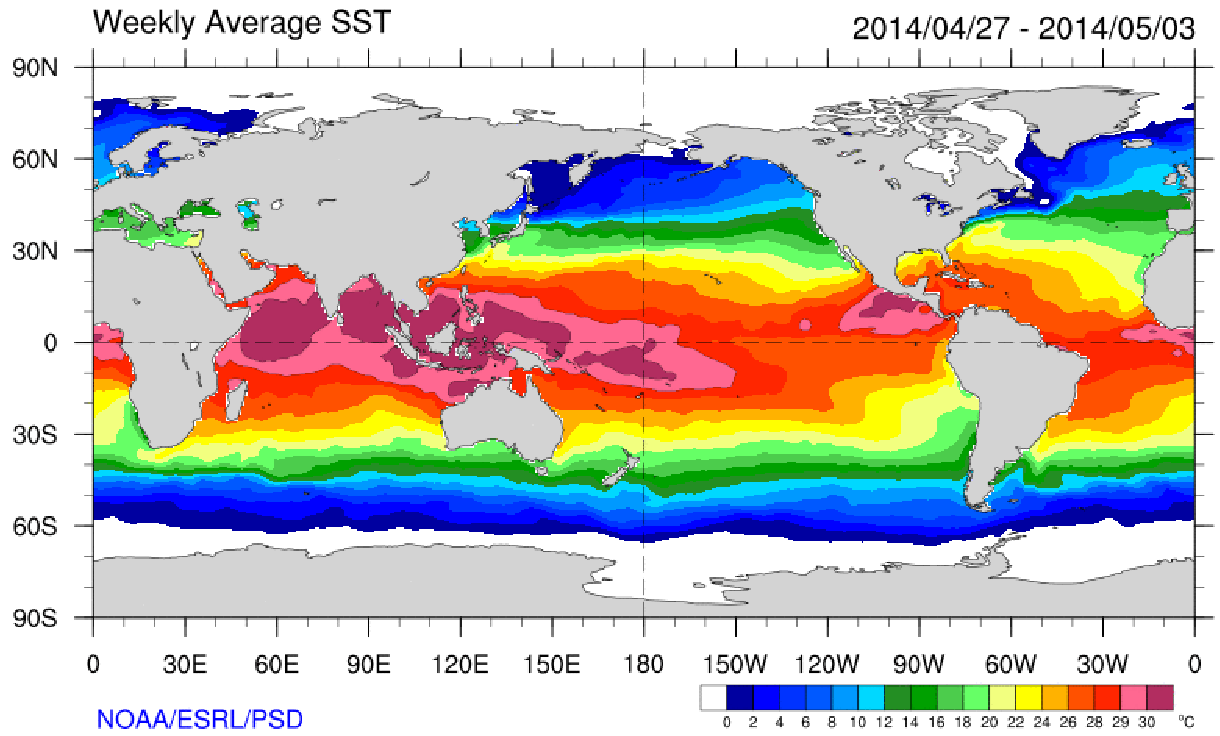


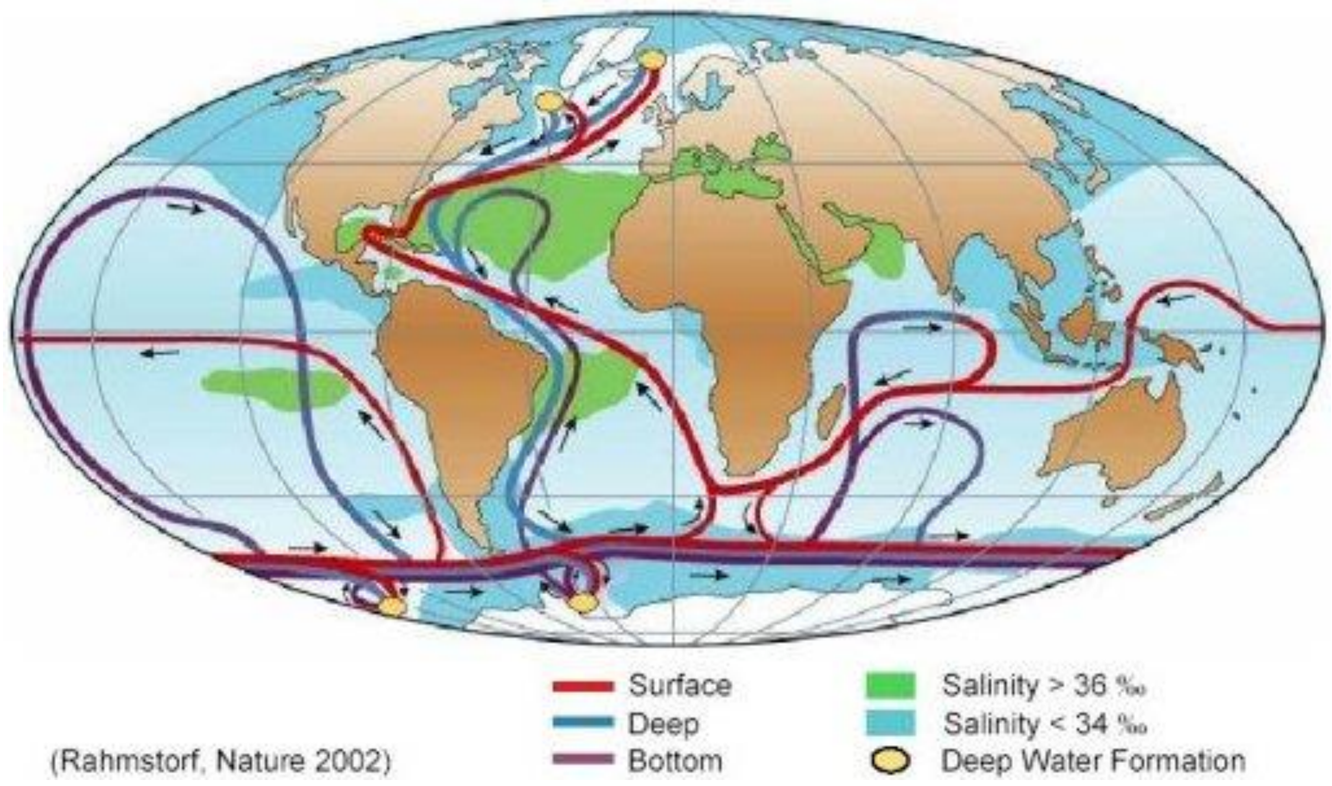
Image credit: Ocean Circulation 2nd Edition, The Open University

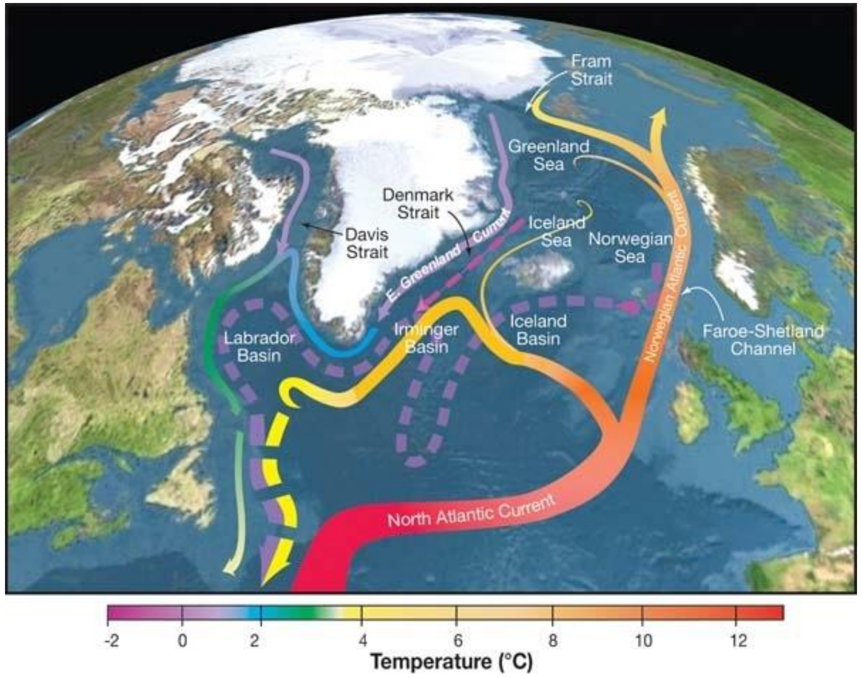


The horizontal ocean circulation is largely wind driven. It is characterized by large-scale sub-tropical and sub-polar gyres, intense western boundary currents, coastal upwelling, strong equatorial currents, and the Antarctic Circumpolar Current.

Thermohaline and overturning circulations

- The thermohaline circulation (THC) is associated with large scale (vertical) overturning circulations and is driven by fluxes of heat and fresh water.
- The Atlantic component of the THC is often referred to as the Atlantic Meridional Overturning Circulation (AMOC).





Ocean heat transports

- Overturning circulations play a particularly important role in meridional ocean heat transports: warm water transported polewards in upper layers is balanced by cold water transported equatorward in deeper layers.
- In the Atlantic, the northward heat transport by the vertical overturning circulation is much larger than that by the horizontal gyre circulation.

Estimates of the globally integrated northward heat transport

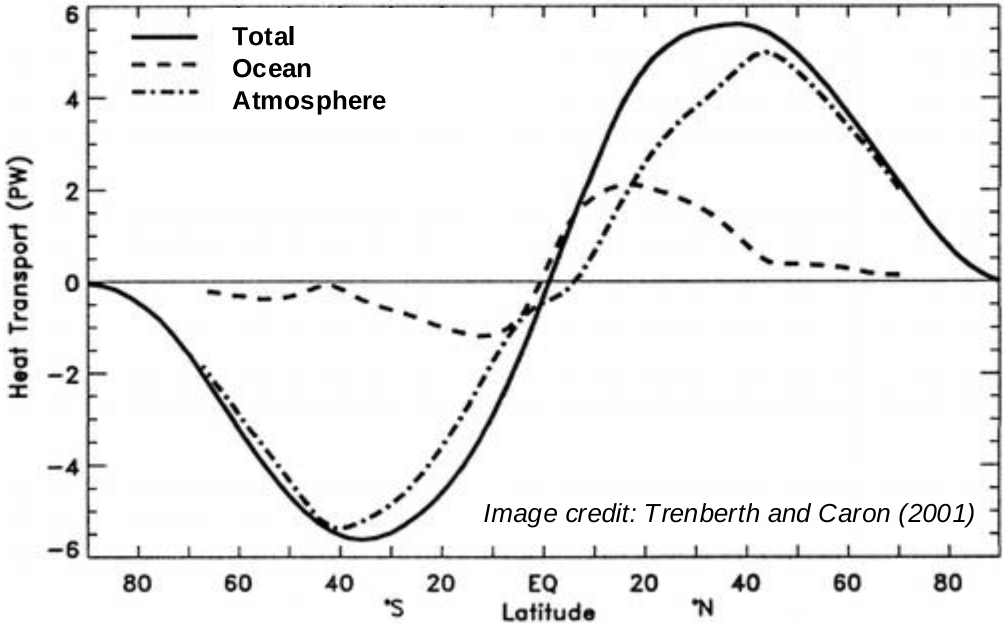


Image credit: Trenberth and Caron (2001)

Atlantic ocean heat transports at 26N

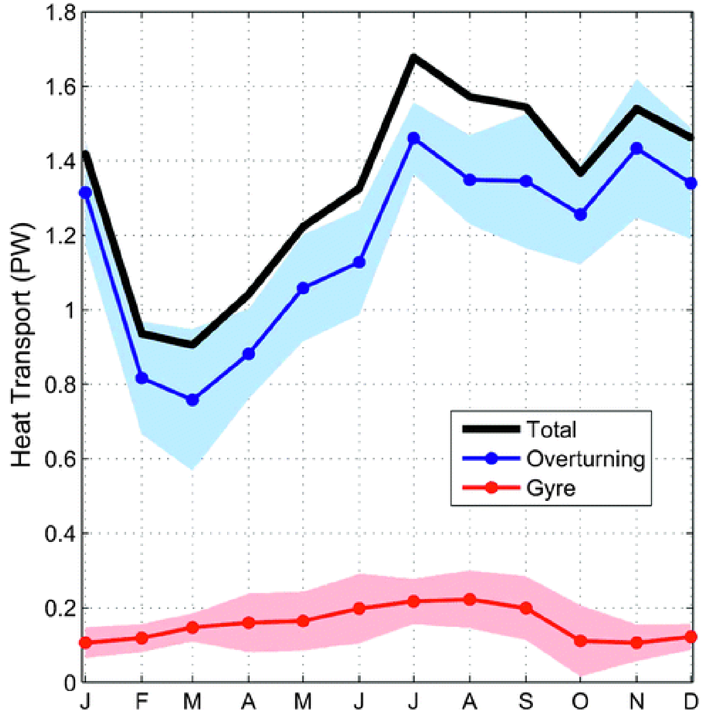


Image credit: Johns et al. (2011)

Oceanography fundamentals (2):
Governing equations and important properties

Governing equations for the ocean (+ typical approximations)

advection (non-linear) | pressure gradient

$$\boxed{\frac{\partial u}{\partial t}} + \boxed{u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z}} - \boxed{fv} = \boxed{-\frac{1}{\rho}\frac{\partial p}{\partial x}} + \boxed{F_x}$$

$$\boxed{\frac{\partial v}{\partial t}} + \boxed{u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z}} + \boxed{fu} = \boxed{-\frac{1}{\rho}\frac{\partial p}{\partial y}} + \boxed{F_y}$$

tendency | rotation/Coriolis | frictional/viscous

Momentum conservation (F=ma)

$$\frac{\partial p}{\partial z} = -\rho g$$

Vertical momentum equation (hydrostatic approximation)

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

Mass conservation (incompressible, Boussinesq approx)

$$\frac{D\Theta}{Dt} = \kappa_T \nabla^2 T \frac{1}{c_p\rho} - \frac{\partial Q_{rad}}{\partial z} \qquad \frac{DS}{Dt} = \kappa_s \nabla^2 S \qquad \rho = \rho(T, S, p)$$

Heat/salt conservation and equation of state for seawater

$\mathbf{u} = (u, v, w) =$ velocity
$\rho =$ density
$p =$ pressure
$T =$ temperature
$\Theta =$ conservative temperature
$S =$ salinity (mass ratio)
$Q_{rad} =$ radiative heat flux
$c_p =$ heat capacity of seawater
$\kappa =$ diffusivity
$g =$ effective gravity
$\Omega =$ Earth's rotation rate
$f = 2\Omega sin\theta =$ Coriolis parameter
$\frac{D}{Dt} = \left(\frac{\partial}{\partial t} + \mathbf{u}\cdot\nabla\right)$

Dimensional analysis

Zonal momentum (neglect frictional/viscous terms for now)

$$\frac{\partial u}{\partial t} + \mathbf{u} \cdot \nabla u - fv = -\frac{1}{\rho_0}\frac{\partial p}{\partial x}$$

Zonal momentum (dimensionless)

$$\left(\frac{1}{f_0 T}\right)\frac{\partial u'}{\partial t'} + \left(\frac{U}{f_0 L}\right)\mathbf{u}' \cdot \nabla u' - \left(\frac{f}{f_0}\right)v' = -\left(\frac{P}{\rho_0 f_0 L U}\right)\frac{\partial p'}{\partial x'}$$

$$\left(\frac{1}{f_0 T}\right) = \frac{\text{inertial timescale}}{\text{timescale of interest}} << 1$$

$$\left(\frac{U}{f_0 L}\right) = \frac{\text{inertial force}}{\text{coriolis force}} = \text{Rossby number (Ro)} << 1$$

$$\boxed{\left(\frac{f}{f_0}\right) = O(1) \qquad \left(\frac{P}{\rho_0 f_0 L U}\right) = O(1)}$$

Dominant balance is between pressure gradient and Coriolis terms

Typical values for basin-scale flow

$$P = \rho_0 g = 10^4 \text{ Pa (pressure for 1 m sea surface height)}$$
$$W = UD/L = 1 \times 10^{-5} \text{ m/s (vertical velocity)}$$
$$\rho_0 = 1000 \text{ kg/m}^3 \text{(reference density)}$$
$$L = 5 \times 10^6 \text{ m (horizontal length)}$$
$$U = 5 \times 10^{-2} \text{ m/s (horizontal velocity)}$$
$$D = 1000 \text{ m (vertical length)}$$
$$T = \text{timescale of interest, s}$$
$$f_0 = 10^{-4} \text{ s}^{-1}$$

Dimensionless variables

$$x' = x/L \qquad v' = v/U$$
$$y' = y/L \qquad w' = w/W$$
$$z' = z/D \qquad t' = t/T$$
$$u' = u/U \qquad p' = p/P$$

Geostrophic balance

Balance between pressure gradient and Coriolis terms:

$$(u_g, v_g) = \left(-\frac{1}{f\rho}\frac{\partial p}{\partial y}, \frac{1}{f\rho}\frac{\partial p}{\partial x}\right)$$

Combining geostrophic and hydrostatic balance* gives thermal wind balance:

$$\left(\frac{\partial u_g}{\partial z}, \frac{\partial v_g}{\partial z}\right) = \frac{g}{f\rho_0}\left(\frac{\partial \rho}{\partial y}, -\frac{\partial \rho}{\partial x}\right)$$

- Horizontal gradients in density (easy to observe) determine vertical shear of currents (hard to observe).
- Geostrophic assumptions hold over much of the open ocean and many ocean observing systems rely on thermal wind (e.g. RAPID array at 26N).
- Note that if density is a function of pressure only (i.e. barotropic), the geostrophic velocity is independent of depth. In this case geostrophic currents at all depths are determined by gradients in sea surface height.

Assumptions:
- Not on the equator (i.e. f >> 0).
- Inertial terms are negligible (i.e. Ro << 1).
- Away from boundary layers (i.e. friction is negligible).
- Timescale of interest >> (1/f).

Surface geostrophic currents follow sea-surface height contours.

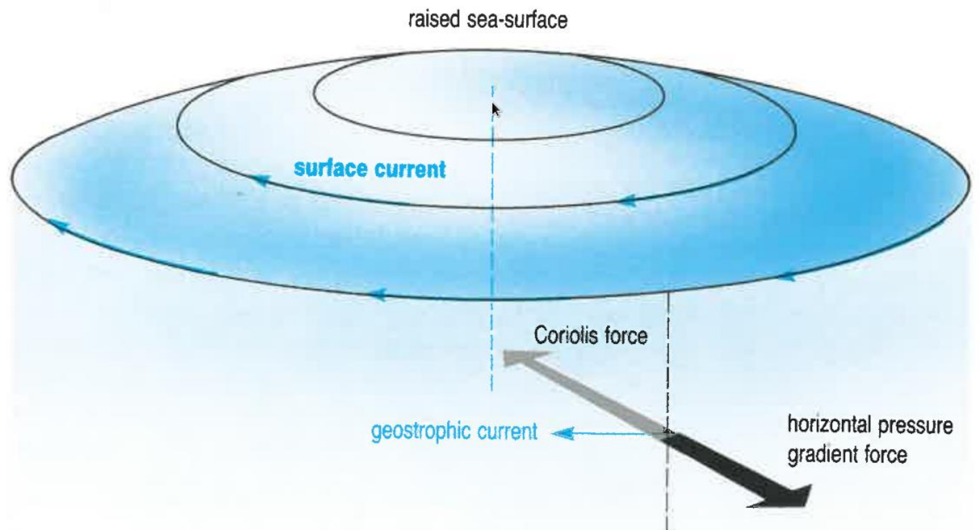


Image credit: Ocean Circulation 2nd Edition, The Open University

**note use of the Boussinesq approximation: i.e. density variations are ignored except where they appear in terms multiplied by g.*

Internal wave fundamentals

- Climatically relevant ocean waves propagate as perturbations to the depth of the ocean thermocline.
- These phenomena can be modelled using the “reduced gravity” shallow water equations.
- The density difference between layers is typically much less than the reference density, so g’ ~ g / 300.
- 10 cm displacement of sea surface height associated with 30 m displacement of the thermocline.
- Vertical density structure can (to some extent) be inferred from satellite observations of sea surface height

Wave basics

$$\text{phase angle}: \theta = kx + ly + mz - \omega t$$

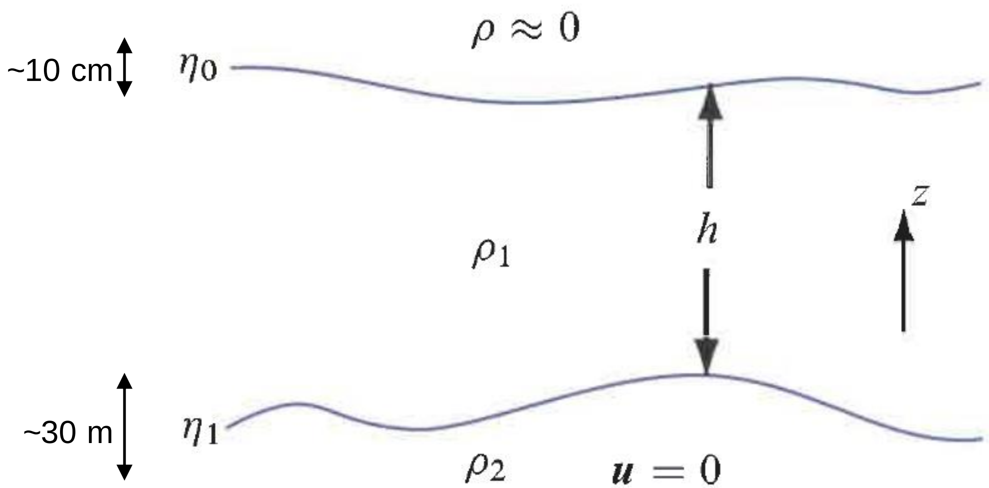
$$\text{wavenumber}: \mathbf{k} \equiv (k, l, m) = \nabla\theta$$

$$\text{frequency}: \omega = -\frac{\partial\theta}{\partial t}$$

$$\text{Phase speed}: c_p = \frac{\omega}{|\mathbf{k}|}$$

$$\text{Group speed}: c_g = \frac{\partial\omega}{\partial\mathbf{k}} \equiv \left(\frac{\partial\omega}{\partial k}, \frac{\partial\omega}{\partial l}, \frac{\partial\omega}{\partial m}\right)$$

Image credit: Atmospheric and Oceanic Fluid Dynamics 2nd Edition, Vallis



$$\boxed{reduced\ gravity = \ g' = g\frac{(\rho_2 - \rho_1)}{\rho_1}}$$

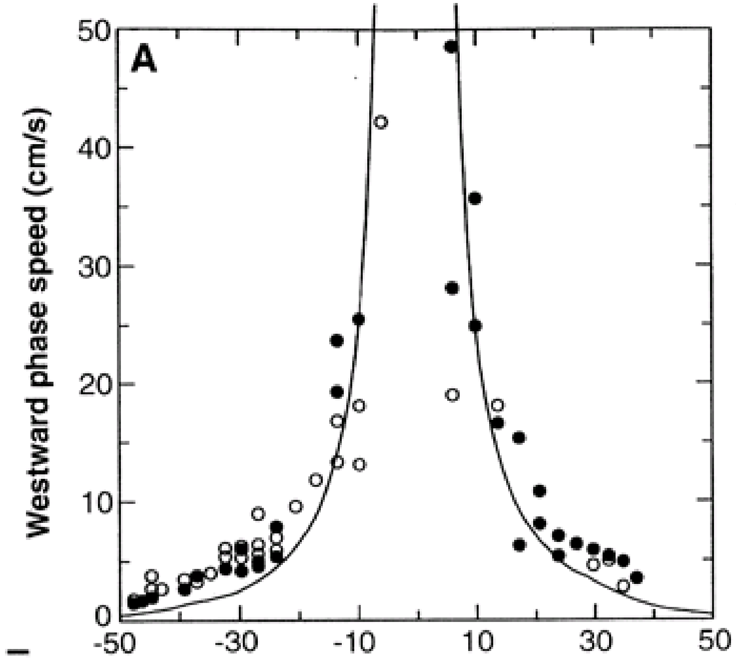
$$\boxed{L_d = \frac{\sqrt{g'H}}{f}}$$

ECMWF

Ocean Rossby waves are (relatively) slow

- Rossby waves are a consequence of latitudinal variations in planetary vorticity (i.e. *f* increases polewards).
- They are solutions to the linearized shallow water equations using the quasi-geostrophic approximation.
- Phase velocity is always westwards relative to the zonal mean flow.
- Group velocity (energy propagation) can be eastward or westward, depending on the zonal vs meridional scale of waves.
- Waves travel faster closer to the equator and longer waves travel faster than shorter waves (dispersive).

Rossby wave phase speed as a function of latitude in the ocean (Chelton et al. 1996)



Rossby wave dispersion relation for reduced gravity model (no background flow)

$$\omega = -\frac{\beta k}{k^2 + l^2 + \frac{1}{L_d^2}} \qquad c_p^x = -\frac{\beta}{k^2 + l^2 + \frac{1}{L_d^2}}$$

~10 years to cross the Atlantic at 40N!

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Also see extra slide 7

Ocean Kelvin waves propagate along boundaries

- Wave solutions to the linearized shallow water equations at a boundary.
- Waves are thus confined to coastlines and the equatorial waveguide.
- Propagate with a boundary to the right (left) in the northern (southern) hemisphere and travel eastward along the equator.
- Kelvin waves are non-dispersive ($c_p = c_g = \sqrt{g'H}$)
- It takes about 2 months for the first baroclinic Kelvin wave to cross the equatorial Pacific (c ~ 3 m/s).
- Equatorial waves triggered by changes to the wind field are an important component of ENSO dynamics.

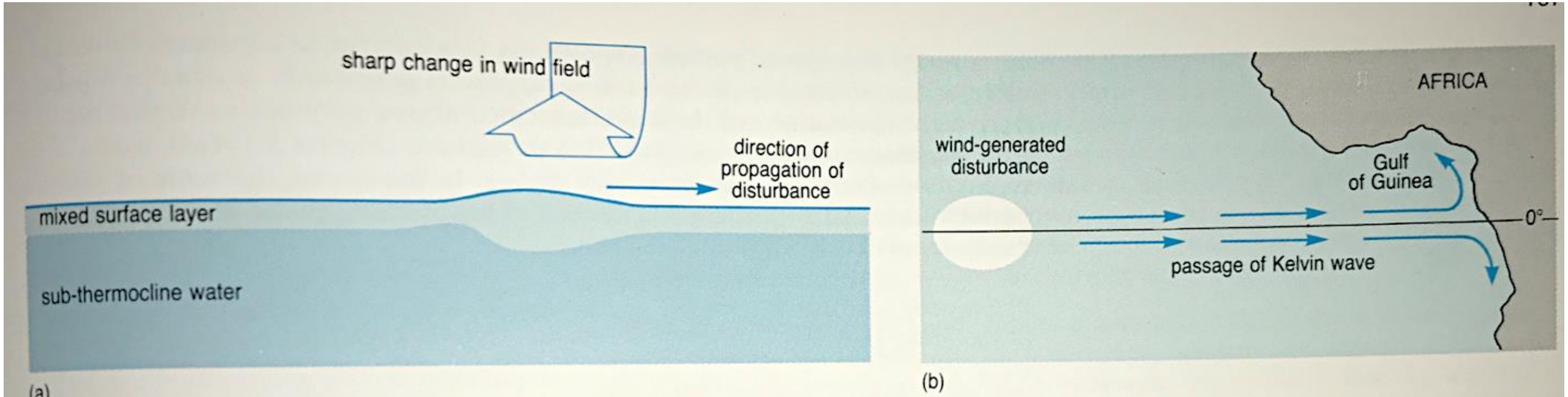


Image credit: Ocean Circulation 2nd Edition, The Open University

Instability in the ocean

- Flow in the ocean and atmosphere is not generally stable: small perturbations give rise to growing unstable modes.
- **Baroclinic instability:**
  - Responsible for atmospheric weather systems and mesoscale ocean eddies.
  - Requires horizontal density gradients.
  - Available potential energy converted to kinetic energy.
  - Occurs preferentially at horizontal scales close to the Rossby radius of deformation (L_d).
- **Barotropic/shear instability:**
  - Requires shear in background flow (e.g. jets).
  - Mean-flow kinetic energy converted to eddy kinetic energy.

Simulated Gulf Stream rings

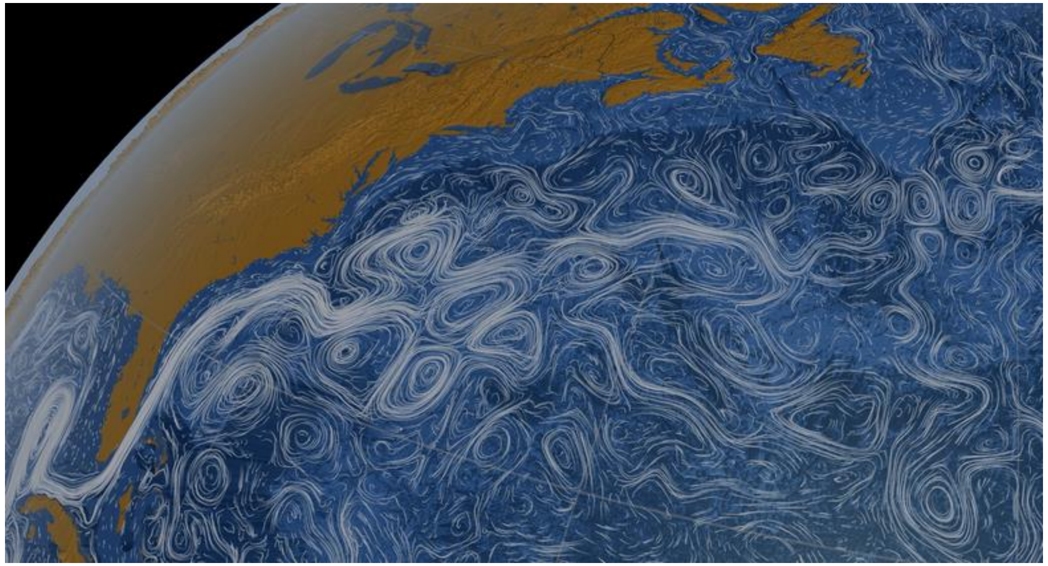


Image credit: NASA

Tropical instability waves related to shear in equatorial currents

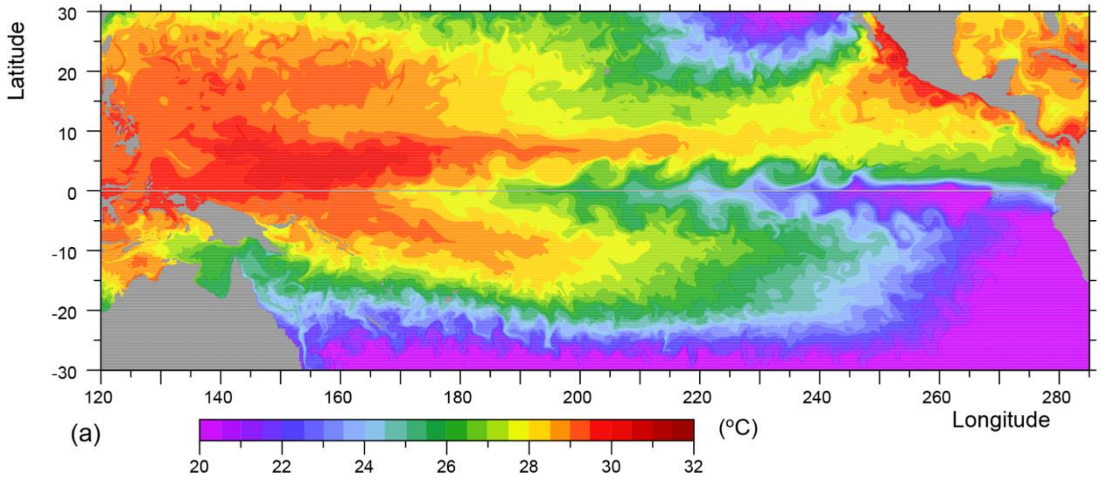


Image credit: Webb (2018), Ocean Science, vol 14.

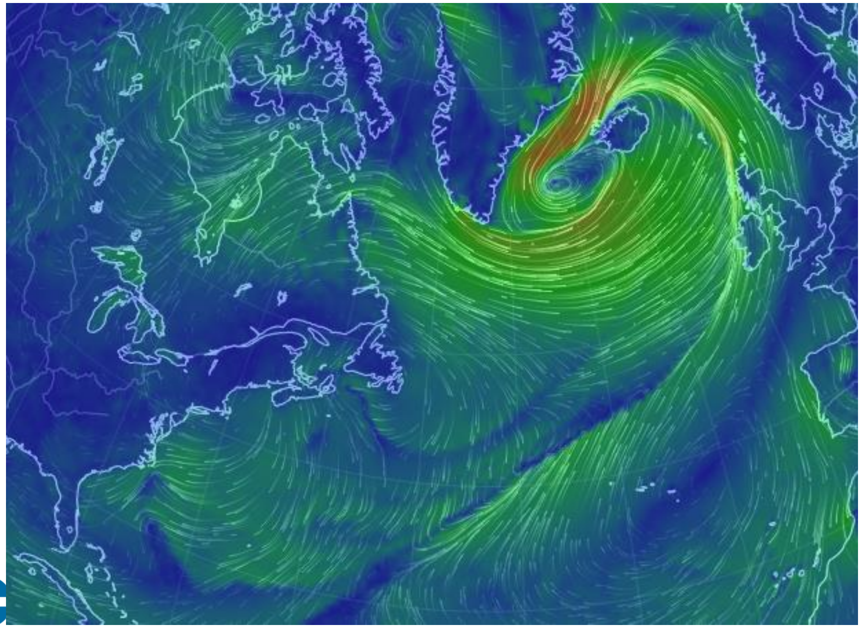
Ocean and atmosphere have different spatial scales

- The Rossby radius of deformation (L_d) controls the typical length scale for ocean and atmosphere "weather".
- Length scale at which rotational and stratification effects (i.e. vorticity advection and vortex stretching) are comparable.

$$\boxed{L_d = \frac{\textit{shallow water gravity wave speed}}{\textit{planetary vorticity}} = \frac{\sqrt{g'H}}{f}}$$

Atmosphere: L_d ~1000 km, period ~4 days

Snapshot of surface winds



Ocean: L_d ~10-200 km, period ~40 days

Snapshot of surface currents

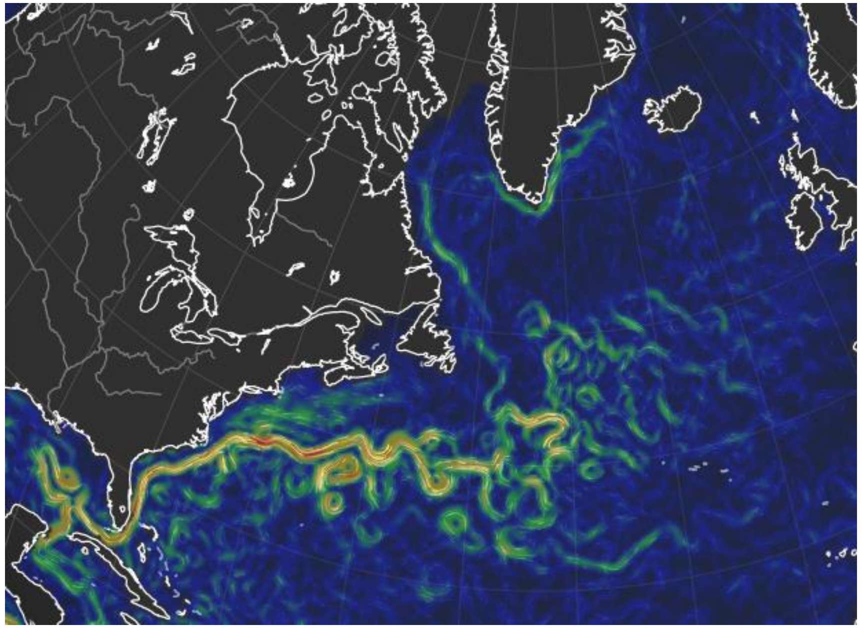


Image credit: earth.nullschool.net (forecast for 2020-03-16)

Examples of air-sea interaction

Air-sea interaction in the tropics is strongly coupled

- Sea-surface temperature (SST) variability in the tropics can have global consequences.
- For example, the turbulent heat fluxes associated with tropical SST variability can trigger deep convection and positive feedbacks that modify the large-scale atmospheric circulation.
- The resulting upper troposphere circulation anomalies can provide a source of planetary Rossby waves, which propagate from the tropics to the extratropics and manifest as global teleconnection patterns (see ENSO and MJO talks).

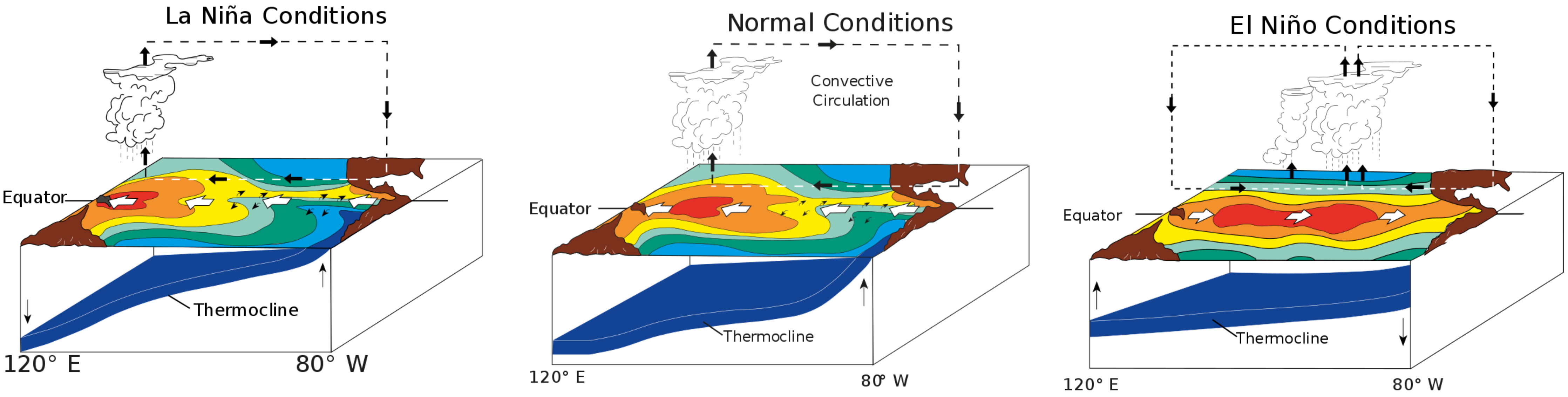


Image credit: Wikipedia/NOAA/PMEL/TAO

Ocean and atmosphere fronts modulate mid-latitude air-sea interaction

SST gradients associated with ocean fronts can influence the mean state and variability of theoveryling atmosphere.

For example, they provide a strong constraint on atmosphere baroclinicity and thus influence the strength and position of the atmospheric storm tracks.

However, atmospheric fronts are also essential for communicating the impact of intense air-sea interactions into the upper troposphere (e.g. within the warm sector of extratropical cyclones).

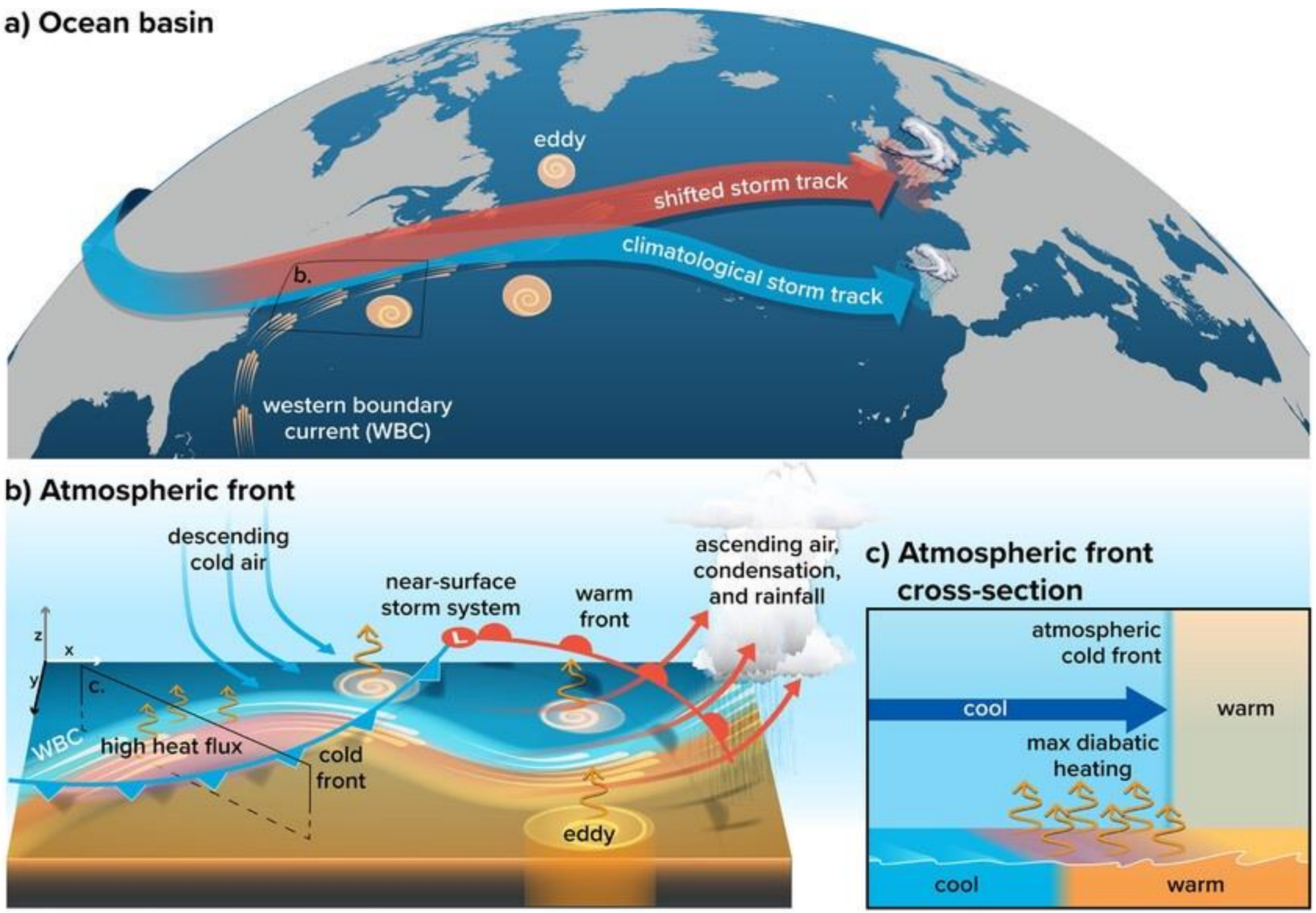
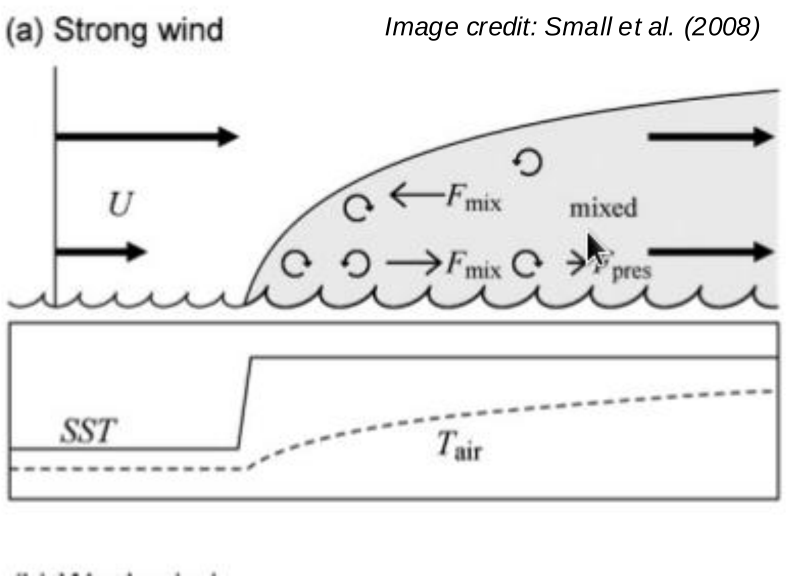


Image credit: Seo et al. (2023)

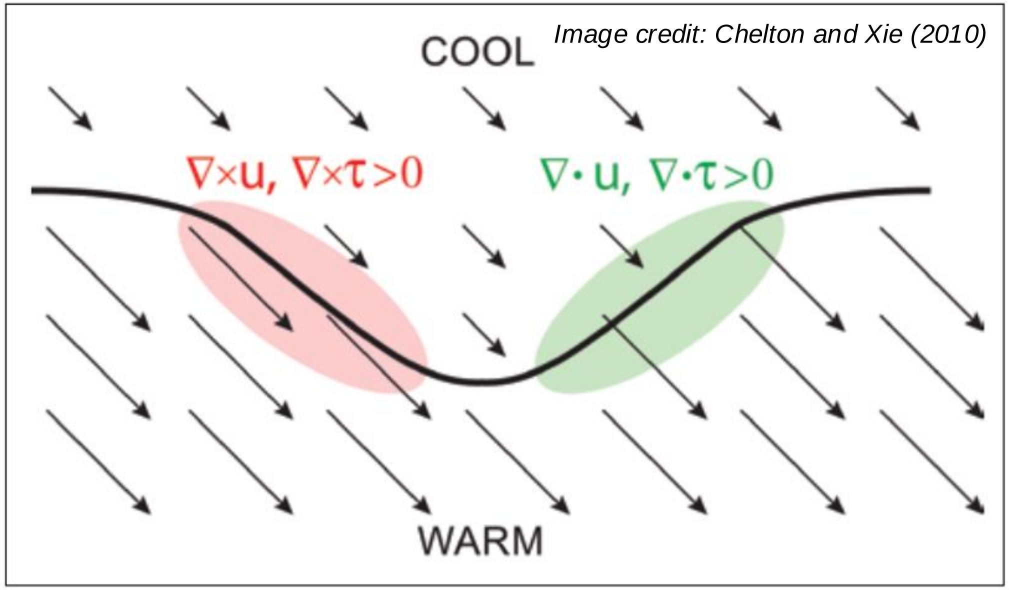
Mid-latitude air-sea interaction is also scale-dependent

At scale basin of ocean basins, there is negative or zero correlation between SST and wind speed/heat fluxes out of the ocean. ***Stronger winds*** → ***ocean heat loss*** → ***SST cooling*** → ***atmosphere driving ocean.***

At scale of ocean fronts and eddies, there is positive correlation between SST and wind speed/heat fluxes out of the ocean: ***Warmer SSTs*** → ***ocean heat loss*** → ***stronger winds*** → ***ocean driving atmosphere.***



A case where air flowing over an SST front experiences enhanced mixing and downward transfer of momentum.



Relative orientation of prevailing winds and SST gradients can impact wind stress curl and dirvergence.

Forecast challenges

Forecast challenges: ocean model resolution

- It is necessary to resolve the Rossby radius of deformation (L_d) in order to accurately simulate sharp ocean fronts and mesoscale eddies.
- The operational configuration of the ECMWF ocean model has a resolution of 1/4th degree (~25 km): eddies are "permitted" in the lower latitudes, but cannot be resolved in the mid-/high-latitudes or coastal regions.
- "Eddy resolving" configurations are very expensive (e.g. 1/12th degree costs 30 x more than 1/4th degree).
- This limits our ability to correctly simulate the position of the Gulf Stream and aspects of air-sea interaction.

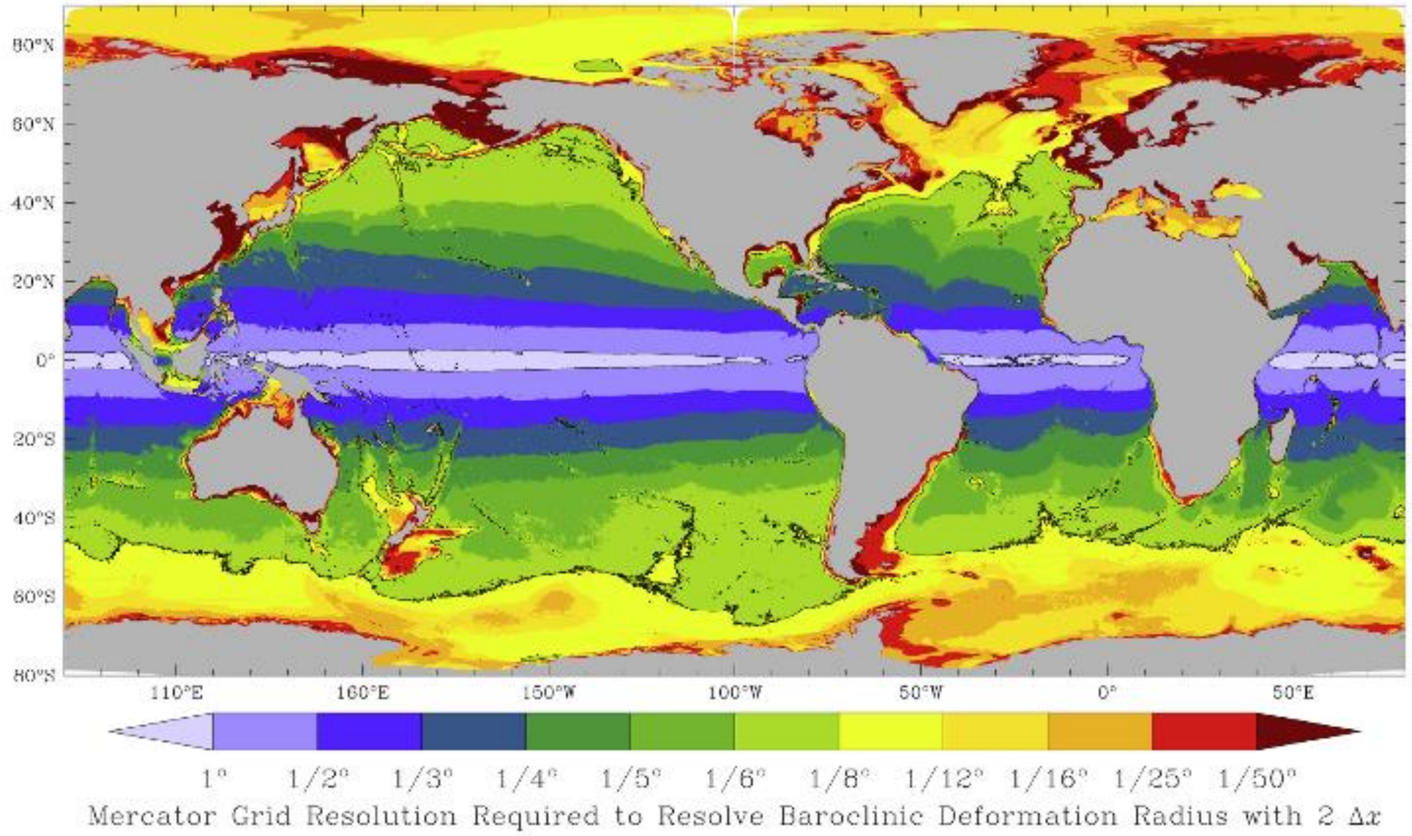


Image credit: Hallberg (2013)

$$L_d = \frac{\sqrt{g'H}}{f}$$

Forecast challenges: ocean model resolution

- Impact of ocean model resolution on the representation of the Gulf Stream

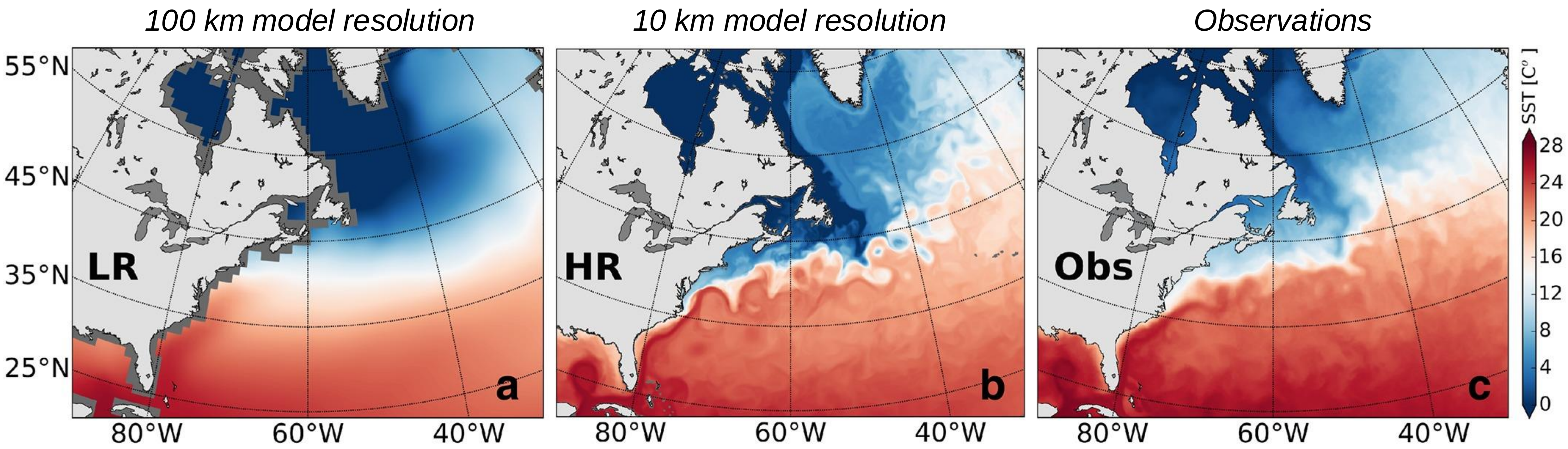
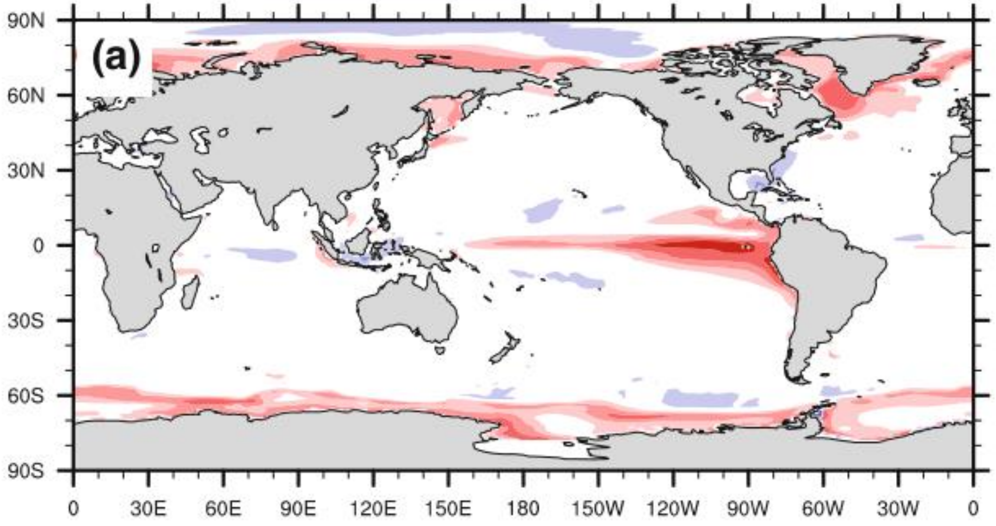


Image credit: Siqueira and Kirtman (2016)

Forecast challenges: air-sea interaction

- Higher resolution ocean in coupled climate model simulates more intense air-sea interaction in regions of high eddy activity, such as western boundary currents and the Antarctic Circumpolar Current.
- Simultaneous correlations between monthly mean SST and heat flux out of the ocean. **Positive (red)** values are indicative of the ocean driving an atmospheric response.

100 km model resolution



10 km model resolution

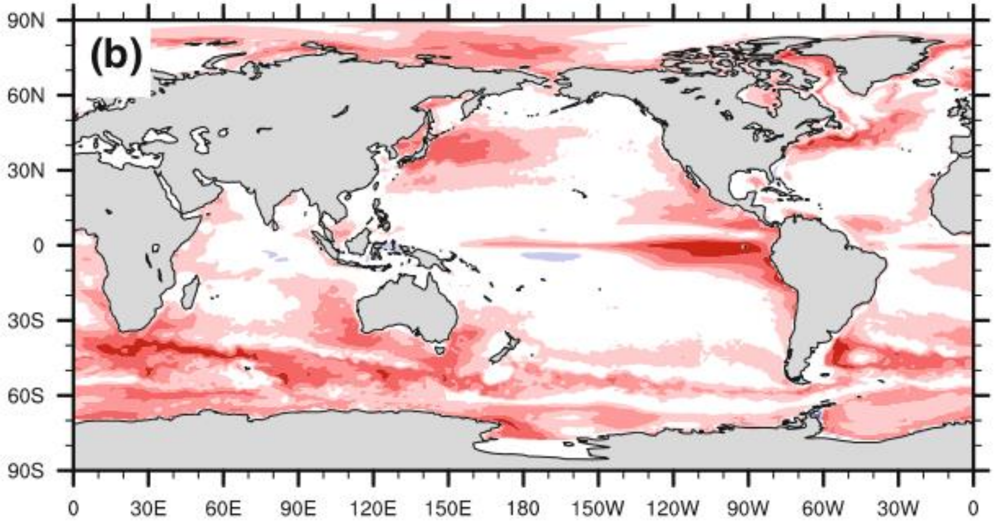
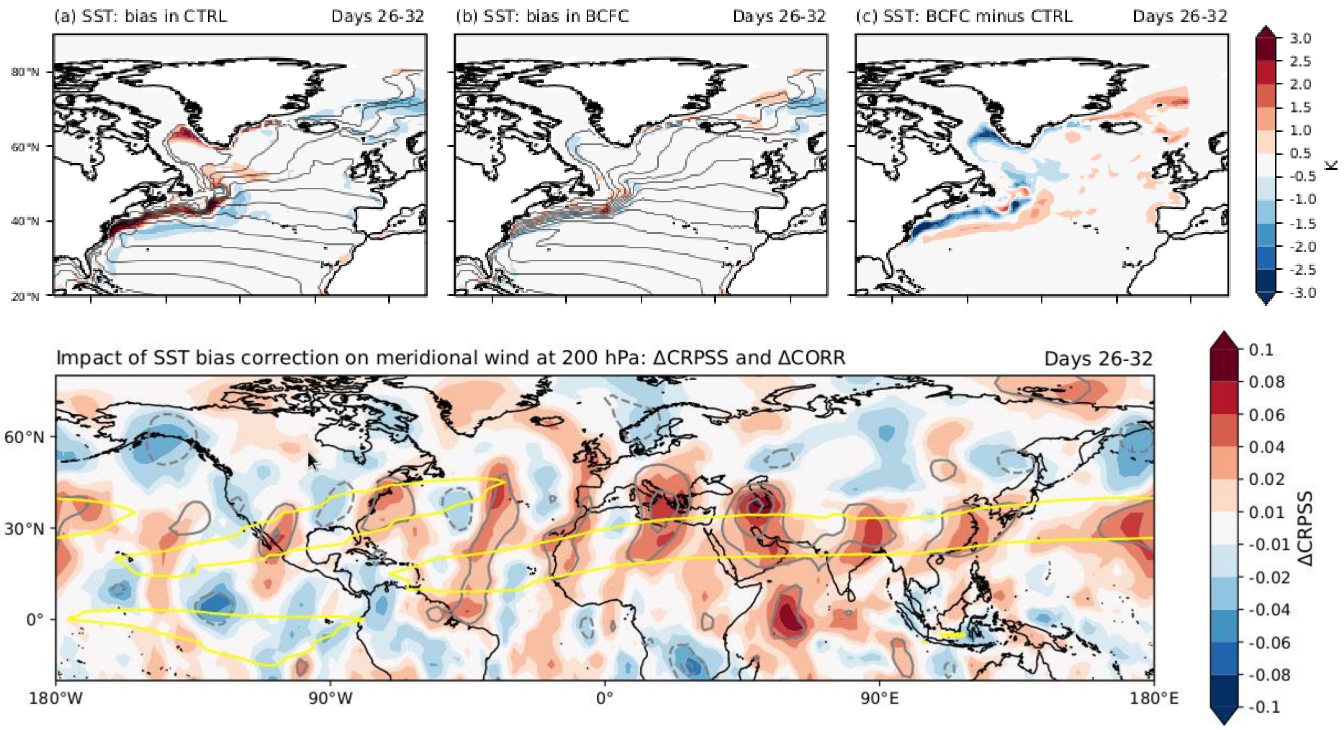


Image credit: Kirtman et al. (2012)

Forecast challenges: impact of Gulf Stream errors on ECMWF forecasts



(a-b) Climatological SSTs seen by the atmosphere (contour spacing 2 K) and biases relative to ESA CCI (c) Difference between BCFC and CTRL SST climatologies (shading). (d-e) Impact of North Atlantic SST bias correction on the forecast skill of weekly mean anomalies in terms of CRPSS (shading) and anomaly correlation (grey contour spacing of 0.1) for meridional wind at 200 hPa (v200). The yellow contours in (e) highlight the position of the northern hemisphere waveguide diagnosed from the meridional gradient of absolute vorticity.

Roberts et al. (2021), https://doi.org/10.1029/2020GL091446

- The position of the Gulf stream influences the location, magnitude, and timing of upward motion and convection associated with cyclones propagating in the North Atlantic storm track (Minobe et al., 2008).
- However, ocean models with a grid spacing of about 25 km, as used in the IFS, struggle to accurately simulate the location and structure of the Gulf Stream.
- These errors can impact weather forecasts, with signals propagating out of the Atlantic along the northern hemisphere waveguide.
- Higher-resolution ocean models that can better resolve the position of the Gulf Stream (grid spacing < 10 km) should improve future ECMWF forecasts.

Forecast challenges: ocean model initialization

- Accurate forecasts require accurate initial conditions.
- Full three-dimensional ocean state cannot be inferred from satellite data.
- Reliant on in situ observations from moorings, ship data, and profiling floats (e.g. Argo).

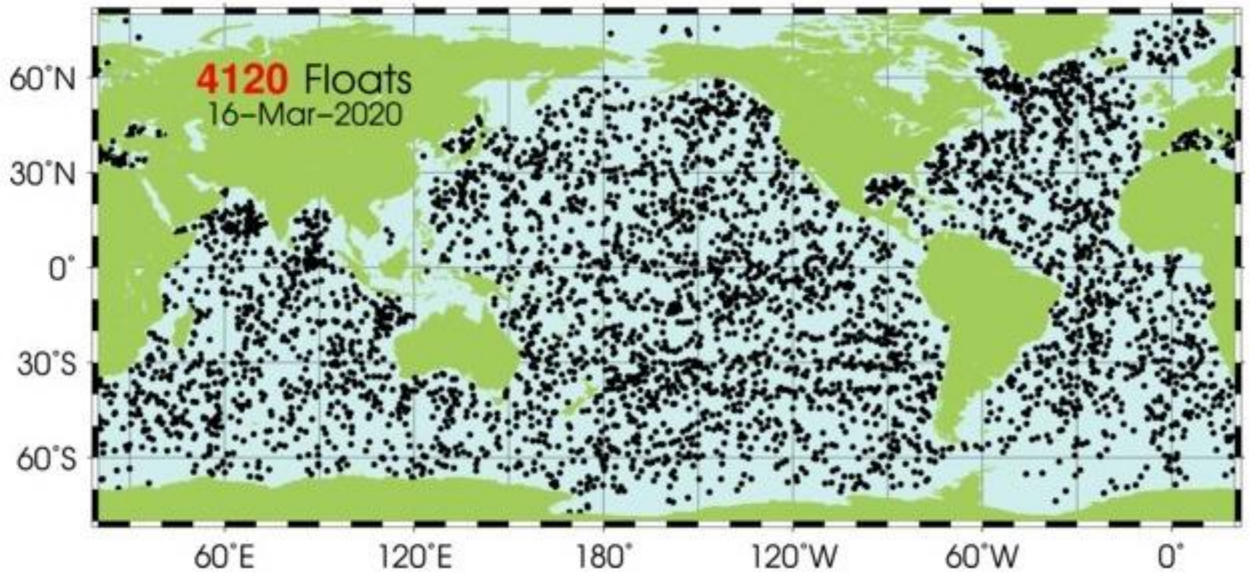
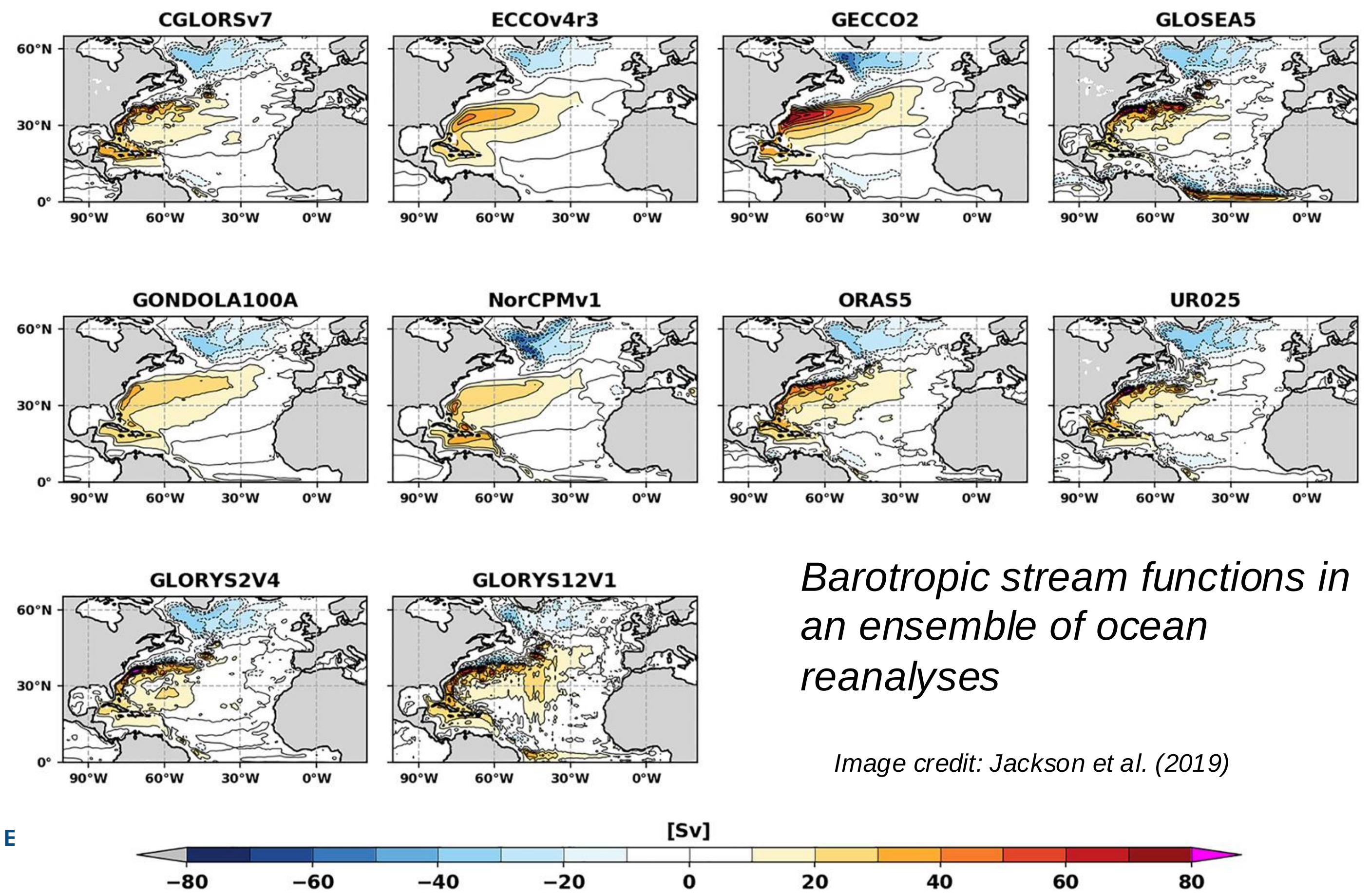


Image credit: argo.uscd.edu

- Since ~2004, the Argo network has measured temperature and salinity in the upper 2000 m with near global coverage.
- It has revolutionized the ocean observing system, but it wasn't designed to constrain the mesoscale (i.e. average spacing >> L_d)
- Floats cycle to 2000m depth every 10 days.
- Floats distributed over the global oceans (except on shelf seas and under ice) with an average spacing of ~300 km

Forecast challenges: ocean model initialization

- Data assimilating ocean reanalyses are (arguably) less mature than their atmospheric counterparts, which is reflected in the diversity of solutions in different products for key ocean properties.
- This is a consequence of (i) the small spatial scales that require very high resolution models, (ii) the complexity of representing coastlines/bathymetry, and (iii) the difficulty of constraining the 3D ocean state from observations.



Barotropic stream functions in an ensemble of ocean reanalyses

Image credit: Jackson et al. (2019)

Summary

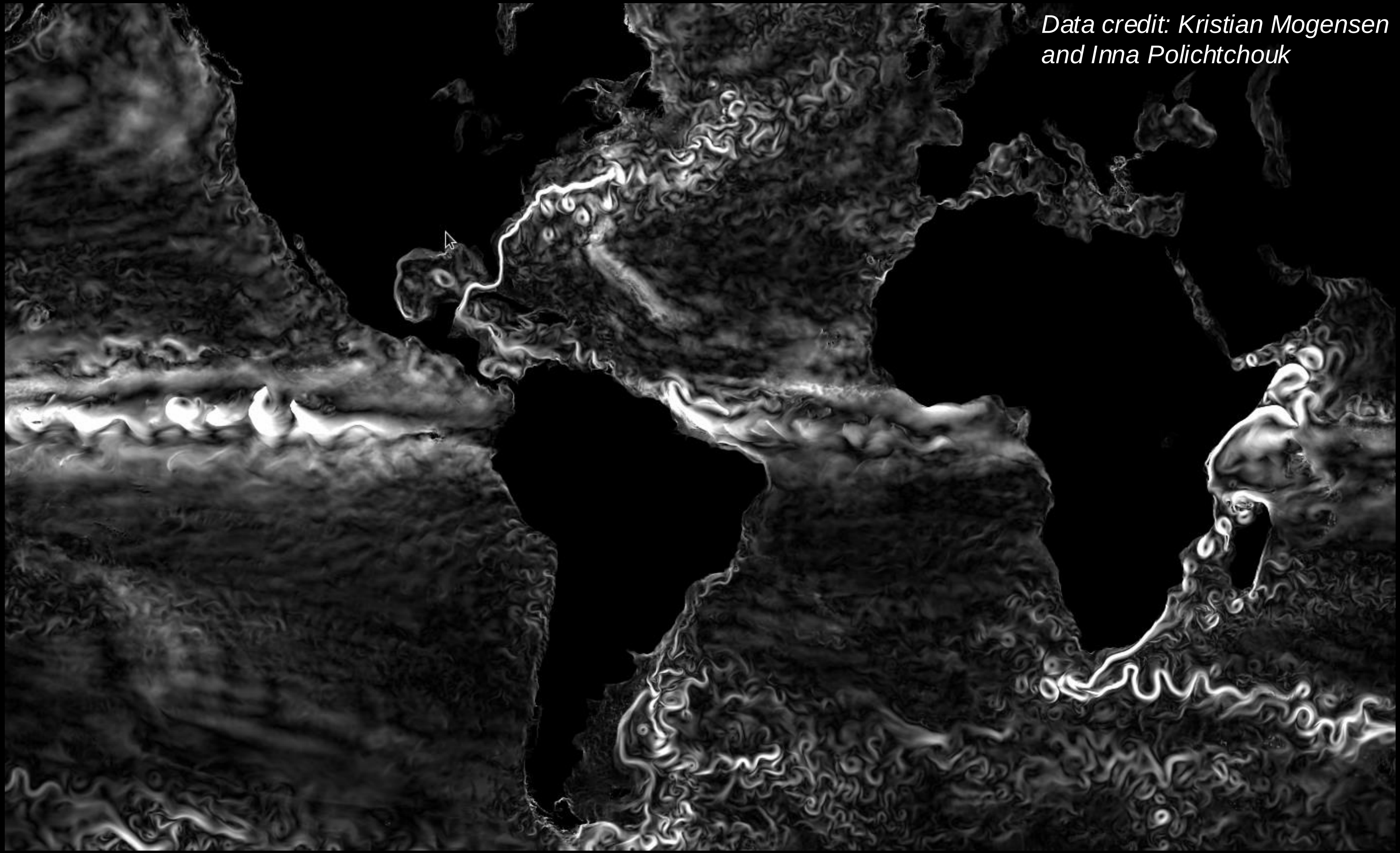
Summary

- **Why is it important for ECMWF to model the ocean?**
  - *Processes*: a dynamic ocean is necessary to represent key coupled ocean-atmosphere feedbacks.
  - *Persistence*: the large heat capacity of the ocean provides “memory” to the climate system.
  - *Predictable dynamics*: well-resolved and predictable ocean dynamics (e.g. equatorial waves) are important for ENSO and interannual variability.
  - *Products*: users are interested in the ocean state.

- **Space and time scales of air-sea interaction:**
  - Coupled ocean-atmosphere interaction occurs across huge range of spatial and temporal scales.
  - Processes that occur on timescales beyond our forecast horizon are still important as they must be correctly modelled in ocean/atmospheric reanalyses to provide accurate forecast initial conditions.

- **Coupled forecasting challenges:**
  - Accurate simulation of ocean boundary currents and mesoscale eddies requires significant increases in ocean resolution. This is computationally very expensive, which motivates the development of novel modelling approaches (e.g. reduced numerical precision, unstructured grids).
  - Model deficiencies and limitations of the ocean observing system result in biases and random errors in our estimates of the instantaneous ocean state used in forecast initial conditions. Reduction of these biases and improved representations of initial condition uncertainty (e.g. ocean stochastic physics) is important for accurate and reliable coupled forecasts.

The future of coupled forecasts at ECMWF?



This image shows a snapshot of surface current speeds in a prototype version of the ECMWF global forecast model (IFS ~4 km, NEMO ~8 km).

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Extra slides

Other important concepts

Ekman transports: Inclusion of a frictional surface boundary results in a departure from geostrophic balance near the ocean surface. The resulting depth-integrated ageostrophic flow (a.k.a. Ekman transport) is to the right (left) of the imposed wind-stress in the northern (southern) hemisphere. *See extra slide 1.*

Ekman suction/pumping: Wind stress curl leads to divergence/convergence of Ekman transports, which drives vertical velocities at the base of the Ekman layer (a.k.a. Ekman "pumping"/"suction") in ocean gyres, along coastlines, and at the equator. *See extra slide 2.*

Potential vorticity (PV): In the ocean interior (i.e. away from frictional boundaries and diabatic forcing), PV is conserved and basin-scale flow is constrained to follow f/H contours. Stretching of a column (increased H) is balanced by increased f (i.e. poleward motion). *See extra slide 3.*

Sverdrup balance: The steady geostrophic flow in the ocean interior is forced from above by Ekman pumping/suction. This relationship is a consequence of vorticity balance and demonstrates that the depth-integrated horizontal gyre circulations are driven by wind-stress curl. *See extra slide 4.*

Western boundary currents: Sverdrup balance does not predict the existence of strong western boundary currents. Western intensification of flow is a consequence of conservation of vorticity and requires the existence of ageostrophic terms (i.e. frictional boundary layers). *See extra slide 5.*

Quasi-geostrophic theory: QG flow is close to geostrophic balance (Ro << 1), but retains time-derivatives. Properties are advected with the *geostrophic* velocity but Coriolis terms in the momentum equations retain *ageostrophic* components. Starting point for investigation of Rossby waves. *See extra slide 6.*

Extra slide 1: Ekman transports

Inclusion of frictional terms results in a departure from geostrophic balance near the ocean surface. Can split flow into geostrophic and ageostrophic components:

$$f\mathbf{k} \times (\mathbf{u_g} + \mathbf{u_{ek}}) + \frac{1}{\rho_0}\nabla_h p = \frac{1}{\rho_0}\frac{\partial \tau}{\partial z}$$

$$f\mathbf{k} \times \mathbf{u_{ek}} = \frac{1}{\rho_0}\frac{\partial \tau}{\partial z}$$ (ageostrophic component)

Can avoid specifying the vertical distribution of stress by integrating over the Ekman layer with appropriate boundary conditions:

$$\tau = (\tau_x, \tau_y) \quad \text{and} \quad \tau(0) = \tau_{wind}, \quad \tau(z_{ek}) = 0$$

$$\mathbf{M_{ek}} = \frac{\tau_{wind} \times \mathbf{k}}{f} \quad \text{where} \quad \mathbf{M_{ek}} = \rho_0 \int_{z_{ek}}^{0} \mathbf{u_{ek}} dz$$

Assumptions:
- As geostrophic, but with a frictional surface boundary.
- Friction included as wind stress vector (***τ***), which represents vertical turbulent fluxes of horizontal momentum.
- Frictional effects from wind stress contained within the near-surface "Ekman layer" (i.e. *$\tau(Z_{ek}) = 0$).*

Thus depth-integrated Ekman flow is to the right (left) of imposed wind-stress in the northern (southern) hemisphere.

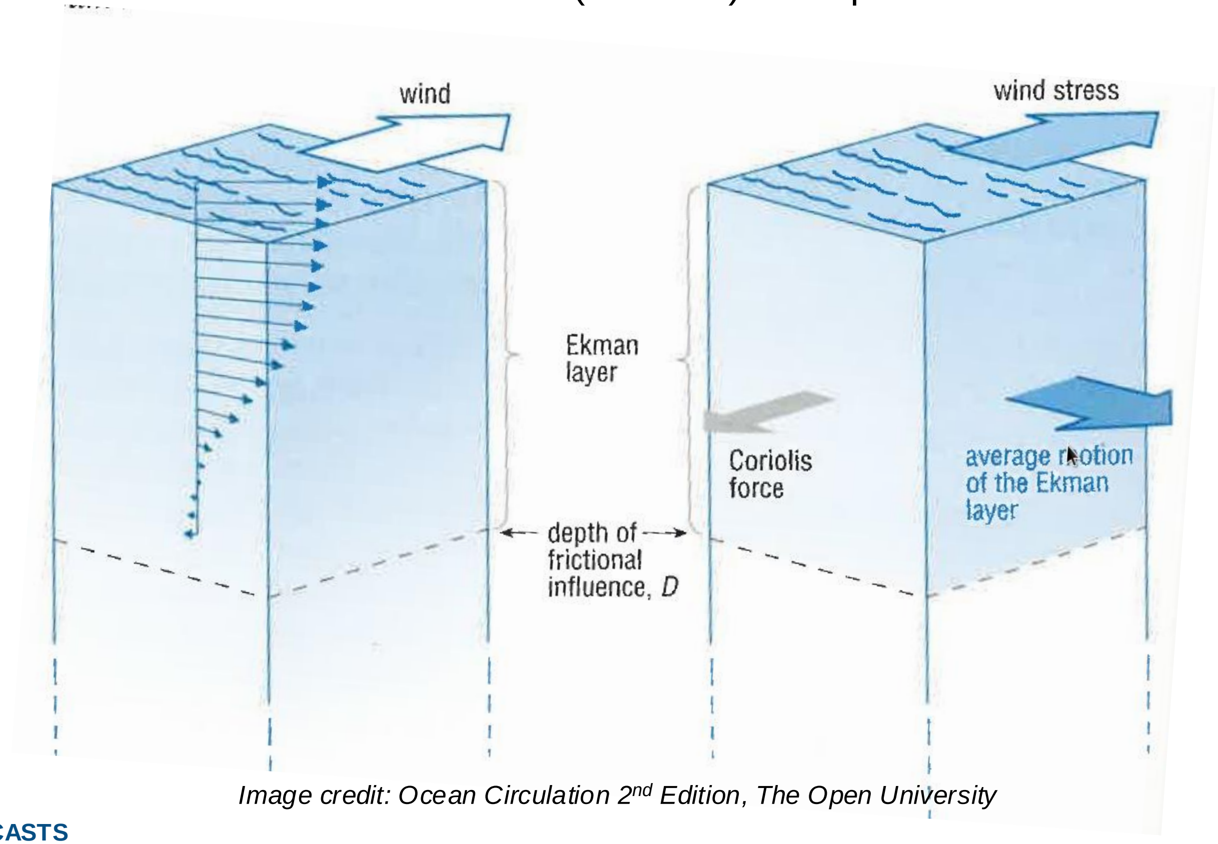


Image credit: Ocean Circulation 2nd Edition, The Open University

Extra slide 2: Ekman pumping

Assumptions:

- Within the Ekman layer, divergence of geostrophic current is small compared to divergence of Ekman transports.

Horizontal divergence of the integrated Ekman transports gives rise to a vertical velocity at the base of the Ekman layer (i.e. Ekman “pumping”):

$$\nabla_h \cdot \mathbf{u} \simeq \nabla_h \cdot \mathbf{u_{ek}} = -\frac{\partial w}{\partial z}$$

$$w_{ek} = \frac{1}{\rho_0}\nabla_h \cdot \mathbf{M_{ek}} = \frac{1}{\rho_0}\nabla_h \times \frac{\tau_{wind}}{f}$$

Vertical velocities are a result of wind-stress curl (and variations in *f* near the equator). In both hemispheres:

- Anticyclonic winds drive downwelling (Ekman pumping).
- Cyclonic winds drive upwelling (Ekman suction).

Divergence of Ekman transports also drives coastal and equatorial upwelling/downwelling.

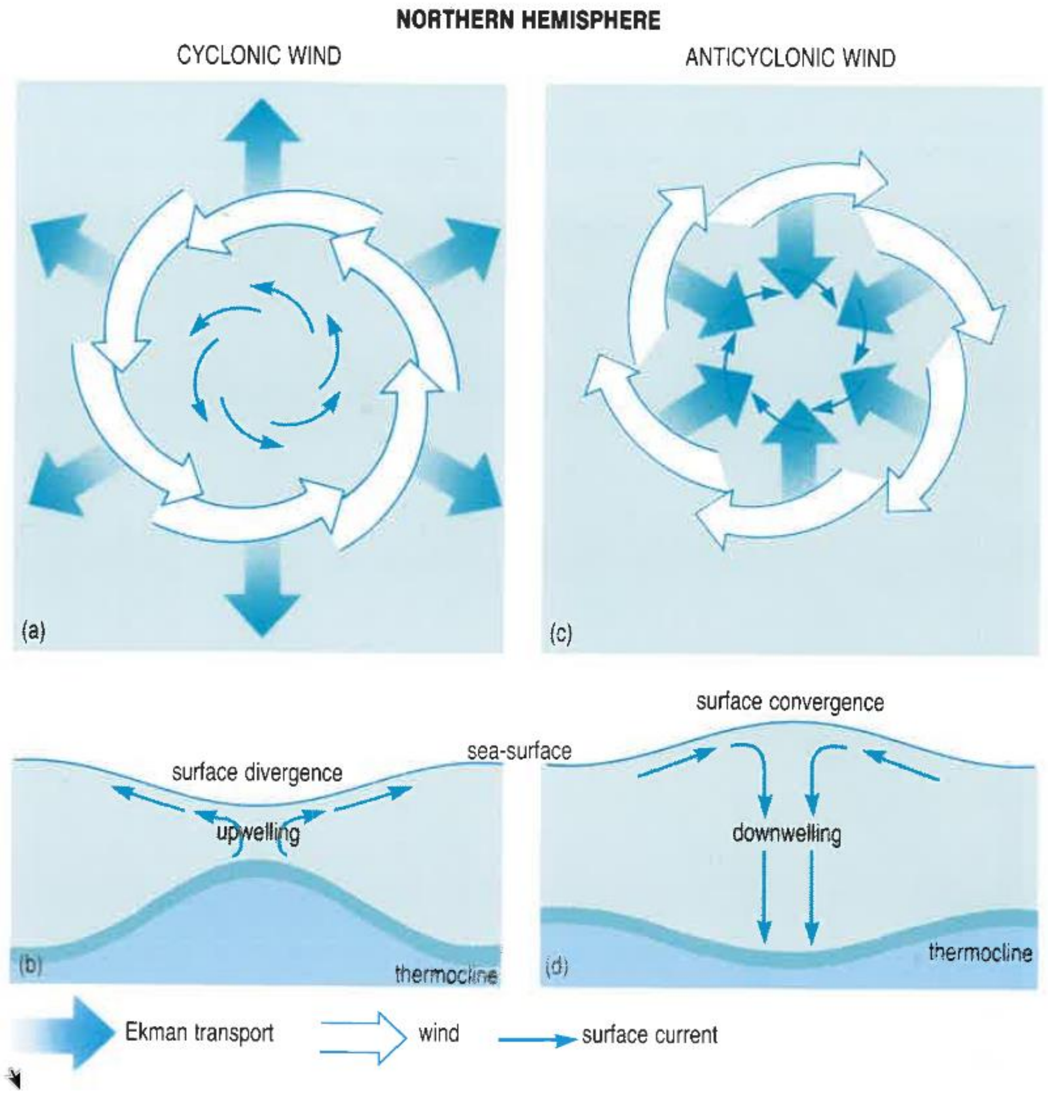


Image credit: Ocean Circulation 2nd Edition, The Open University

Extra slide 3: Potential vorticity conservation

Potential vorticity (PV) is a quantity that is conserved with fluid flow derived by combining conservation of mass and vorticity. For a layer of constant density and thickness (H):

$$PV = \frac{\zeta + f}{H} \quad \text{where} \quad \zeta = \nabla_h \times \mathbf{u_h}$$

$$\text{At planetary scale (i.e. Ro} << 1) \quad f >> \zeta \quad \text{and} \quad PV = \frac{f}{H}$$

In the ocean interior, PV is conserved with fluid flow:

$$\frac{D}{DT}\left(\frac{\zeta + f}{H}\right) = 0 \quad \text{(unless direct forcing or friction)}$$

At the planetary scale, flow is constrained to follow *f*/H contours. Stretching of a column (i.e. increased H) is balanced by increased *f* (i.e. poleward motion).

PV conservation breaks down under the following conditions:

- Frictional boundaries
- Heating/cooling at the surface.
- Mixing across isopycnals (i.e. changing stratification).

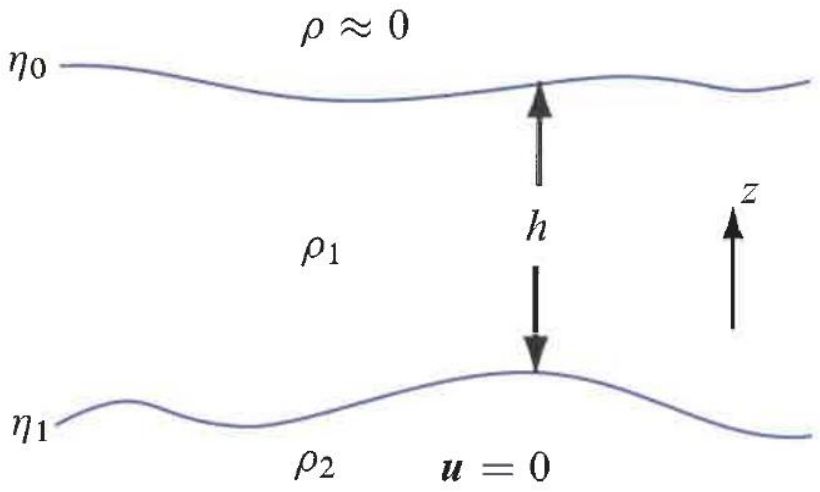
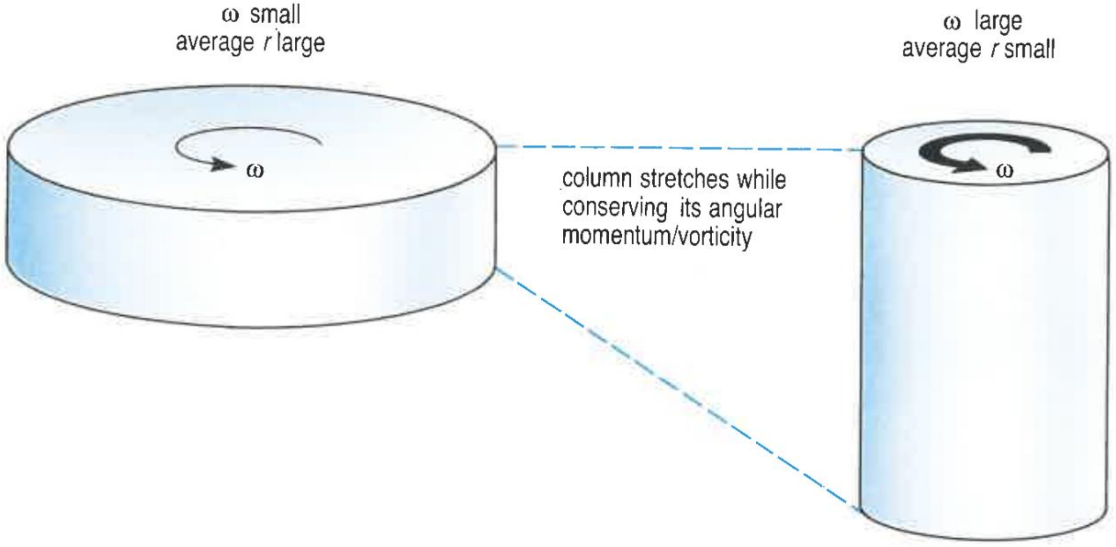


Image credit: Atmospheric and Oceanic Fluid Dynamics 2nd Edition, Vallis



(b)

Image credit: Ocean Circulation 2nd Edition, The Open University

Extra slide 4: Sverdrup balance

Combine continuity equation with divergence of geostrophic flow in the ocean interior:

$$\nabla_h \cdot \mathbf{u_g} + \frac{\partial w}{\partial z} = 0 \quad \text{and} \quad \beta = \frac{\partial f}{\partial y}$$

$$\nabla_h \cdot \mathbf{u_g} = \frac{\partial}{\partial x}\left(-\frac{1}{f\rho_0}\frac{\partial p}{\partial y}\right) + \frac{\partial}{\partial y}\left(\frac{1}{f\rho_0}\frac{\partial p}{\partial x}\right) = -\frac{\beta}{f}v_g$$

$$\boxed{\beta v_g = f\frac{\partial w}{\partial z}}$$ (Sverdrup relation)

This equation relates geostrophic currents to vertical velocities and is an expression of (linearized) vorticity balance.

Squashing (stretching) columns of water moves them equatorward (poleward).

Assumptions:

- Steady geostrophic flow in ocean interior (below Ekman layer, away from frictional boundaries).
- Forced from above by Ekman pumping/suction.
- Assumed can integrate to a level of no motion where w=0

Now integrate from the base of the Ekman layer to the a known level of no motion (e.g. ocean floor):

$$w(z_{ek}) = w_{ek} \quad \text{and} \quad w(z_{bottom}) = 0$$

$$\beta \int_{z_{bottom}}^{z_{ek}} v_g dz = f \int_{z_{bottom}}^{z_{ek}} \frac{\partial w}{\partial z} dz$$

$$\frac{\beta}{\rho_0} V_g = f w_{ek} \quad \text{where} \quad V_g = \int_{z_{bottom}}^{z_{ek}} v_g \rho_0$$

Substitue w_{ek} and manipulate:

$$\frac{\beta}{\rho_0} V_g = \frac{f}{\rho_0}\left(\frac{\partial}{\partial x}\left(\frac{\tau^y_{wind}}{f}\right) - \frac{\partial}{\partial y}\left(\frac{\tau^x_{wind}}{f}\right)\right)$$

$$\boxed{V = V_g + V_{ek} = \frac{1}{\beta}\nabla_h \times \tau_{wind}}$$

We get an expression for the total depth-integrated meridional flow (Sverdrup transport) that is a function of *wind-stress only.*

Extra slide 5: Wind-driven gyres

Can now define a streamfunction (Ψ) in terms of the zonal integral of wind-stress curl that represents the 2D depth-integrated flow:

$$\frac{\partial \Psi}{\partial x} = \frac{V}{\rho_0} \quad \text{and} \quad \frac{\partial \Psi}{\partial y} = -\frac{U}{\rho_0}$$

$$\Psi(x, y) = \frac{1}{\rho_0 \beta} \int_{\text{eastern boundary}}^{x} \nabla_h \times \tau_{wind} dx$$

Integrating from the eastern boundary gives results consistent with the existence of strong western boundary currents.

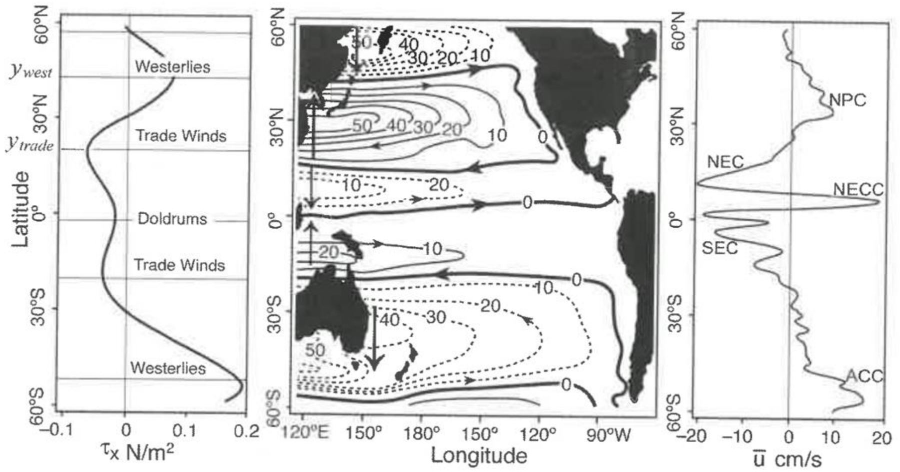


FIGURE 10.21. (left) The zonal-average of the zonal wind stress over the Pacific Ocean. (middle) The Sverdrup transport stream function (in $Sv = 10^6 \, m^3 s^{-1}$) obtained by evaluation of Eq. 10-20 using climatological wind stresses. Fig. 10.2. Note that no account has been made of island ... ough them. The

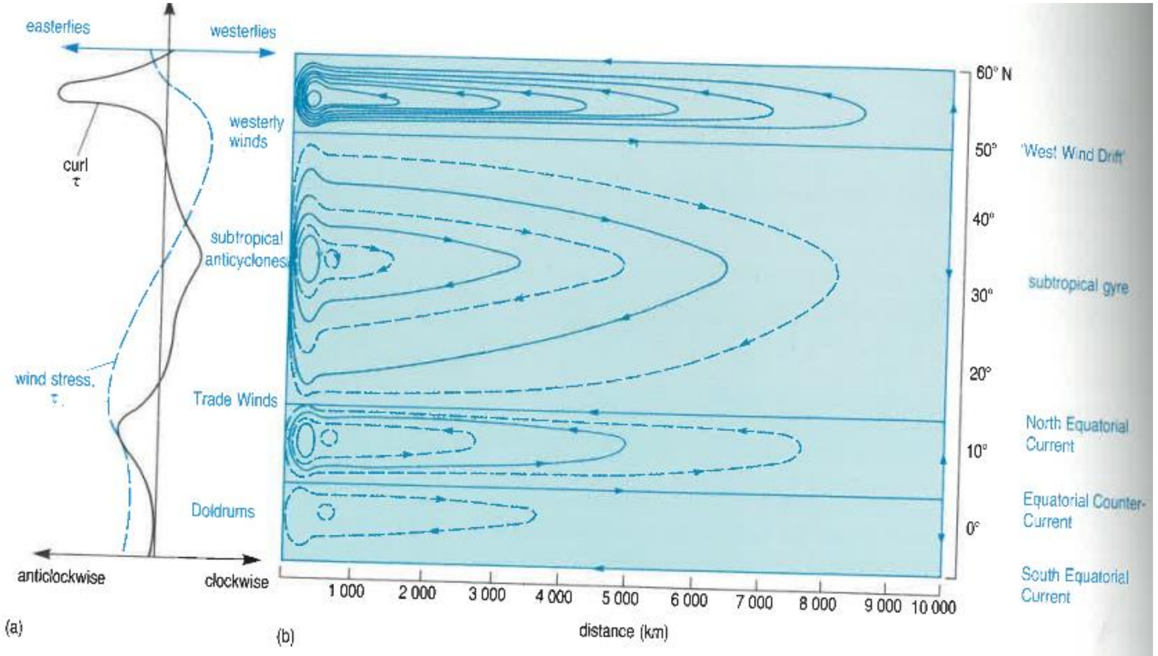
FORECASTS

Image credit: Atmosphere, Ocean, and Climate Dynamics, Marshall & Plumb

However, Sverdrup theory does not predict the existence of a strong WBC - could also integrate from the western boundary. Western intensification of flow is a consequence of conservation of vorticity and requires the existence of ageostrophic terms (i.e. frictional boundary layers).

Stommel (1948) and Munk (1950) developed models of the depth-integrated wind-driven flow that predict the existence of western boundary currents by including representations of linear drag and momentum diffusion, respectively.

Munk gyres showing western boundary currents.



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Extra slide 6: Quasi-geostrophic theory

- No tendency/development term in geostrophic balance: cannot be used investigate time-dependent phenomena.
- Quasi-geostrophic theory is close to geostrophic balance (Ro << 1), but retains time-derivatives.
- Properties are advected with the *geostrophic* velocity but Coriolis terms in the momentum equations retain *ageostrophic* components.
- Shallow water quasi-geostrophic potential vorticity (q):

$$q = \nabla^2\Psi + \beta y - \frac{1}{L_d^2}\Psi$$

= relative vorticity + planetary vorticity + vortex stretching

$$\Psi = \frac{g}{f_0}\eta \quad \text{and} \quad u_g = -\frac{\partial \Psi}{\partial y}, \quad v_g = \frac{\partial \Psi}{\partial x}$$

Assumptions:

- Scale of flow similar to deformation length scale ($L_d = \sqrt{gH}/f$).
- Close to geostrophic balance (Ro << 1).
- Variations in Coriolis parameter are small (f = f0 + βy)
- Time can be scaled advectively (T = L/U)

- Quasi-geostrophic potential vorticity is materially conserved with the flow:

$$\frac{Dq}{Dt} = \frac{D}{Dt}\left(\nabla^2\Psi + \beta y - \frac{1}{L_d^2}\Psi\right) = 0$$

- Single prognostic variable (q), from which all other dynamic variables can be diagnosed (PV inversion).
- Starting point for investigation of Rossby waves.

Extra slide 7: Rossby wave propagation mechanism

- Propagation mechanism is a function of wavenumber relative to deformation radius (L_d).
- Short waves: balance between relative vorticity and planetary vorticity.
- Long waves: balance between vortex stretching and planetary vorticity
- Waves can be a mixture of both mechanisms.

$$u_t - fv = -g'\eta_x, \quad v_t + fu = -g'\eta_y, \quad \eta_t + H(u_x + v_y) = 0$$

Propagation mechanism for short waves (i.e. $f + \varsigma$ = constant)

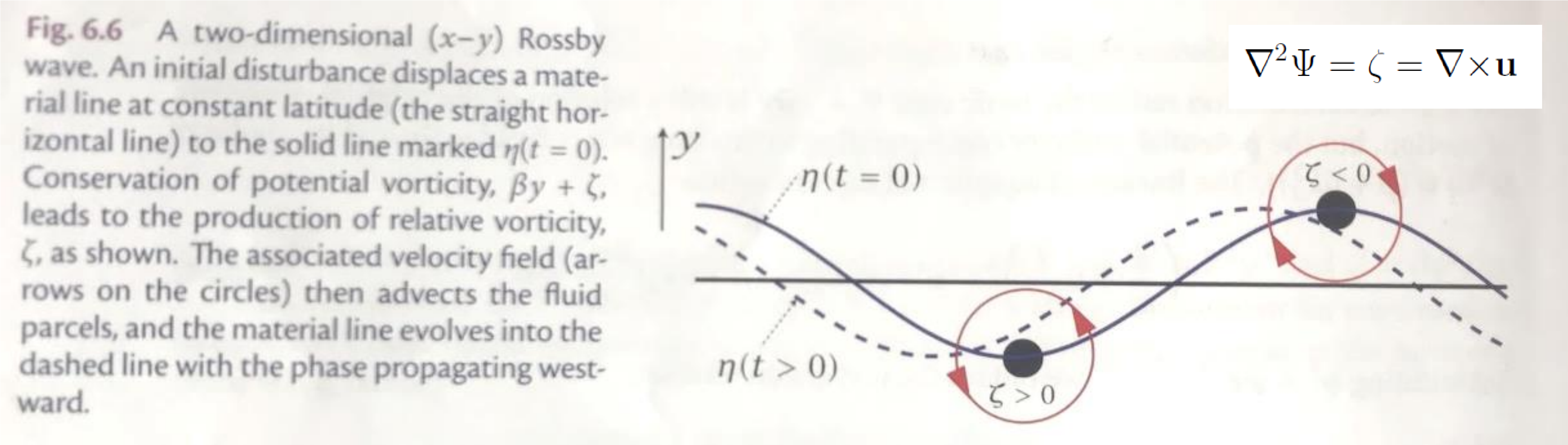
Fig. 6.6 A two-dimensional (x–y) Rossby wave. An initial disturbance displaces a material line at constant latitude (the straight horizontal line) to the solid line marked $\eta(t=0)$. Conservation of potential vorticity, $\beta y + \zeta$, leads to the production of relative vorticity, ζ, as shown. The associated velocity field (arrows on the circles) then advects the fluid parcels, and the material line evolves into the dashed line with the phase propagating westward.

Image credit: Atmospheric and Oceanic Fluid Dynamics 2nd Edition, Vallis