

The role of the oceans for prediction across timescales

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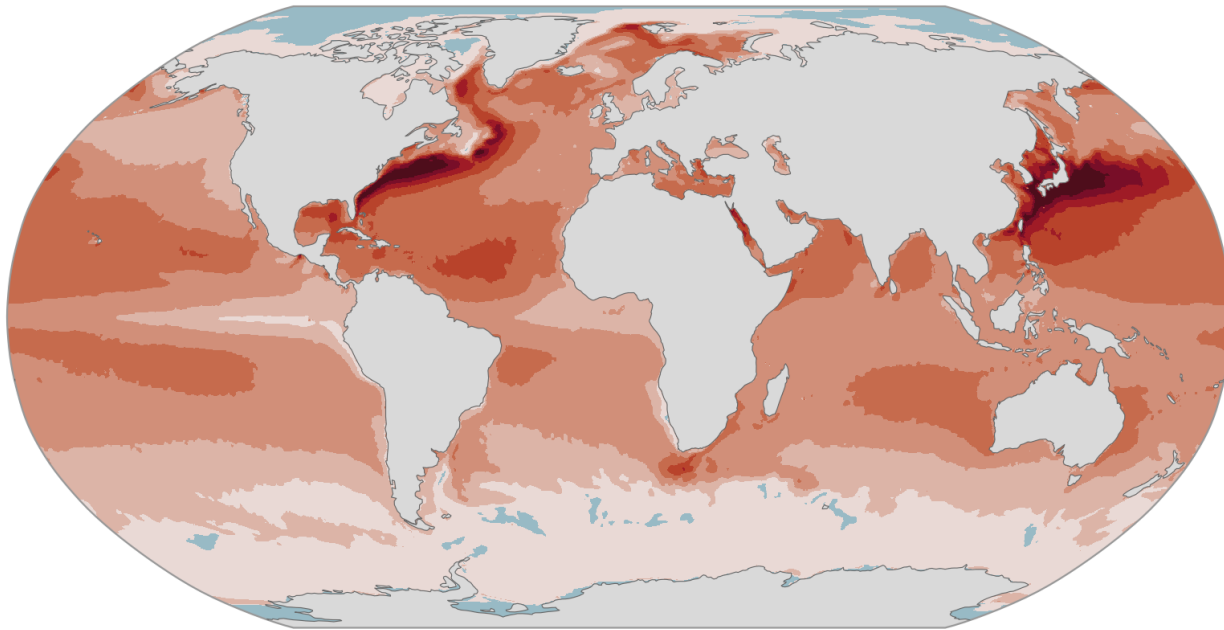
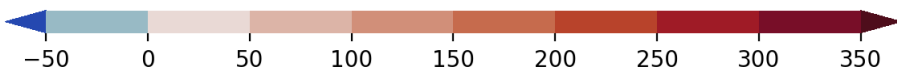


The oceans and atmosphere redistribute uneven solar heating

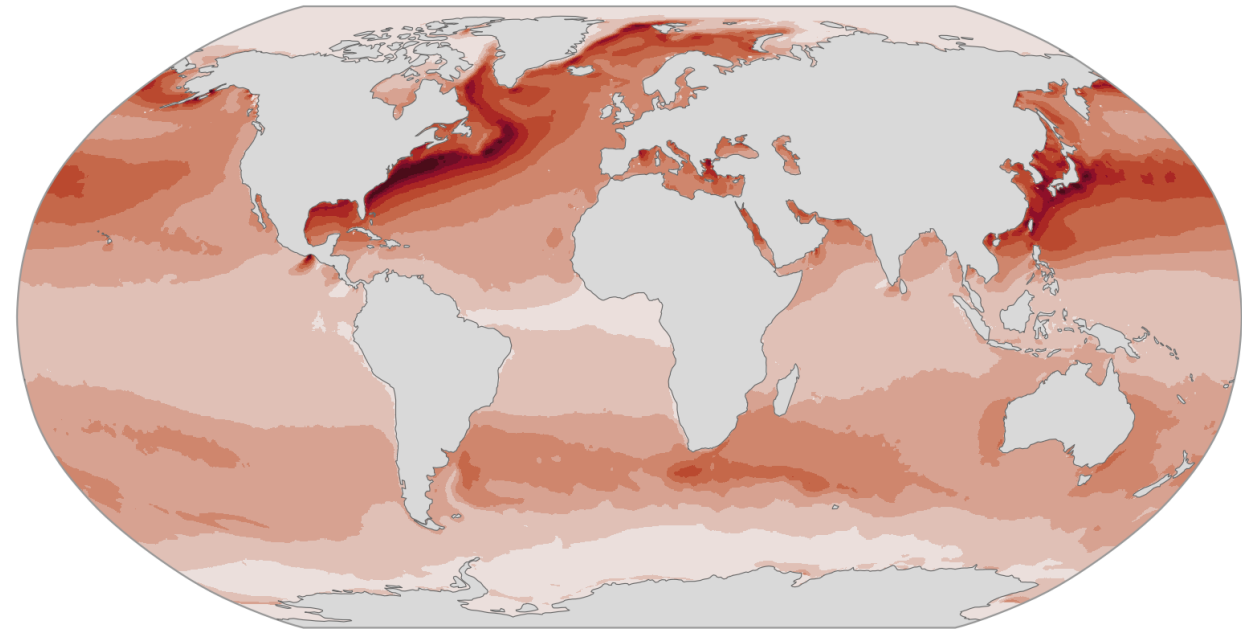
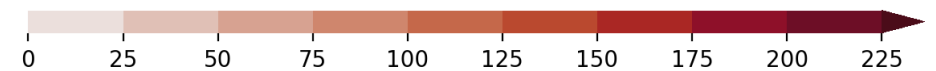
The turbulent heat release from the ocean, and its day-to-day variability, is concentrated in "hot spots" over western boundary currents and strong sea surface temperature gradients, where heat carried poleward by the ocean is given up to the atmosphere.

Turbulent (sensible+latent) heat flux into the atmosphere (DJF, 1993-2012)

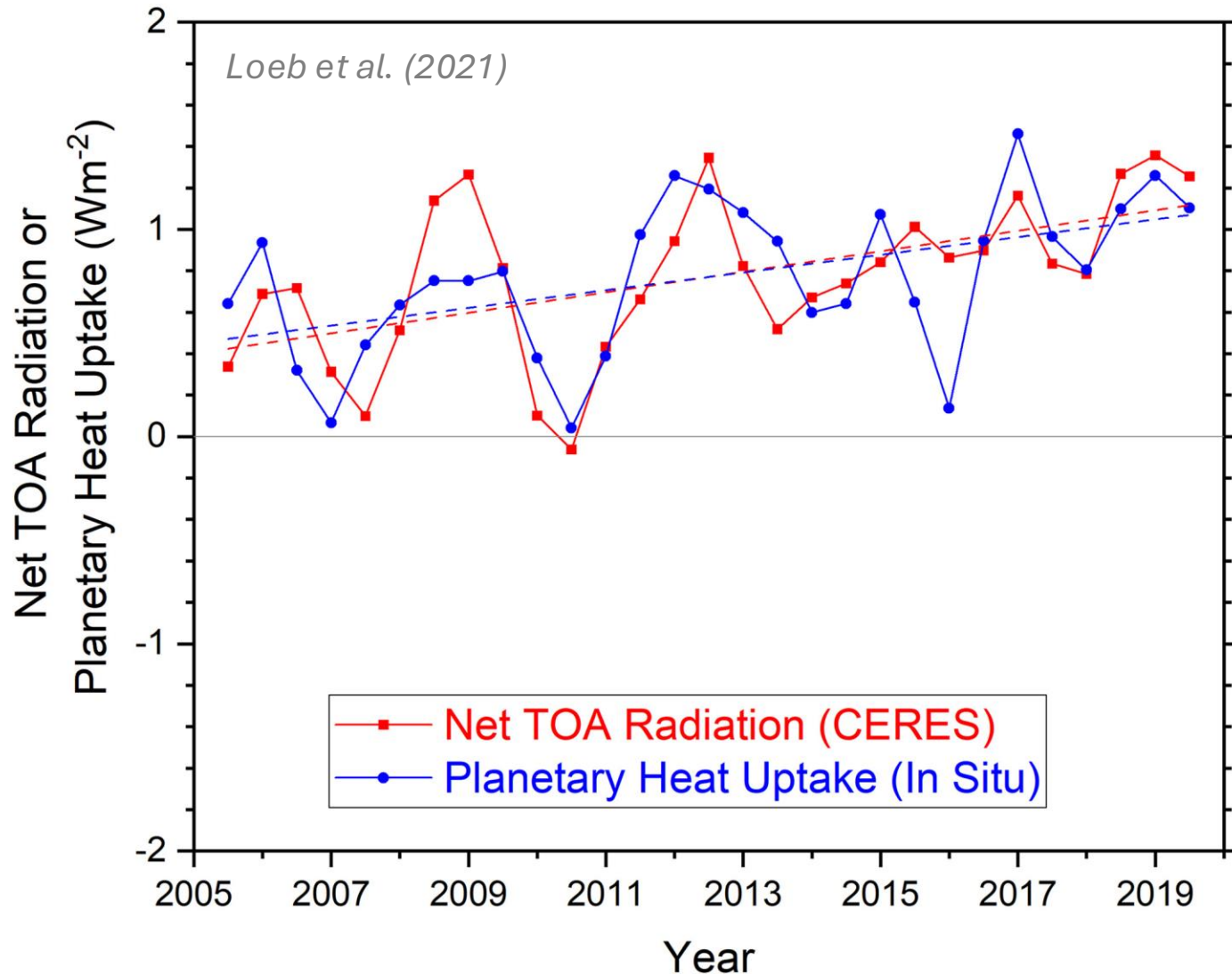
Climatological mean (W/m^2)



Standard deviation of daily means (W/m^2)



Earth's energy budget is (mostly!) balanced by ocean heat uptake



About 90% of the excess energy associated with top-of-atmosphere (TOA) radiative imbalance is stored in the ocean (von Schuckmann et al., 2023).

Global energy balance in response to climate change can be approximated:

$$\Delta T = (F - N) / \alpha$$

Global mean surface temperature change (K)

Climate forcing (W m^{-2})

Net TOA radiation $\approx$ ocean heat uptake (W m^{-2})

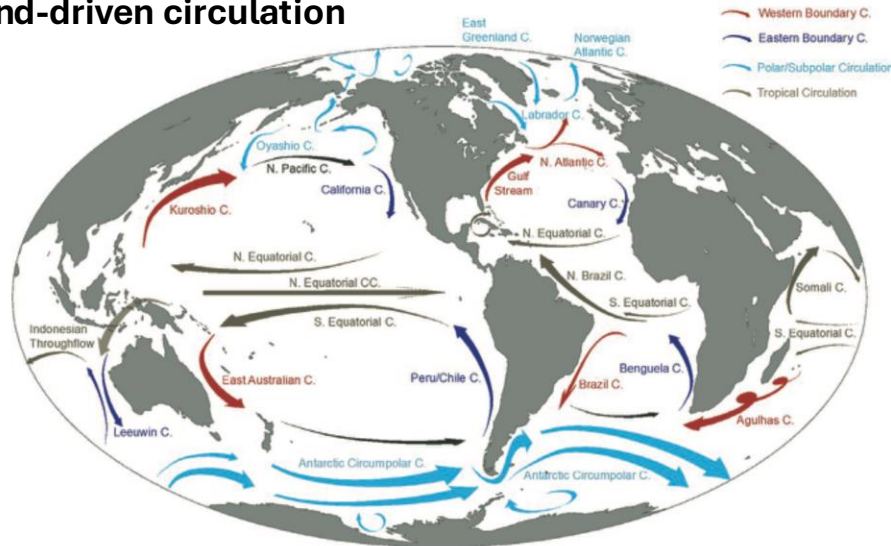
Climate sensitivity parameter ($\text{W m}^{-2} \text{K}^{-1}$)

At steady state, $N \rightarrow 0$ and $\Delta T \rightarrow F/\alpha$. However, ocean heat uptake (N) moderates the rate of time-dependent climate change.

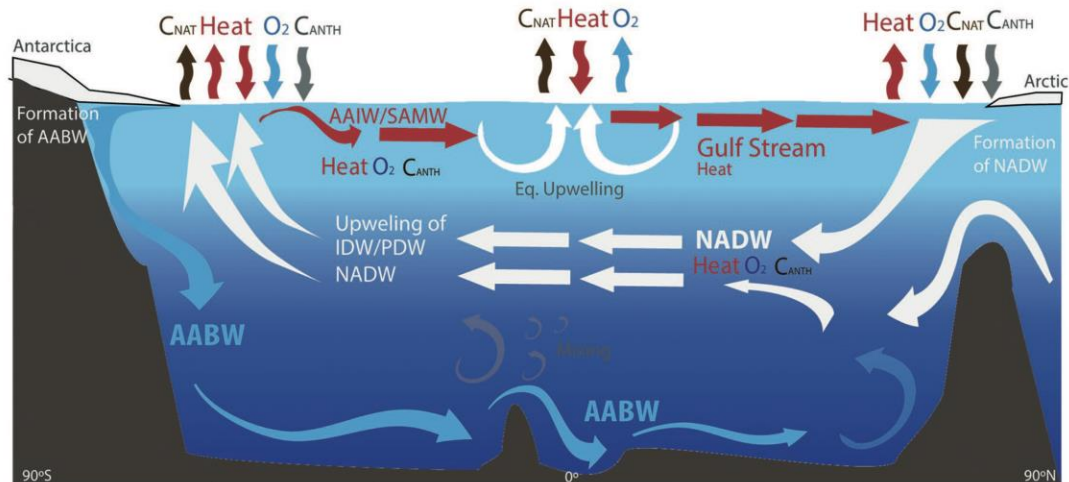
For given forcing (F) and climate sensitivity (α), the transient climate response (ΔT) is reduced when ocean heat uptake (N) is higher.

Ocean circulation changes contribute to regional climate uncertainty

Wind-driven circulation

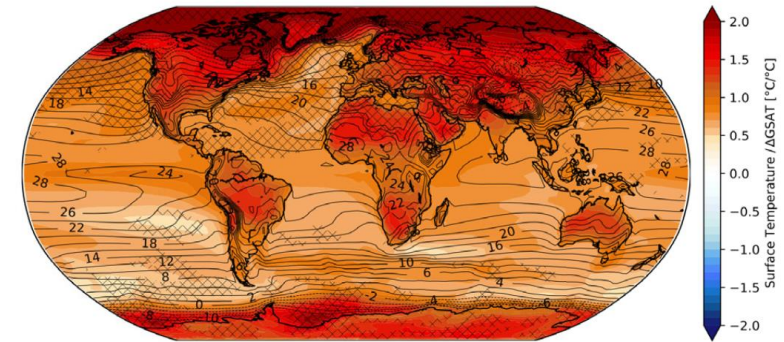


Overtaking circulation

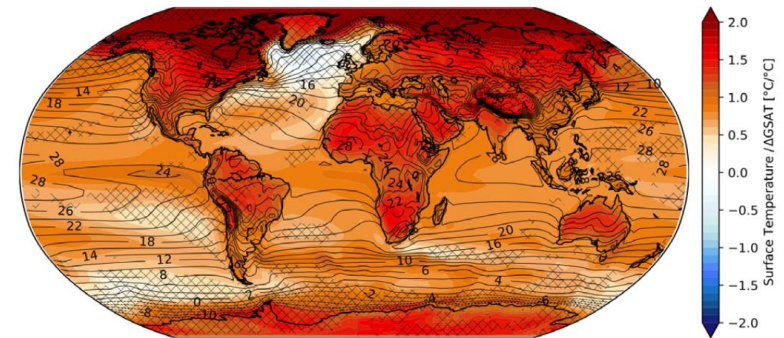


Bertrand Delorme and Yassir Eddebbar

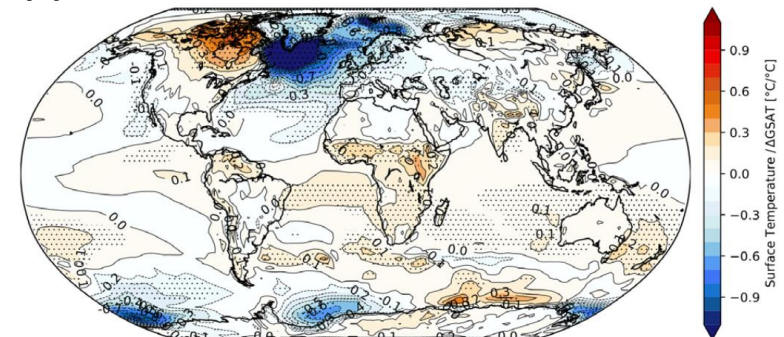
(a) ΔT_{2m} under $4xCO_2$ + weak AMOC decline



(b) ΔT_{2m} under $4xCO_2$ + strong AMOC decline



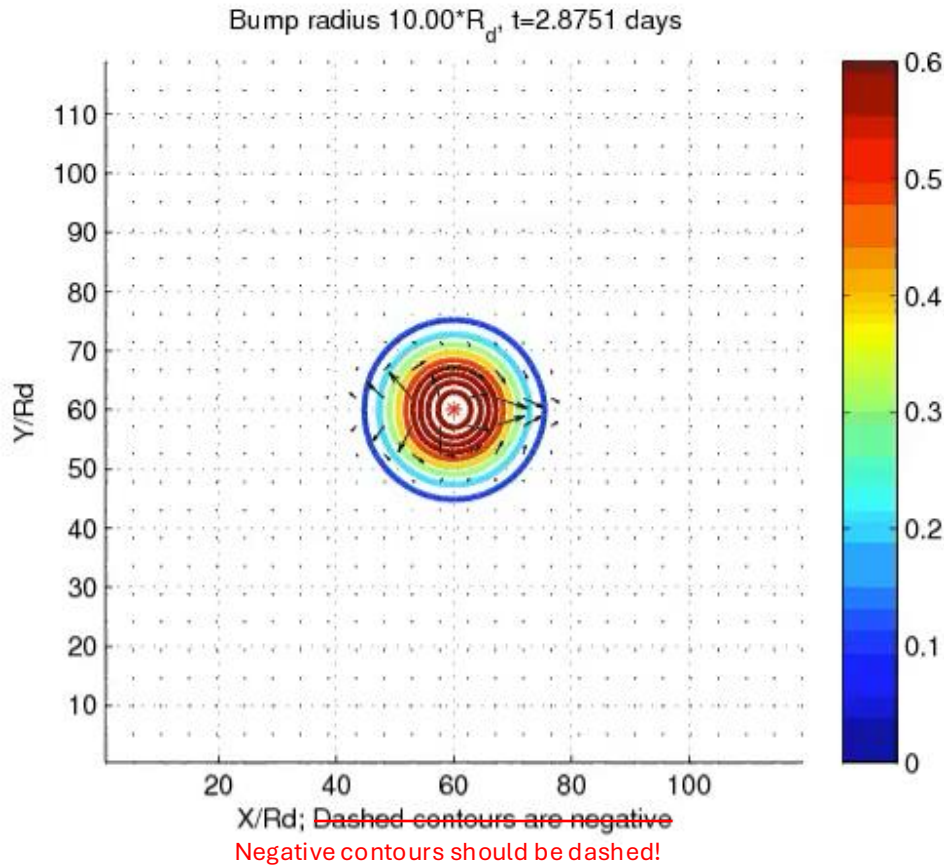
(c) Difference



Bellomo et al. (2021)

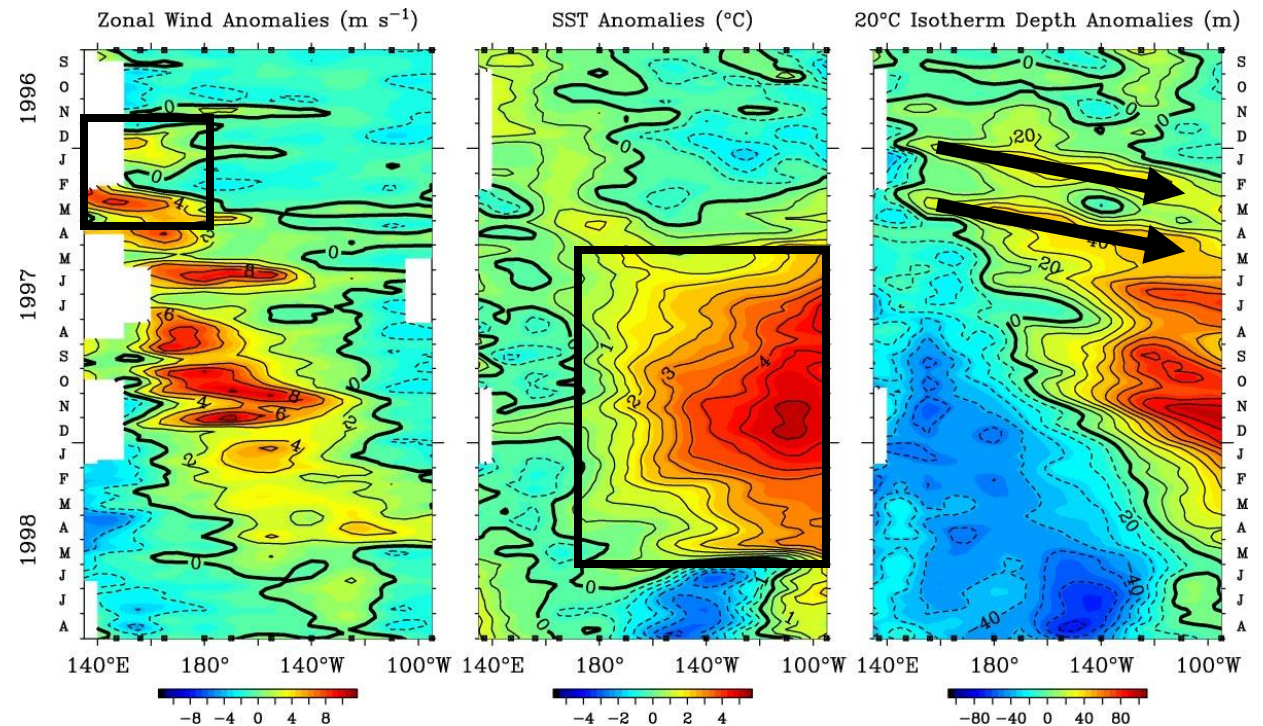
Internal ocean waves are important for coupled climate variability

Rossby and Kelvin wave response to an equatorial disturbance in a reduced-gravity shallow water model



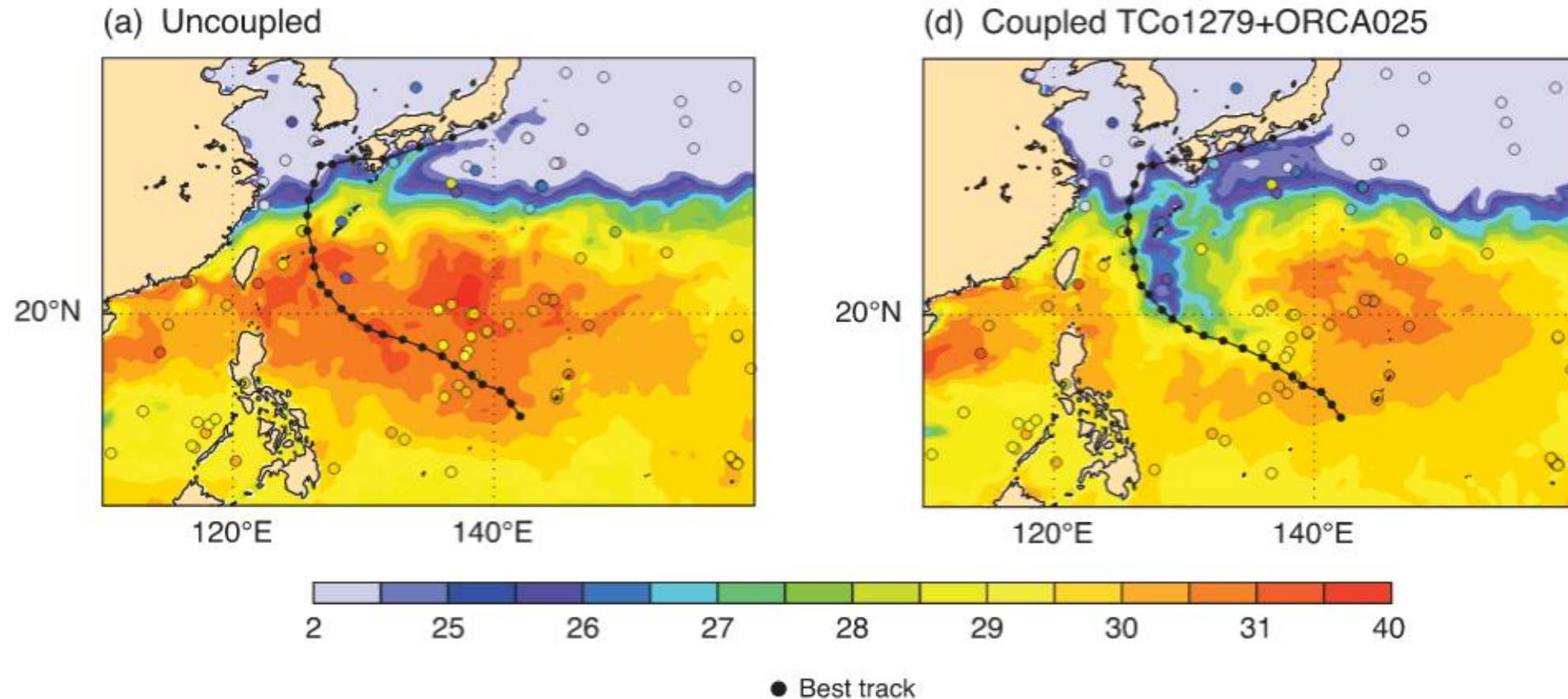
Large-scale equatorial wave dynamics are well-resolved by global ocean models and provide a source of predictability for modes of coupled ocean-atmosphere variability (e.g. ENSO).

Example: The 1997-98 El Niño was preceded by MJO-related westerly wind events that generated equatorial Kelvin waves.



Air-sea interactions and coupled feedbacks are important for NWP

Case study of tropical cyclone Neoguri (2014) 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.



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).

Governing equations for the ocean (+ typical approximations)

Horizontal momentum ($F=ma$)

$$\begin{aligned}
 &\text{advection (non-linear)} && \text{pressure gradient} && \text{surface stress + mixing} \\
 &\boxed{\frac{\partial u}{\partial t}} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - \boxed{fv} = -\frac{1}{\rho_0} \frac{\partial p}{\partial x} + \frac{1}{\rho_0} \frac{\partial \tau_x}{\partial z} + \boxed{F_x^h} \\
 &\boxed{\frac{\partial v}{\partial t}} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + \boxed{fu} = -\frac{1}{\rho_0} \frac{\partial p}{\partial y} + \frac{1}{\rho_0} \frac{\partial \tau_y}{\partial z} + \boxed{F_y^h} \\
 &\text{tendency} && \text{rotation/Coriolis} && \text{lateral friction}
 \end{aligned}$$

$\mathbf{u} = (u, v, w)$ = velocity

ρ = density

ρ_0 = reference density (Boussinesq)

p = pressure

g = effective gravity

$f = 2\Omega \sin \phi$ = Coriolis parameter

Ω = Earth's rotation rate, ϕ = latitude

$\boldsymbol{\tau} = (\tau_x, \tau_y)$ = stress (positive downward)

F_x^h, F_y^h = lateral friction (per unit mass)

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

Vertical momentum (hydrostatic approximation i.e. $L \gg D$)

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

Volume conservation (Boussinesq approximation i.e. $\delta\rho \ll \rho_0$)

The atmosphere enters the momentum equations through $\boldsymbol{\tau}$ at the surface ($z = \eta$)

Governing equations: tracers and surface boundary conditions

vertical fluxes and mixing

$$\frac{D\Theta}{Dt} = \frac{1}{\rho_0 c_p} \frac{\partial Q}{\partial z} + \kappa_h \nabla_h^2 \Theta$$

Heat conservation

$$\frac{DS}{Dt} = \frac{\partial F_S}{\partial z} + \kappa_h \nabla_h^2 S$$

Salt conservation

Rate of change following flow **lateral fluxes and mixing**

$$\rho = \rho(\Theta, S, p)$$

Equation of state (tracers influence density gradients and dynamics)

Surface boundary conditions

$$\tau(z = \eta) = \tau_{oc} = \tau_a - \tau_w$$

(momentum: wind stress minus wave uptake)

$$Q(z = \eta) = Q_{net} = Q_{SW} + Q_{LW} + Q_H + Q_E$$

(net surface heat flux)

$$F_S(z = \eta) = S_0 (E - P)$$

(freshwater flux as virtual salt flux)

$$p(z = \eta) = p_a$$

(atmospheric pressure loading)

All fluxes positive downward (into the ocean)

$$\frac{D}{Dt} = \frac{\partial}{\partial t} + \mathbf{u} \cdot \nabla = \text{material derivative}$$

Θ, S = conservative temperature, salinity

Q, F_S = vertical heat and salt fluxes

c_p = heat capacity of seawater, κ_h = lateral diffusivity

$\tau_{oc}, \tau_a, \tau_w$ = stress into ocean, from atmosphere, into waves

Q_{SW}, Q_{LW}, Q_H, Q_E = shortwave, longwave, sensible, latent

E, P = evaporation, precipitation, S_0 = surface salinity

η = sea surface height, p_a = atmospheric pressure

Coupling is two-way: surface fluxes (e.g. Q_H, Q_E, τ_a) depend on ocean and atmosphere state (e.g. SST, surface currents/winds, temperature, and humidity) and typically estimated via bulk formulae (e.g. Large and Yeager, 2009).

Ocean and atmosphere balances: same dynamics, different scales

Scale	Atmosphere (synoptic)	Ocean mesoscale	Ocean (basin)
L	10^6 m	10^5 m	10^7 m
U	10 m/s	10^{-1} m/s	10^{-2} m/s
L/U (advective)	10^5 s (~1 day)	10^6 s (~10 days)	10^9 s (~30 years)
$Ro = U/f_0L$	10^{-1}	10^{-2}	10^{-5}

Dimensionless u-momentum (neglecting frictional/viscous terms)

$$\left(\frac{1}{f_0T}\right)\frac{\partial u'}{\partial t'} + \left(\frac{U}{f_0L}\right)\mathbf{u}' \cdot \nabla \mathbf{u}'$$

$$- \left(\frac{f}{f_0}\right)v' = - \left(\frac{P}{\rho_0 f_0 L U}\right)\frac{\partial p'}{\partial x'}$$

Tendency and non-linear advection terms are small

Coriolis and pressure gradient terms dominate i.e. geostrophic balance

$x' = x/L, \quad y' = y/L$
 $z' = z/D$
 $u' = u/U, \quad v' = v/U$
 $w' = w/W$
 $t' = t/T$
 $p' = p/P$
 $f_0 =$ reference Coriolis parameter ($\approx 10^{-4} \text{ s}^{-1}$ at mid-latitudes)

$L =$ horizontal length scale
 $D =$ vertical length scale
 $U =$ horizontal velocity scale
 $W = UD/L =$ vertical velocity scale
 $T =$ timescale of interest
 $P =$ pressure scale

$$\left(\frac{1}{f_0T}\right) = \frac{\text{inertial timescale}}{\text{timescale of interest}} \ll 1$$

$$\left(\frac{U}{f_0L}\right) = \frac{\text{inertial terms}}{\text{Coriolis terms}} = \text{Rossby number (Ro)} \ll 1$$

$$\left(\frac{f}{f_0}\right) = O(1)$$

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

The ocean and atmosphere are both close to hydrostatic ($L \gg D$) and **geostrophic balance** ($Ro \ll 1$) at large scales. However, the ocean equivalent of synoptic variability is both much smaller and much slower.

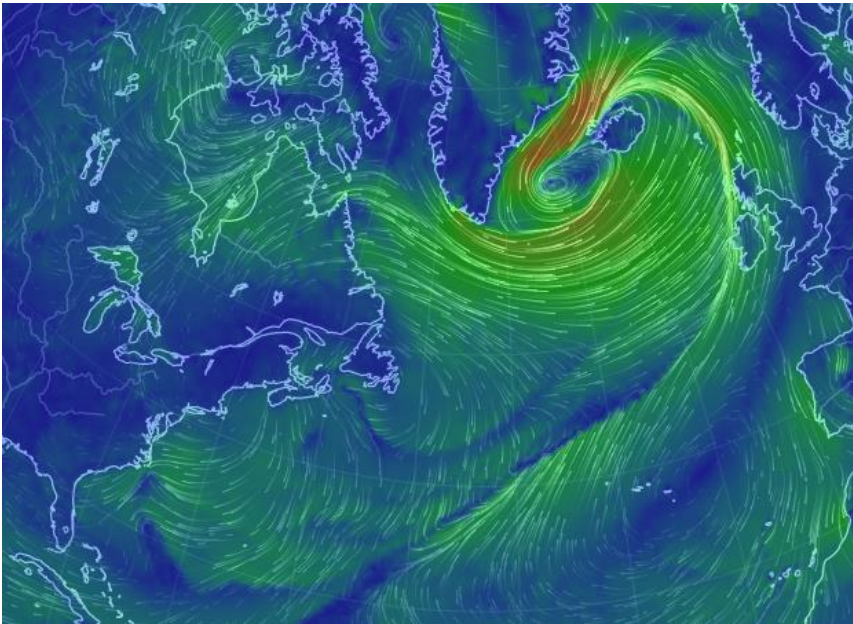
Ocean ‘weather’ scales with the first baroclinic Rossby radius

The Rossby radius of deformation (L_d) is the scale at which rotation and stratification are equally important.

$$g' = \text{reduced gravity} = g \frac{\Delta \rho}{\rho_0}$$

$$L_d = \frac{\text{gravity wave speed}}{\text{planetary vorticity}} = \frac{\sqrt{g'H}}{f}$$

Snapshot of surface winds



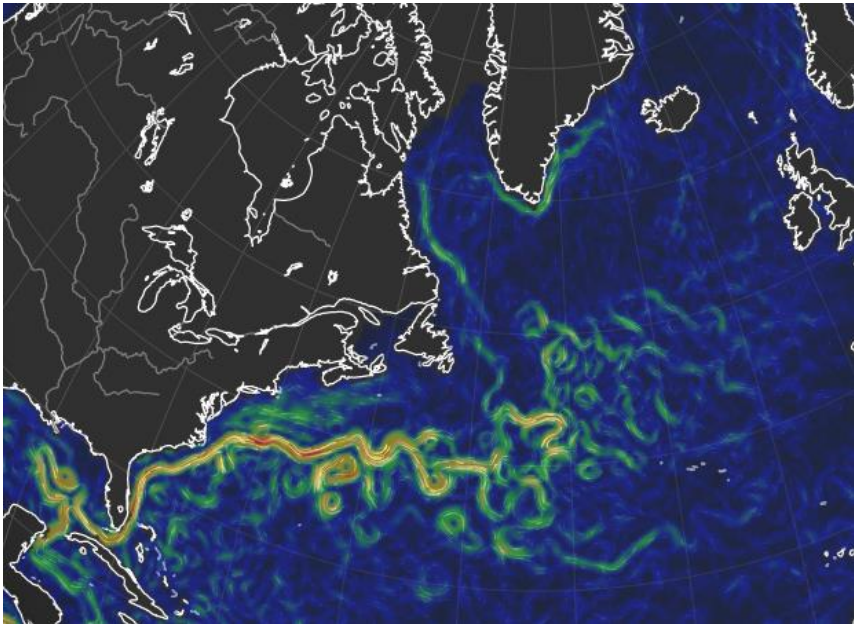
Atmosphere

Ocean

H	$\sim 10^4 \text{ m (troposphere)}$	$\sim 500 \text{ m (thermocline)}$
g'	$g \frac{\Delta \theta}{\theta_0} \approx 0.1 \text{ m s}^{-2}$	$g \frac{\Delta \rho}{\rho_0} \approx 0.02 \text{ m s}^{-2}$
$c = \sqrt{g'H}$	$\sim 30 \text{ m s}^{-1}$	$\sim 3 \text{ m s}^{-1}$
$L_d = c/f$	$\sim 300 \text{ km}$	$\sim 30 \text{ km}$

Order-of-magnitude values for $f \approx 10^{-4} \text{ s}^{-1}$ (mid-latitudes).

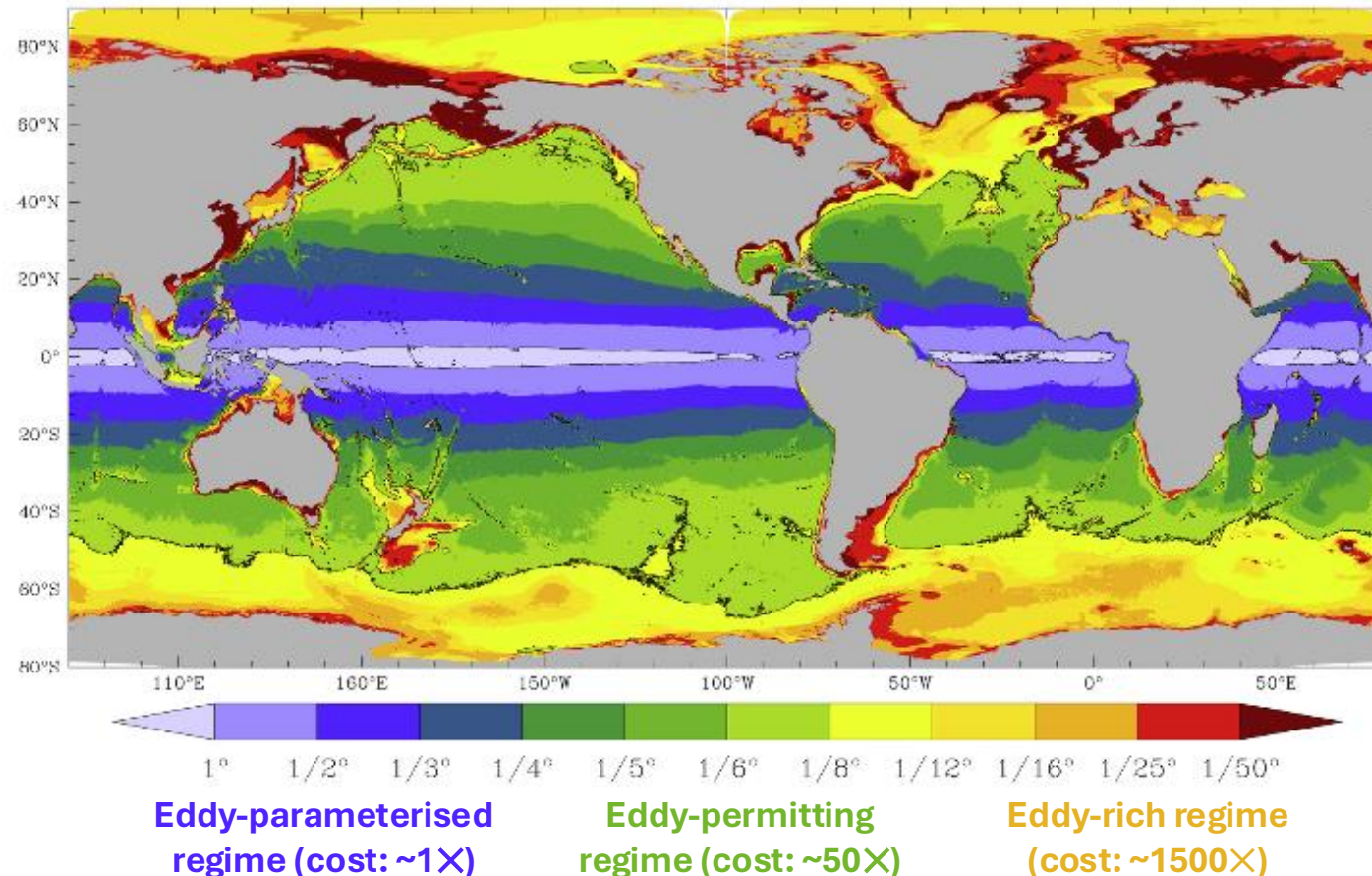
Snapshot of surface currents



Ocean models need to resolve L_d to accurately simulate the mesoscale

- The Rossby radius of deformation (L_d) decreases in the higher latitudes and coastal regions.
- **Eddy-permitting models** ($\Delta x \approx 25$ km) are sufficient to resolve mesoscale features in the low latitudes, but they provide an inaccurate or incomplete representation of the mesoscale in the mid- and high-latitudes.
- **Eddy-rich models** ($\Delta x \approx 10$ km) resolve the ocean mesoscale over the majority of the open oceans, but even higher resolutions are required to fully resolve L_d in the high latitudes and shallow coastal regions.

Resolution required to resolve L_d by $2 \Delta x$

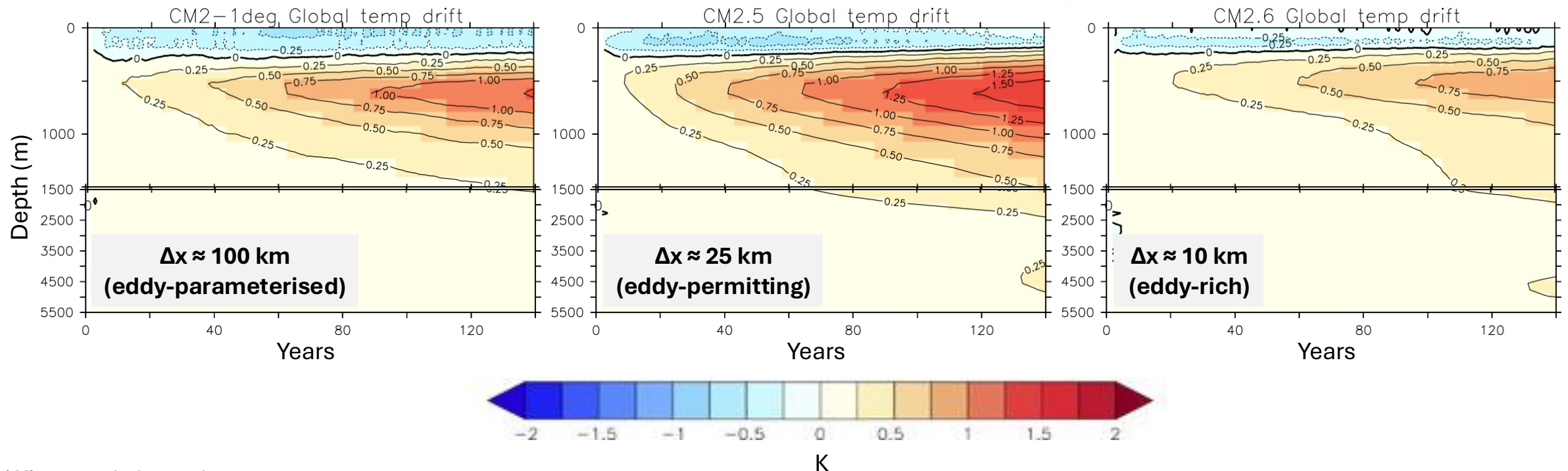


$$L_d = \frac{\sqrt{g'H}}{f}$$

Ocean eddies are crucial for realistic ocean budgets

- At climate timescales, an accurate representation of ocean eddies is crucial for a realistic simulation of ocean volume, heat, and salt transports and associated air-sea interactions and atmospheric feedbacks.
- Eddy parameterisations (e.g. Gent-McWilliams) improve biases but miss the inherently transient, nonlinear, and spatially heterogeneous nature of mesoscale eddy-mean flow interactions.
- Eddy-permitting models can drift more than coarse models because eddies are neither fully resolved nor parameterised. However, these deficiencies can be mitigated at daily to seasonal lead times through accurate initialisation.

Drift in global mean temperature as a function of depth in GFDL climate models (same atmosphere)

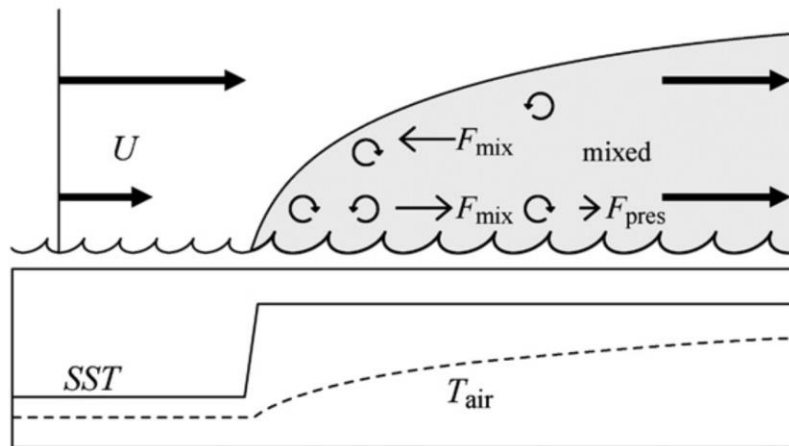


Mid-latitude air-sea interaction is scale-dependent

At the 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**.

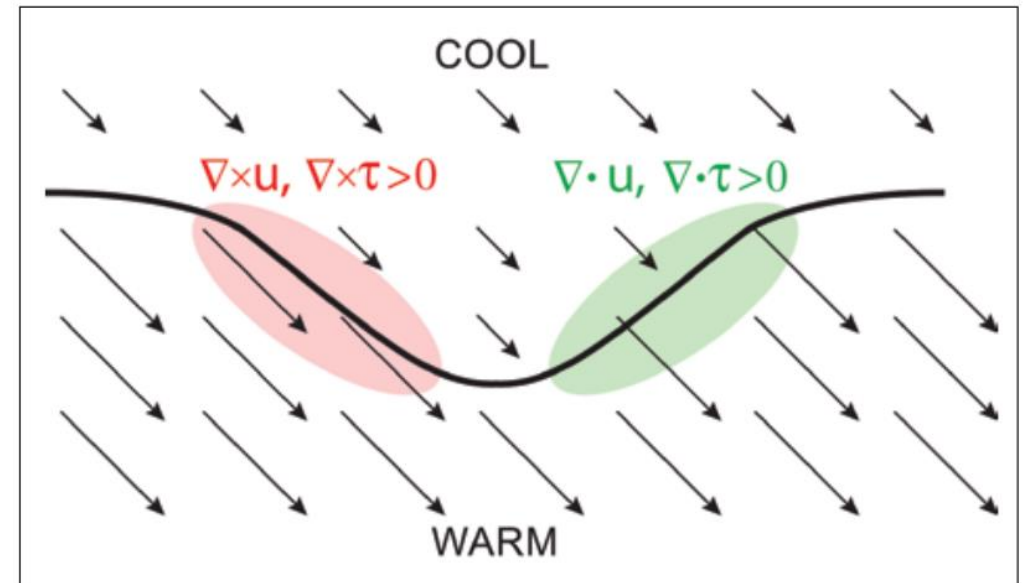
At the scale of ocean fronts and eddies, there is positive correlation between SST and wind speed/heat fluxes out of the ocean: **Warmer SSTs** → **unstable boundary layer** → **stronger surface winds** → **ocean driving atmosphere**.

(a) Strong wind



Air flowing from cold to warm water over an SST front experiences enhanced mixing and downward transfer of momentum.

Small et al. (2008)



Winds blowing across SST fronts drives near-surface divergence. Winds blowing along SST fronts results in wind stress curl.

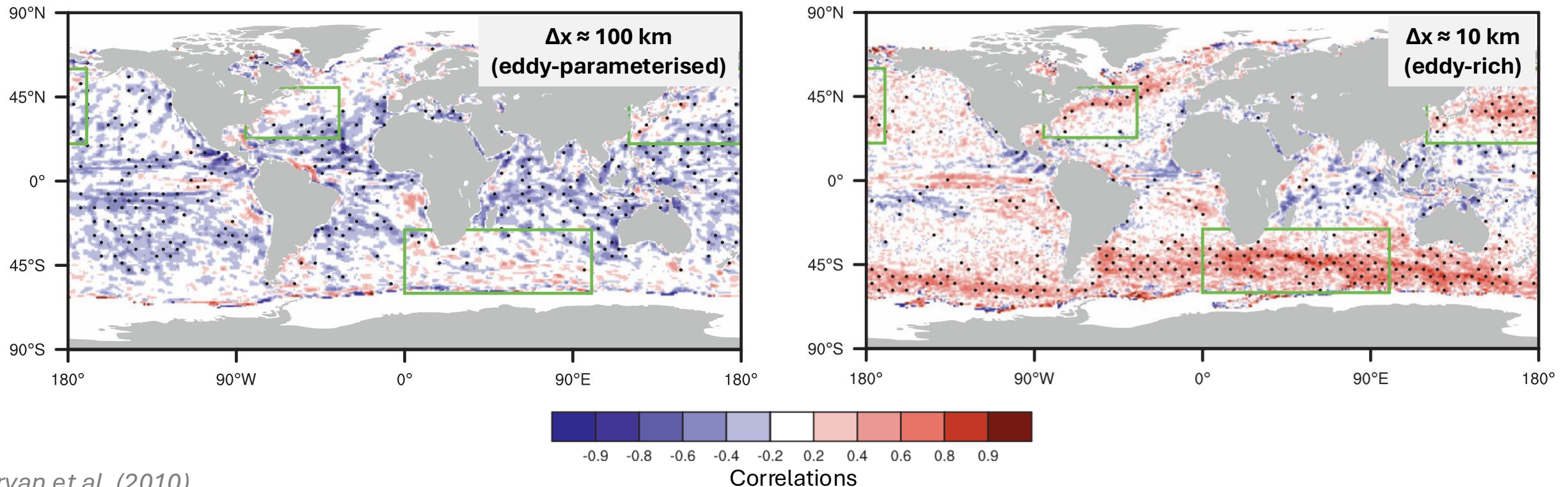
Chelton and Xie (2010)

Eddy-parameterised models miss scale-dependent air-sea interactions

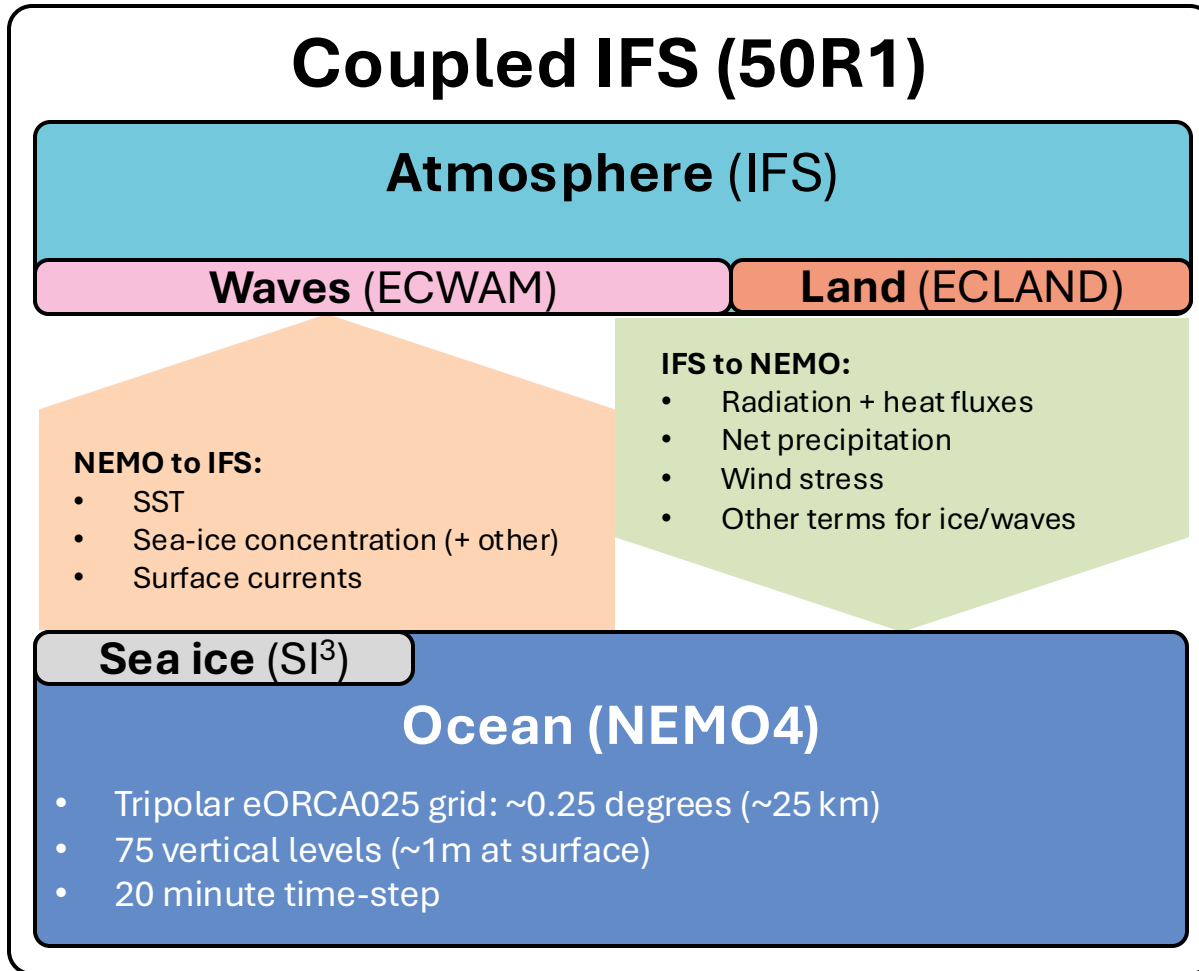
Higher resolution ocean configurations simulate more intense mesoscale air-sea interaction in regions of high eddy activity, such as **western boundary currents** and the **Antarctic Circumpolar Current**.

- **Positive correlations:** ocean driving a local atmospheric response.
- **Negative correlations:** atmosphere driving the ocean.

Temporal correlation of spatially high-pass filtered surface wind speed with SST in CCSM



All operational IFS forecasts are coupled to ocean and sea ice models



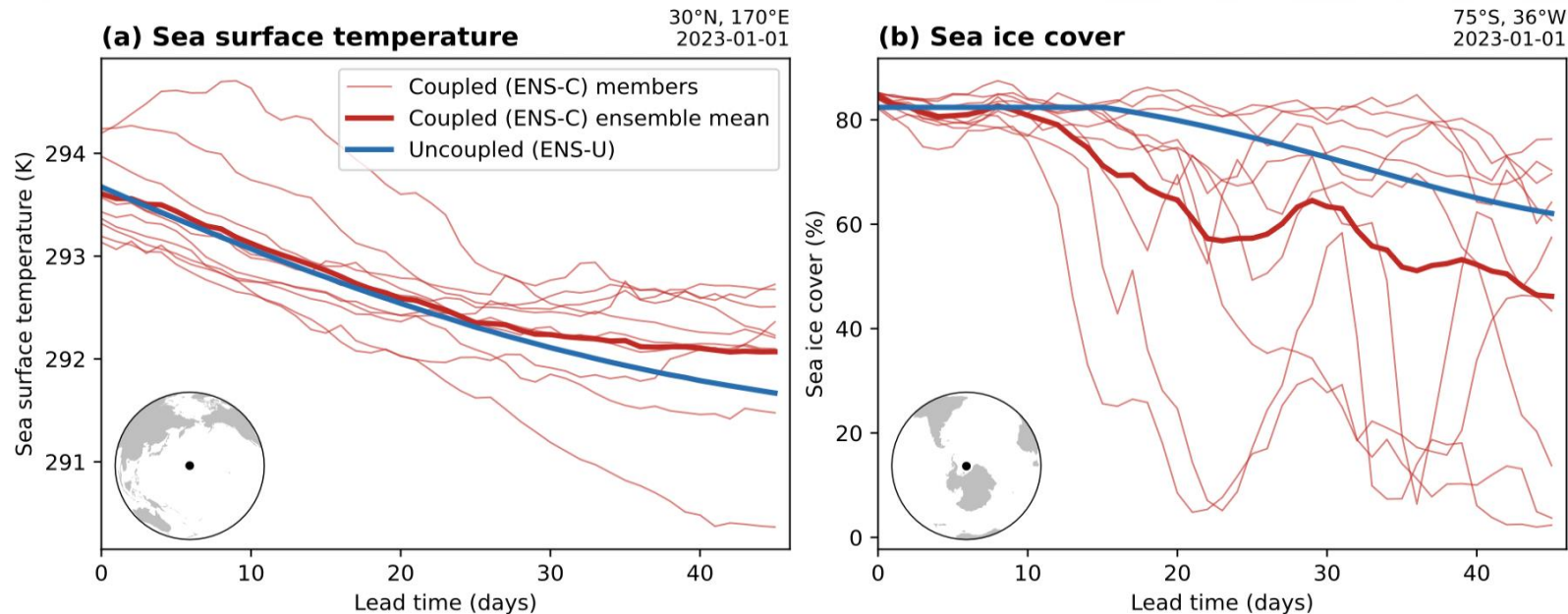
All operational configurations of the ECMWF Integrated Forecasting System (IFS) include dynamic representations of the ocean, atmosphere, sea ice, land surface, and ocean waves.

The ocean (NEMO4) and sea-ice (SI³) components of the IFS Cycle 50R1 use the same eddy-permitting eORCA025 tripolar grid.

System	Atm Res	Ocn Res	Max Lead time
Medium range	~9 km	~25 km	15 days
Sub-seasonal	~35 km	~25 km	46 days
Seasonal	~35 km	~25 km	13 months

When and where do we see the impacts of ocean coupling?

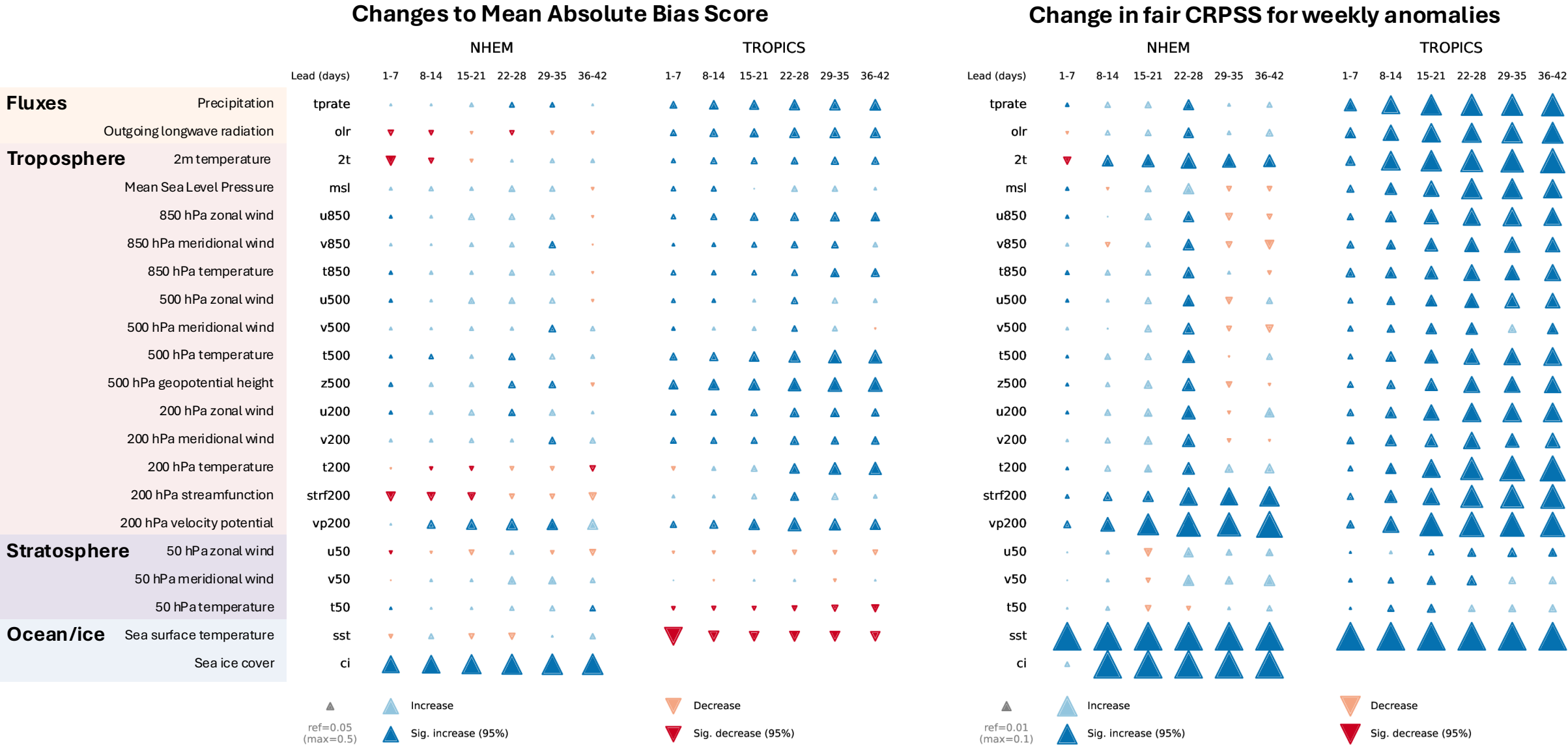
	ENS-C (Coupled)	ENS-U (Uncoupled)
Atmosphere	IFS Cycle 49R2, Tco319 L137 (~35 km)	
SSTs	NEMO4 eORCA025_Z75 (~25 km)	Climatology + persistence of initial SST anomaly
Sea ice cover	SI3 eORCA025_Z75 (~25 km)	Day 1-15: persistence of initial ice cover. Day 16-45: blends persistence → climatology. Beyond day 45: ice cover climatology.
Initial conditions	ERA5, OCEAN6	ERA5
Reforecast	10+1 members, 46 days from 1 st of each month, 216 start dates, 2006-2023	



Impact of ocean coupling in subseasonal reforecasts

Reforecast configuration

- 10 perturbed members
- 46 days
- 216 start dates, 2006-2023



Impact of coupling on 200 hPa temperature anomalies

- Reforecast configuration**
- 10 perturbed members
 - 46 days
 - 216 start dates, 2006-2023

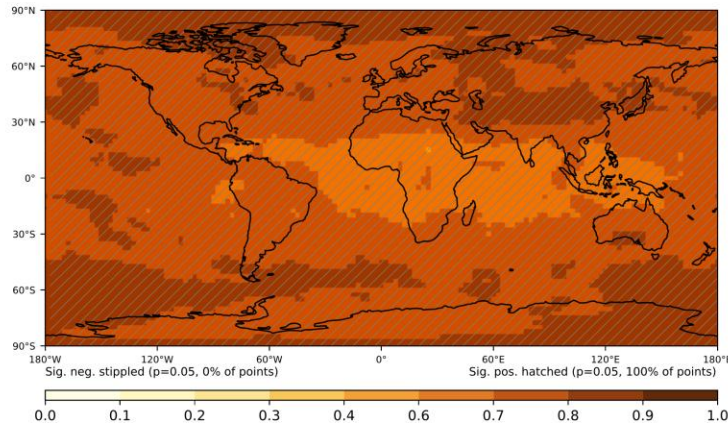
Fair version of the continuous ranked probability skill score (fCRPSS)

Uncoupled

FCRPSS: Temperature at 200hPa (vs EA12)

impq1: 20060101-20231201

PERIOD: 0-168



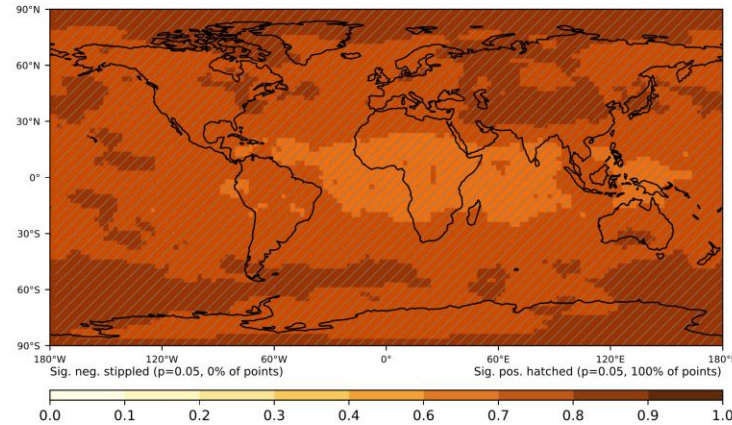
Week 1

Coupled

FCRPSS: Temperature at 200hPa (vs EA12)

impq: 20060101-20231201

PERIOD: 0-168



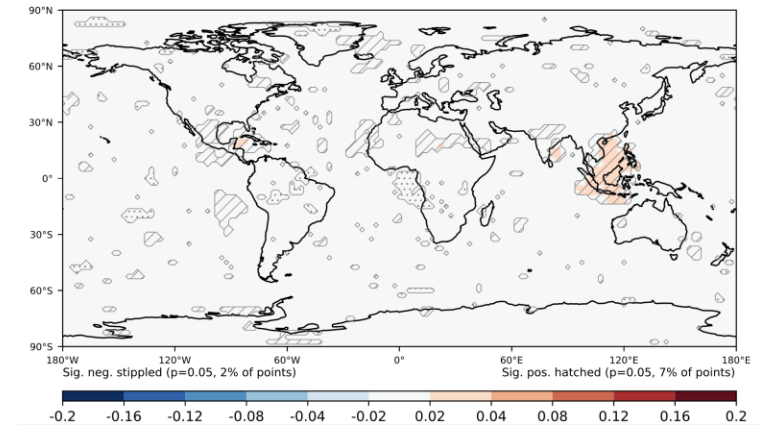
Δ fCRPSS for weekly anomalies

Coupled minus Uncoupled

Δ fCRPSS: Temperature at 200hPa (vs EA12)

impq - impq1: 20060101-20231201

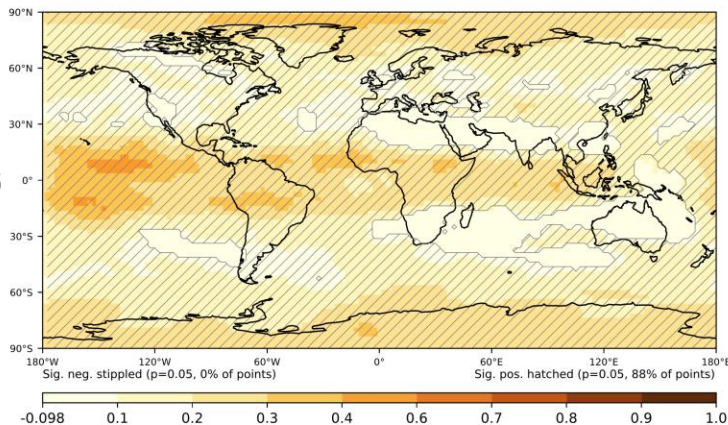
PERIOD: 0-168



FCRPSS: Temperature at 200hPa (vs EA12)

impq1: 20060101-20231201

PERIOD: 840-1008

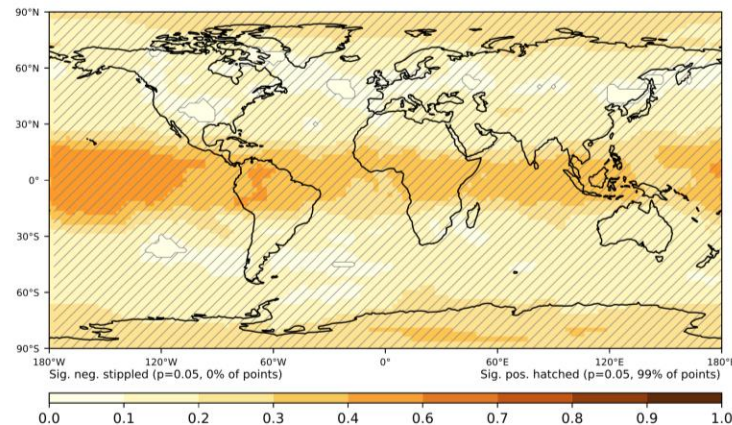


Week 6

FCRPSS: Temperature at 200hPa (vs EA12)

impq: 20060101-20231201

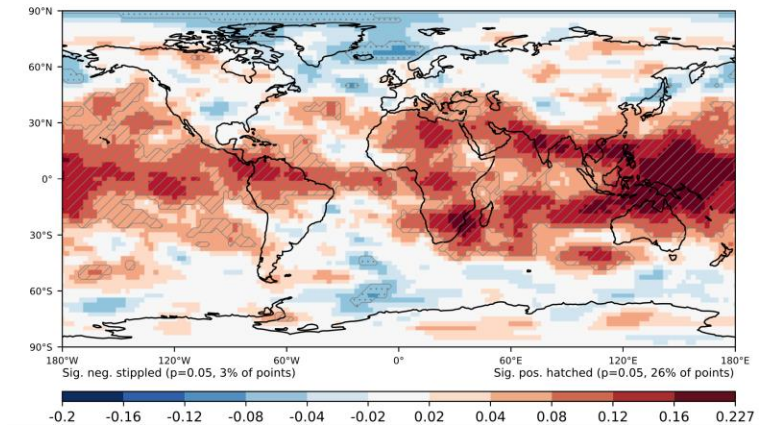
PERIOD: 840-1008



Δ fCRPSS: Temperature at 200hPa (vs EA12)

impq - impq1: 20060101-20231201

PERIOD: 840-1008



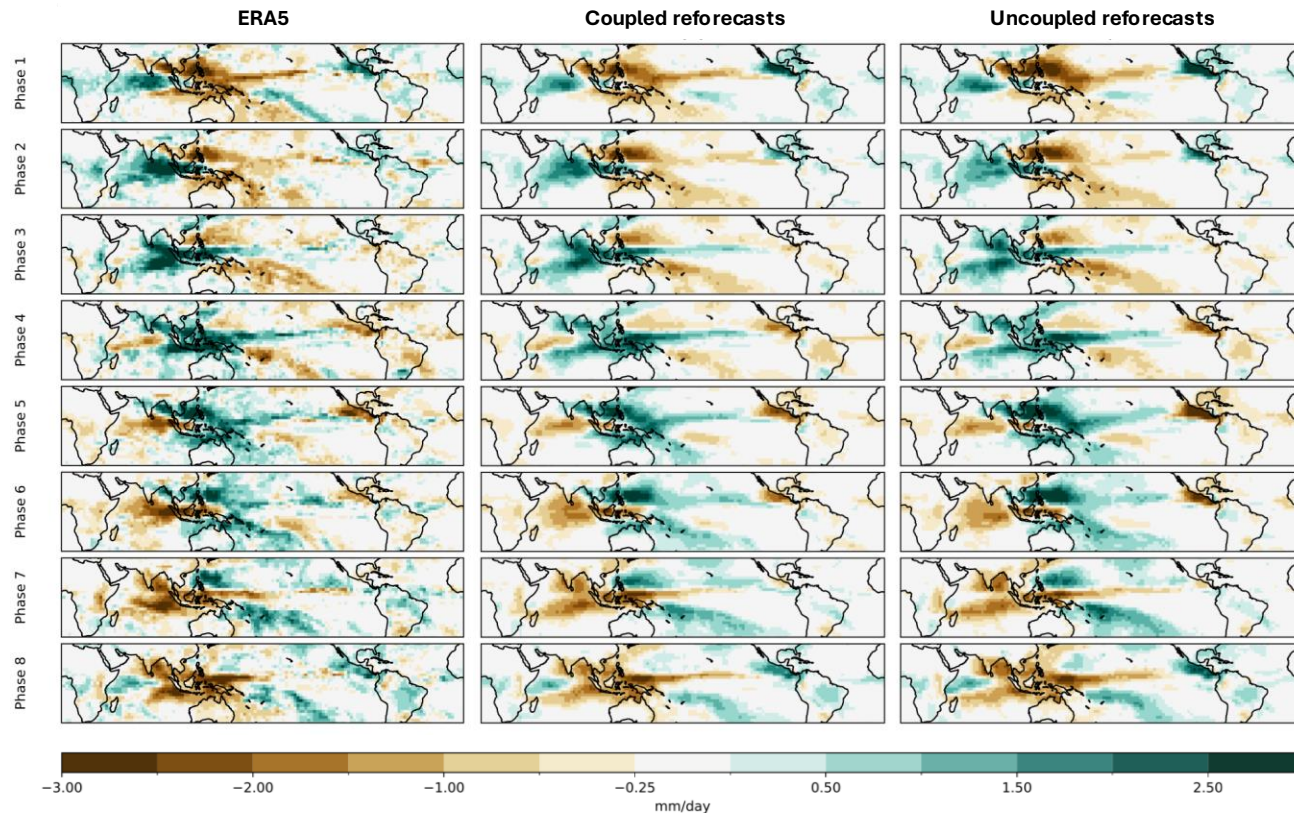
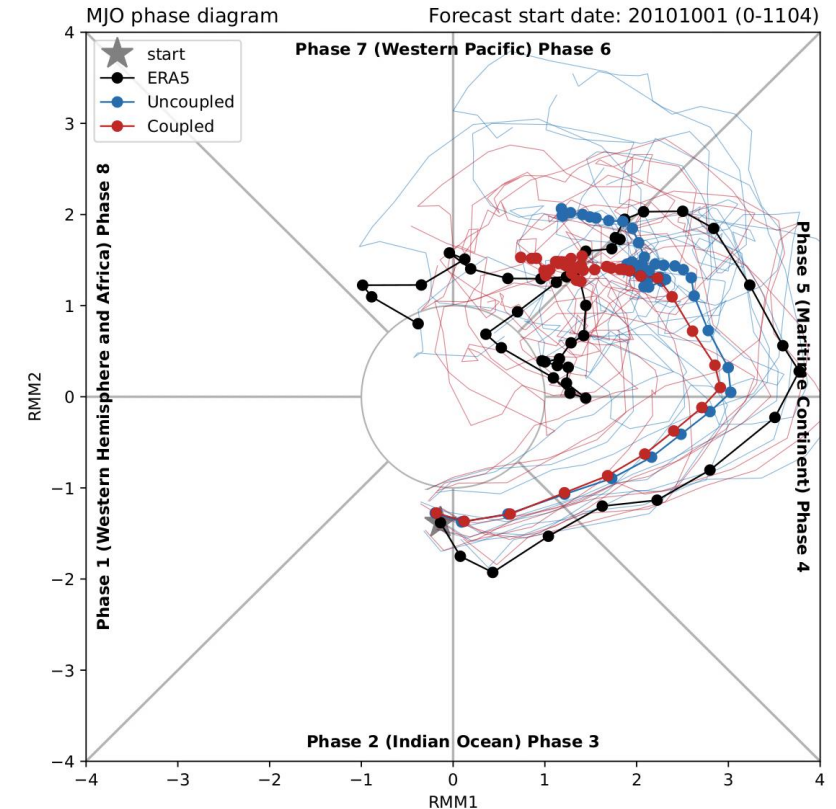
Improved with ocean coupling
Degraded with ocean coupling

Impact of ocean coupling on the Madden-Julian Oscillation (MJO)

The Madden-Julian Oscillation (MJO) has a timescale of ~30-60 days and is the leading mode of intraseasonal variability in the tropics associated with eastward propagating circulation and precipitation anomalies.

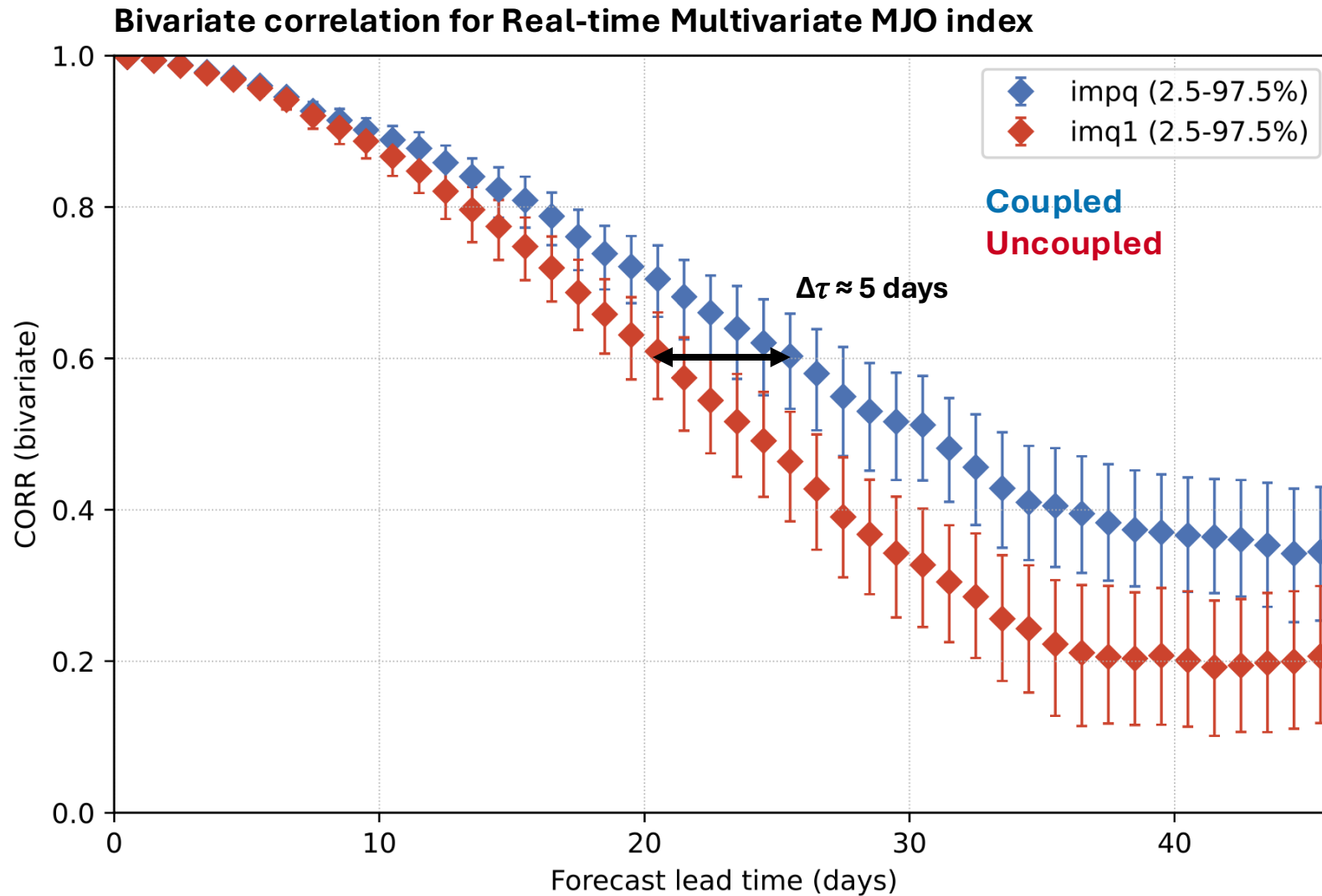
Both reforecasts simulate realistic MJO propagation and associated precipitation anomalies. It is difficult to discriminate between “coupled” and “uncoupled” simulations in case studies.

MJO Total Precipitation Composites (Days 1-46, lag=0)



Also see Woolnough et al. (2007), Kim et al. (2010), Klingaman and Woolnough (2014), and DeMott et al. (2015).

Ocean coupling improves the skill of MJO predictions



Averaged over many cases, forecast skill is improved with ocean coupling.

Predictability horizon (CORR > 0.6) extended by ~5 days.

In ECMWF model, improved skill is associated with reduced phase bias ($\approx$ faster propagation).

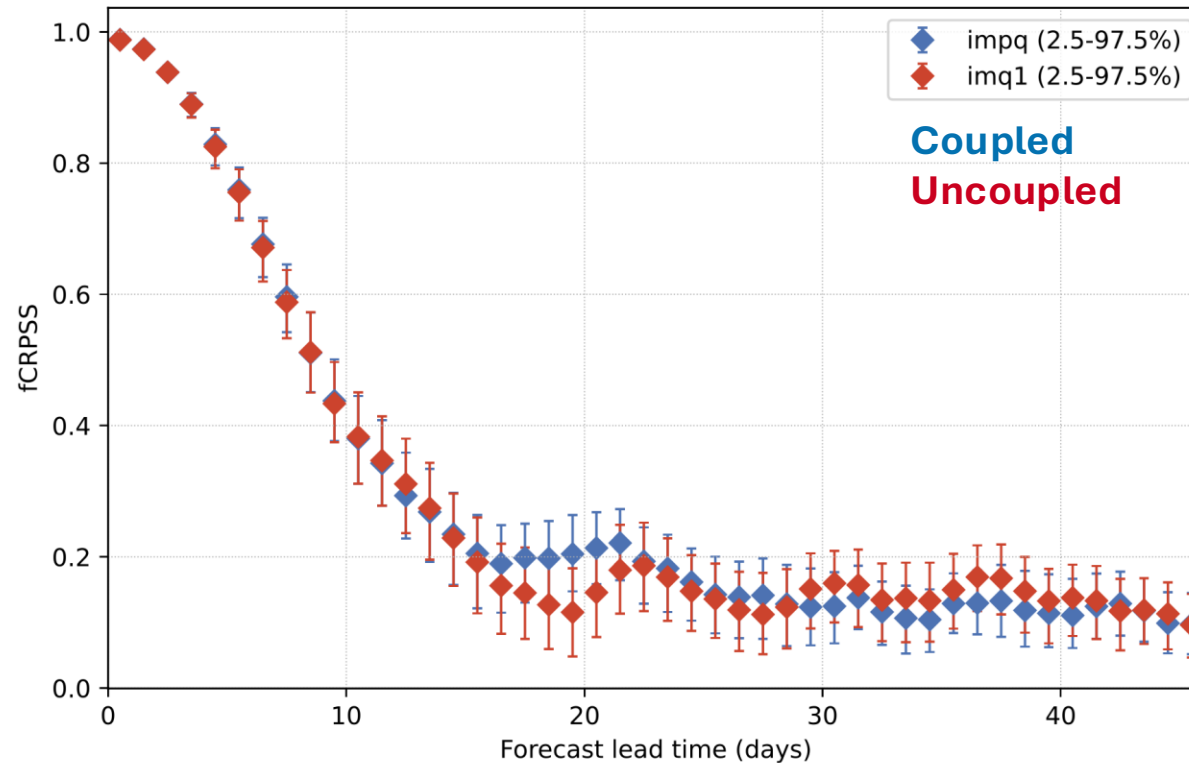
Previous studies have emphasised the importance of coupling for more realistic interactions between convection, moisture advection, ocean mixing, and SSTs (e.g. *DeMott et al. 2015*).

Ocean coupling has a limited impact on extratropical circulation indices

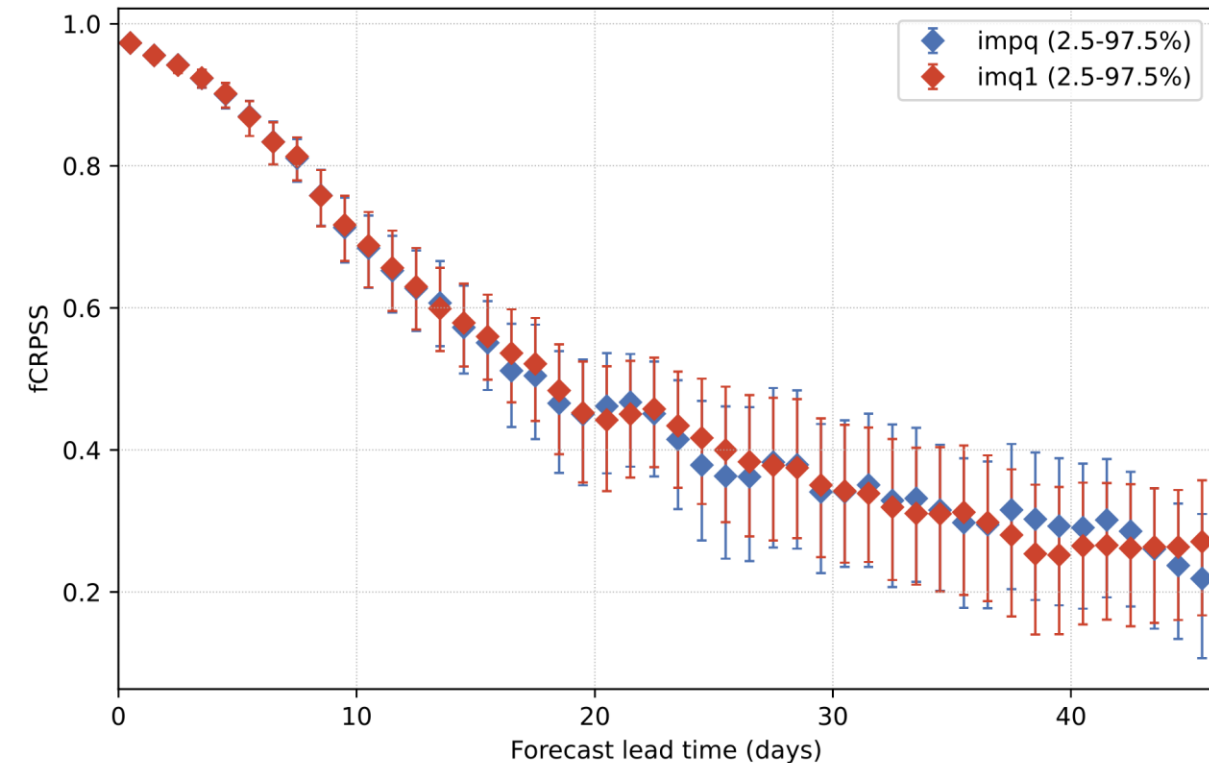
Sea surface temperature and sea-ice anomaly forecasts are dramatically improved by coupling. However, these improvements at the ocean/atmosphere interface do not translate to statistically robust impacts on atmospheric circulation indices at S2S lead times in the ECMWF coupled model.

Fair version of the continuous ranked probability skill score (fCRPSS)

500 hPa North Atlantic Oscillation (NAO) index



Zonal mean zonal winds at 50hPa and 60°N



Insights from eddy-rich subseasonal reforecasts

Free-running coupled climate models show clear benefits from eddy-rich oceans. *Do these benefits carry over to initialised subseasonal predictions, where biases have less time to develop?*

Experiment design:

- Single-member reforecasts with **eddy-rich** and **eddy-permitting** ocean at different atmosphere resolutions.
- All initialised from the same eddy-rich ocean/sea-ice analysis (GLORYS12).

Caveats:

- Deterministic evaluation of 264 start dates limits statistical power. Ensembles would be very expensive!
- Double-penalty effects can penalise higher resolution forecasts (i.e. more realistic features, but in the wrong place or at the wrong time).

Experiment	Atmosphere	Ocean/ice	Members	Length (days)	Period	Initial conditions
LRA-EPO	IFS 48r1 (~35 km)	NEMO4-SI ³ (~25 km)	1	32	1995–2016	ERA5/GLORYS12v1 ^b
HRA-EPO	IFS 48r1 (~9 km)	NEMO4-SI ³ (~25 km)	1	32	1995–2016	ERA5/GLORYS12v1 ^b
LRA-ERO	IFS 48r1 (~35 km)	NEMO4-SI ³ (~8 km)	1	32	1995–2016	ERA5/GLORYS12v1 ^b
HRA-ERO	IFS 48r1 (~9 km)	NEMO4-SI ³ (~8 km)	1	32	1995–2016	ERA5/GLORYS12v1 ^b

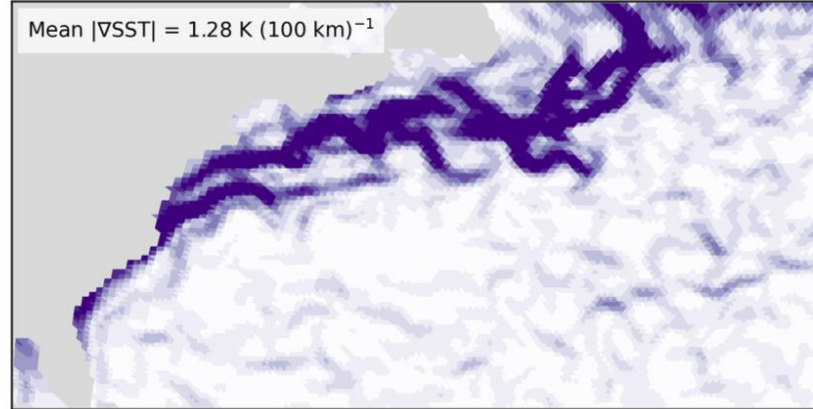
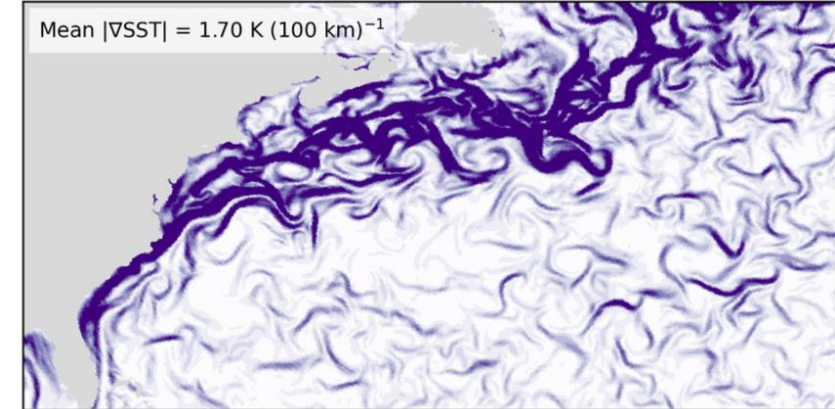
At $t=0$, atmosphere resolution can matter more than ocean resolution

$|\nabla \text{SST}|$ in the Gulf Stream region for January 1st 2016

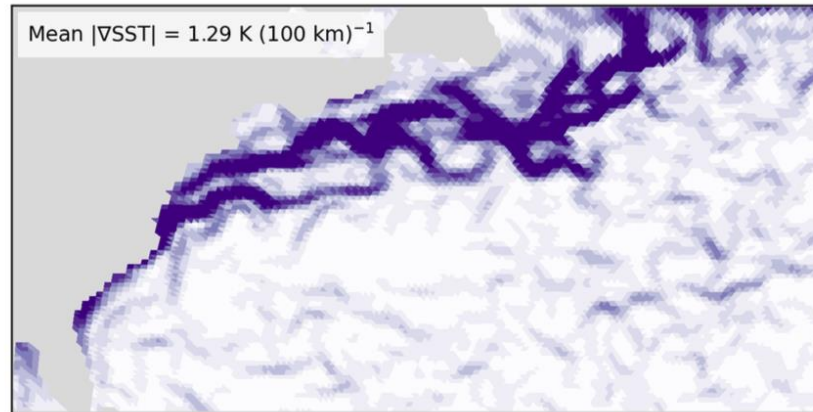
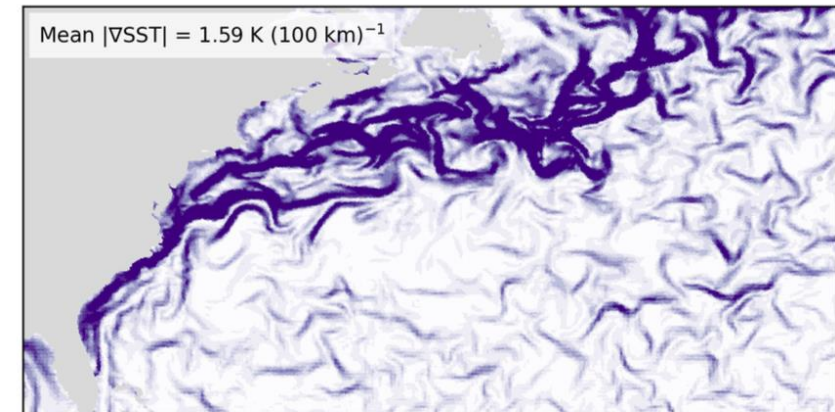
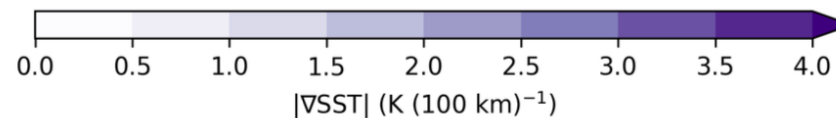
Tco319 (~35 km) atmosphere

Tco1279 (~9 km) atmosphere

Eddy-rich
(~8 km)



Eddy-permitting
(~25 km)



All configurations capture scale-dependent coupling

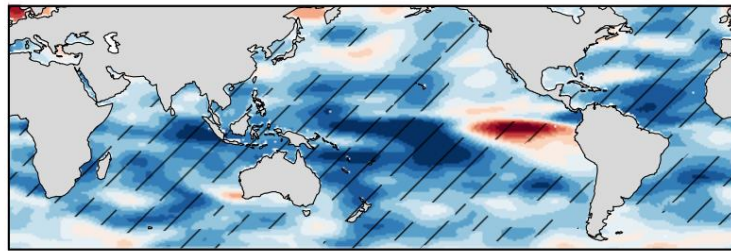
Unlike eddy-parameterised models (Bryan et al. 2010), eddy-permitting initialised forecasts capture ocean-driven, small-scale coupling (i.e. positive wind-SST correlations).

Scale-dependent correlation of 10 m wind speed and SST

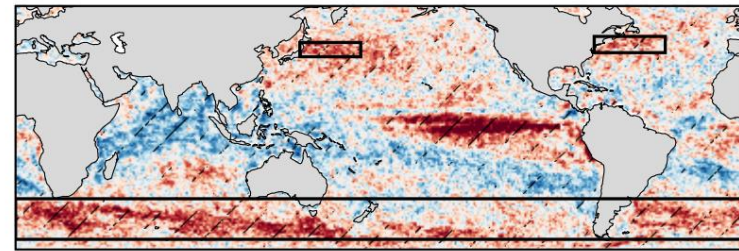
ONDJFM 2008-2016 start dates | 26 to 32 days lead time | $6^\circ \times 18^\circ$ boxcar filter

ESA-CCI SST and
scatterometer
winds

(a) Observations (low-passed)

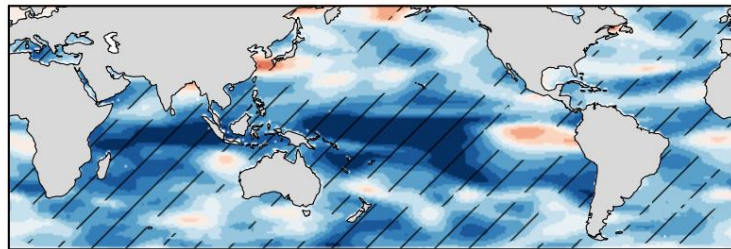


(b) Observations (high-passed)

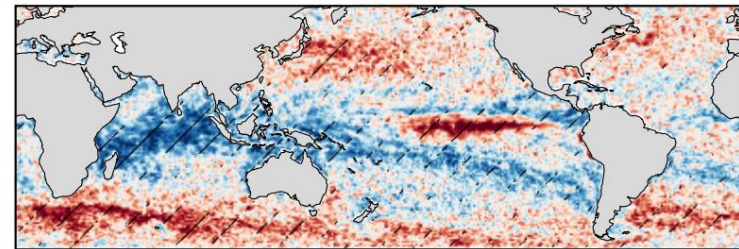


Eddy-permitting
(~25 km)

(c) LRA-EPO (low-passed)

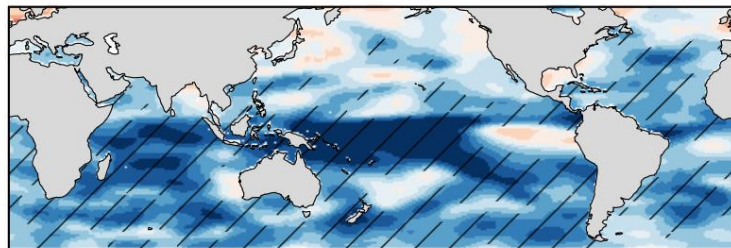


(d) LRA-EPO (high-passed)

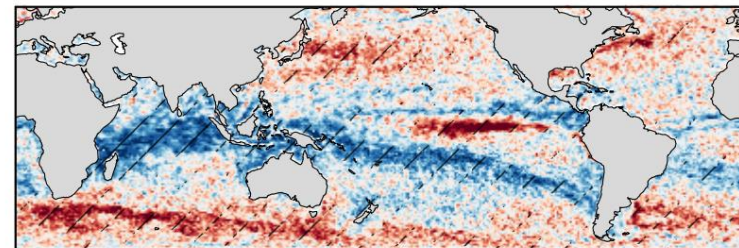
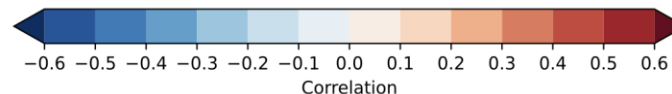


Eddy-rich
(~8 km)

(e) LRA-ERO (low-passed)



(f) LRA-ERO (high-passed)



Limited impact of an eddy-rich ocean on atmosphere at S2S lead times

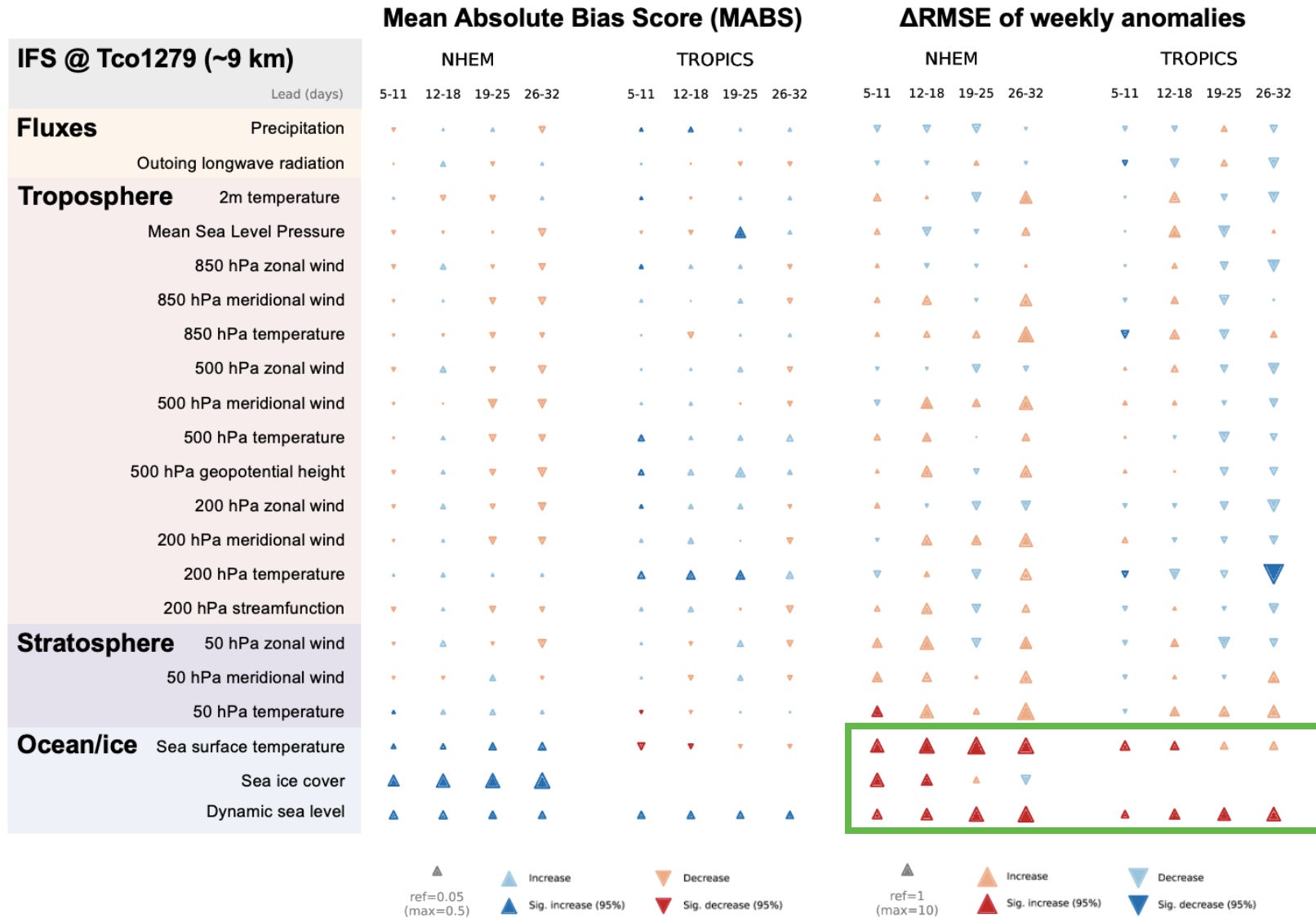


Small but statistically robust improvements to the mean state of some ocean and sea ice fields.

Some evidence for very small changes in the predictability of weekly mean ocean and sea ice. The details are sensitive to atmospheric configuration and **double-penalty effects**.

No strong evidence that an eddy-rich ocean substantially improves predictability of large-scale atmospheric fields in single-member S2S reforecasts.

Limited impact of an eddy-rich ocean on atmosphere at S2S lead times



Small but statistically robust improvements to the mean state of some ocean and sea ice fields.

Some evidence for very small changes in the predictability of weekly mean ocean and sea ice. The details are sensitive to atmospheric configuration and **double-penalty effects**.

No strong evidence that an eddy-rich ocean substantially improves predictability of large-scale atmospheric in single-member S2S reforecasts.

No consistent impact of an eddy-rich ocean on MJO forecast skill

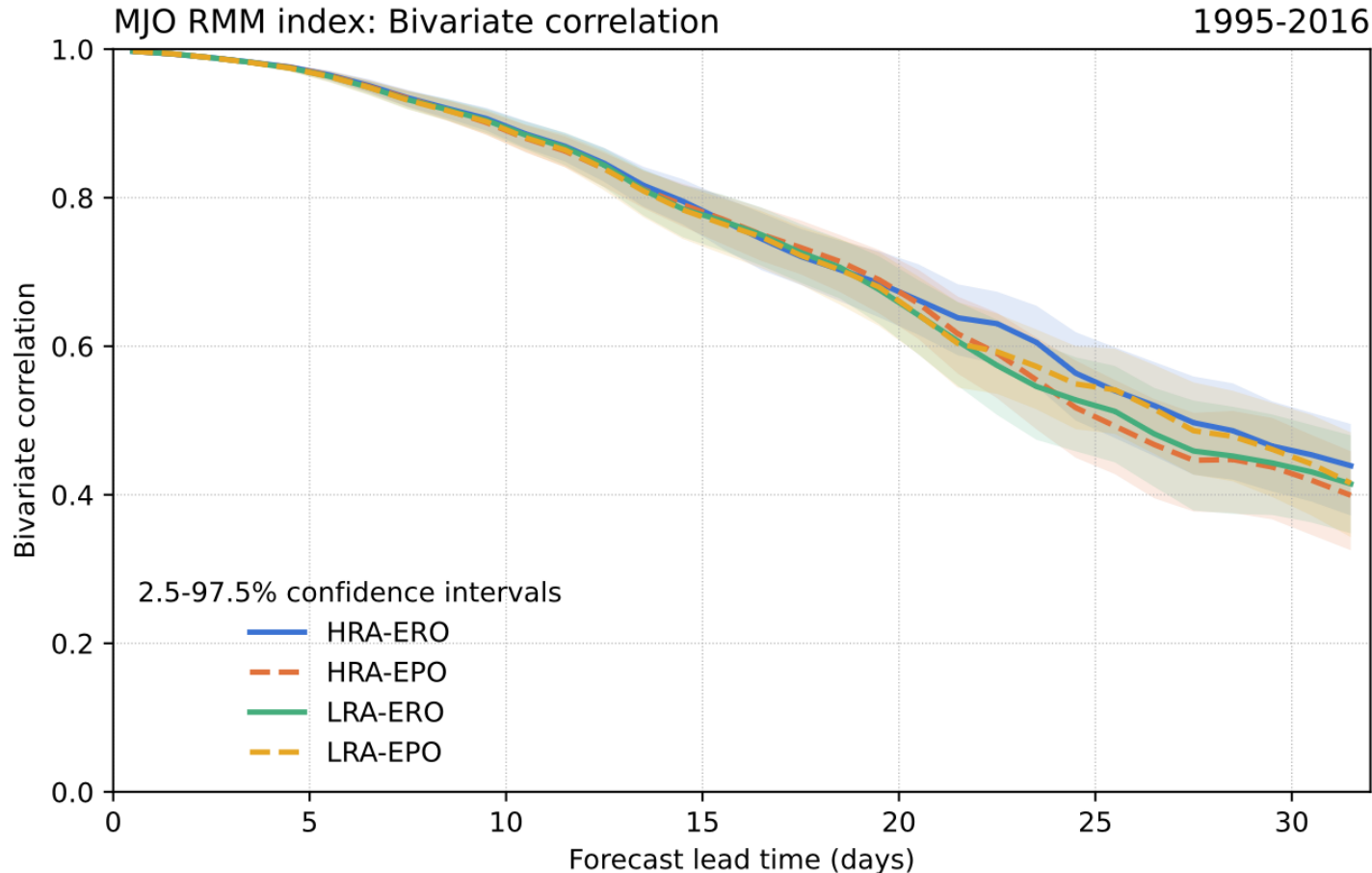


Figure R2. Bivariate correlation of the RMM index against lead time in single-member reforecasts with eddy-rich (solid lines) and eddy-permitting (dashed lines) oceans coupled to Tco1279 (HRA-ERO, HRA-EPO) and Tco319 (LRA-ERO, LRA-EPO) atmospheres, initialised on the first of each month between 1995 and 2016. Shading shows 95% confidence intervals estimated by bootstrap resampling over start dates.

Bivariate correlations of the RMM index are almost identical in all four configurations for the first three weeks.

At longer lead times, eddy-rich reforecasts have slightly higher correlations than eddy-permitting reforecasts when coupled to the Tco1279 atmosphere, but slightly lower correlations when coupled to the Tco319 atmosphere.

However, all differences are within estimated sampling uncertainty.

Summary

1. The ocean and atmosphere share the same underlying dynamics, but ocean “weather” is much smaller and slower than its atmospheric counterpart.
2. The benefits of coupling for initialised predictions depend on the balance between improvements from more realistic air-sea interactions and the impact of systematic errors (e.g. SST biases).
3. At subseasonal lead times, the largest benefits from ocean coupling are in the tropics (e.g. MJO skill extended by ~5 days), with only marginal impacts on extratropical circulation indices.
4. In free-running climate models, eddies must be resolved or adequately parameterised for realistic ocean transports and representation of the mean state.
5. In ECMWF subseasonal forecasts, eddy-rich oceans have limited impacts on the large-scale atmospheric circulation because accurate initialisation mitigates many of the deficiencies of eddy-permitting models.
6. Nevertheless, other factors may be important for an eddy-rich ocean in coupled forecasts:
 - Impacts on weather extremes that are not captured in standard verification.
 - Higher-resolution (and more accurate) ocean and sea ice forecast products.
 - Benefits for coupled data assimilation, such as reduced representativity errors.

Snapshot of surface current speeds in the ECMWF global coupled model (IFS ~4 km, NEMO ~8 km)



References and further reading

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**Extra slides not covered in the
lecture**

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:

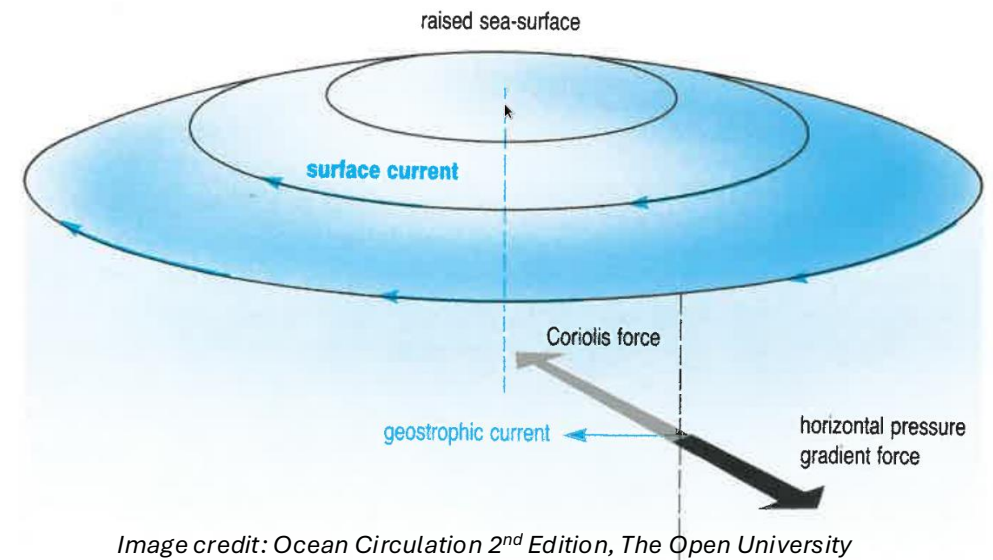
$$\left(\frac{\partial u_g}{\partial z}, \frac{\partial v_g}{\partial z} \right) = \frac{g}{f\rho_0} \left(\frac{\partial \rho}{\partial y}, -\frac{\partial \rho}{\partial x} \right)$$

- 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. $f \gg 0$).
- Inertial terms are negligible (i.e. $Ro \ll 1$).
- Away from boundary layers (i.e. friction is negligible).
- Timescale of interest $\gg (1/f)$.

Surface geostrophic currents follow sea-surface height contours.



*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 $g' \sim g / 300$.
- 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

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

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

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

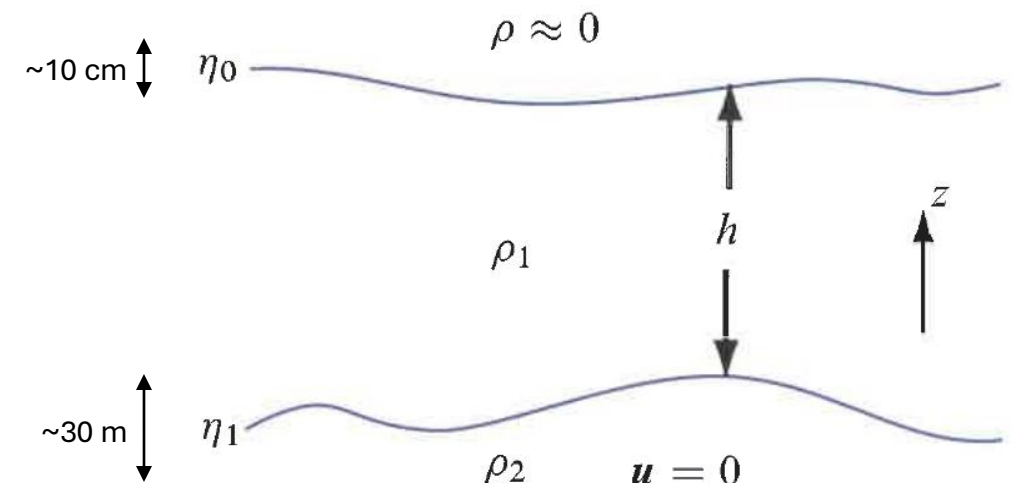
Phase speed : $c_p = \frac{\omega}{|\mathbf{k}|}$

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

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

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

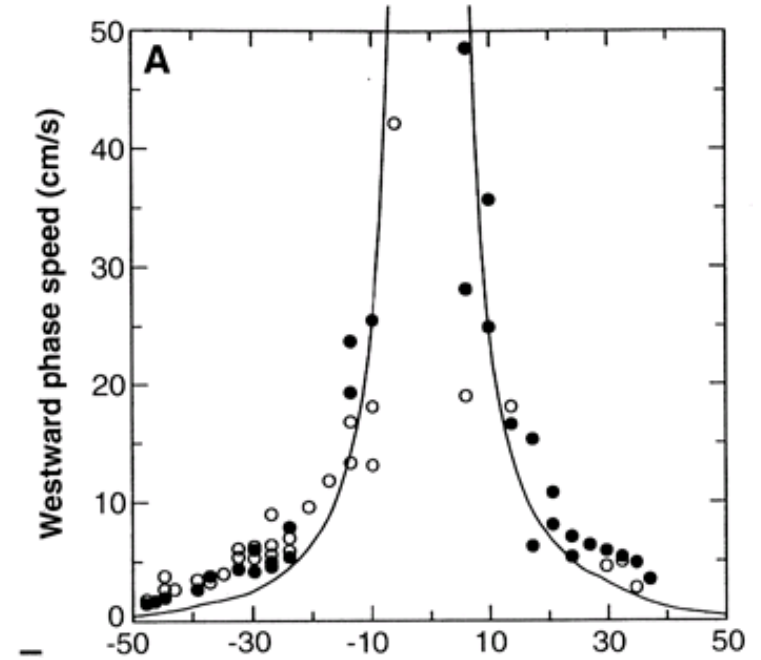
Image credit: Atmospheric and Oceanic Fluid Dynamics 2nd Edition, Vallis



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)



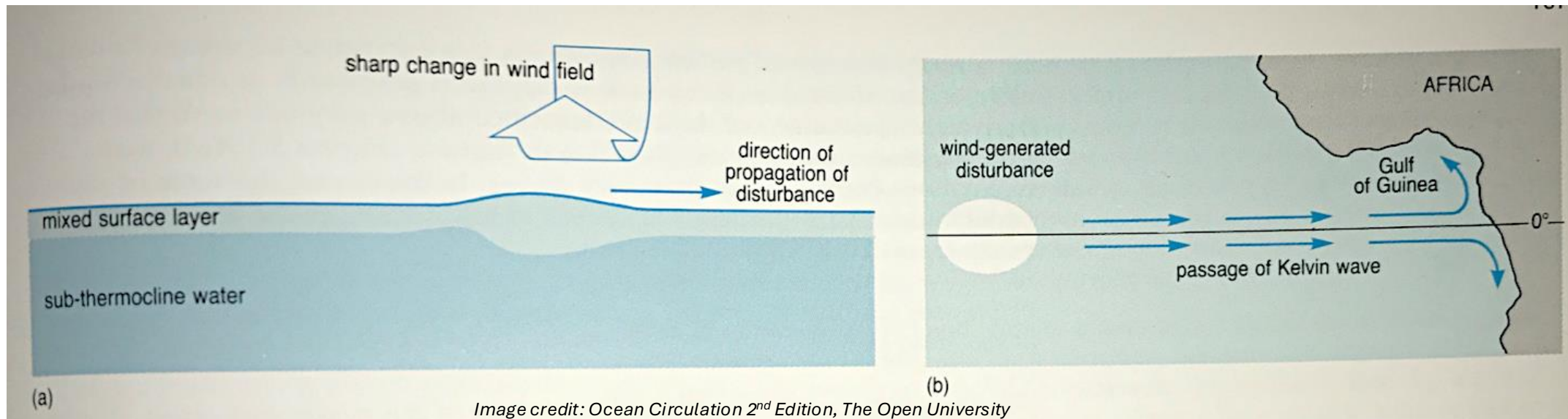
~10 years to cross the Atlantic at 40N!

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

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

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 ($c \sim 3$ m/s).
- Equatorial waves triggered by changes to the wind field are an important component of ENSO dynamics.



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

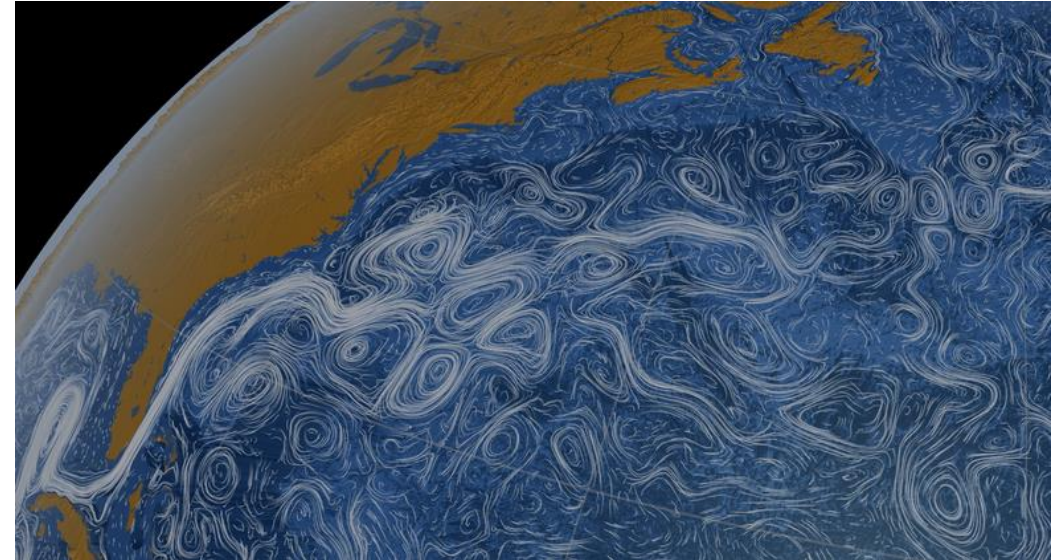


Image credit: NASA

Tropical instability waves related to shear in equatorial currents

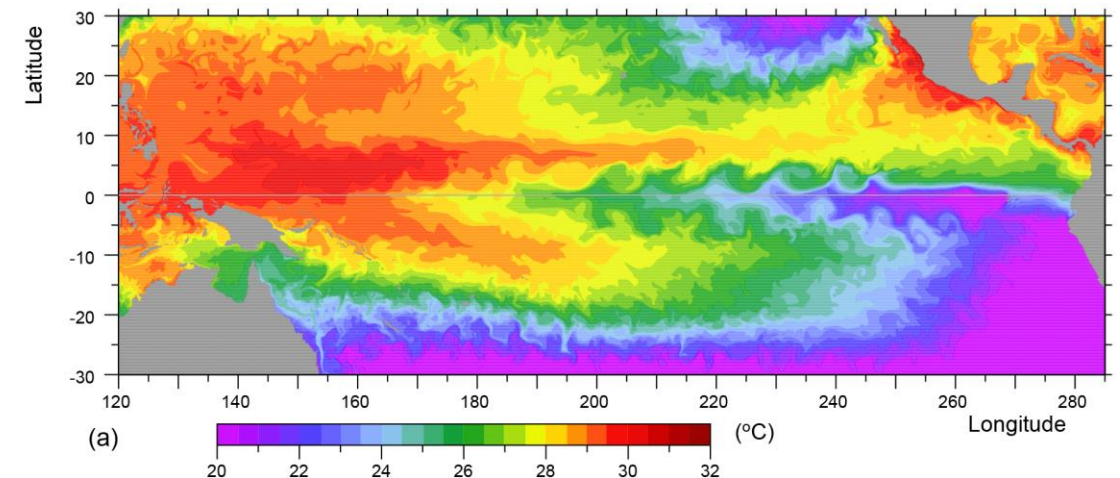


Image credit: Webb (2018), Ocean Science, vol 14.

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 ($Ro \ll 1$), 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:

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

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

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 ($\boldsymbol{\tau}$), 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.

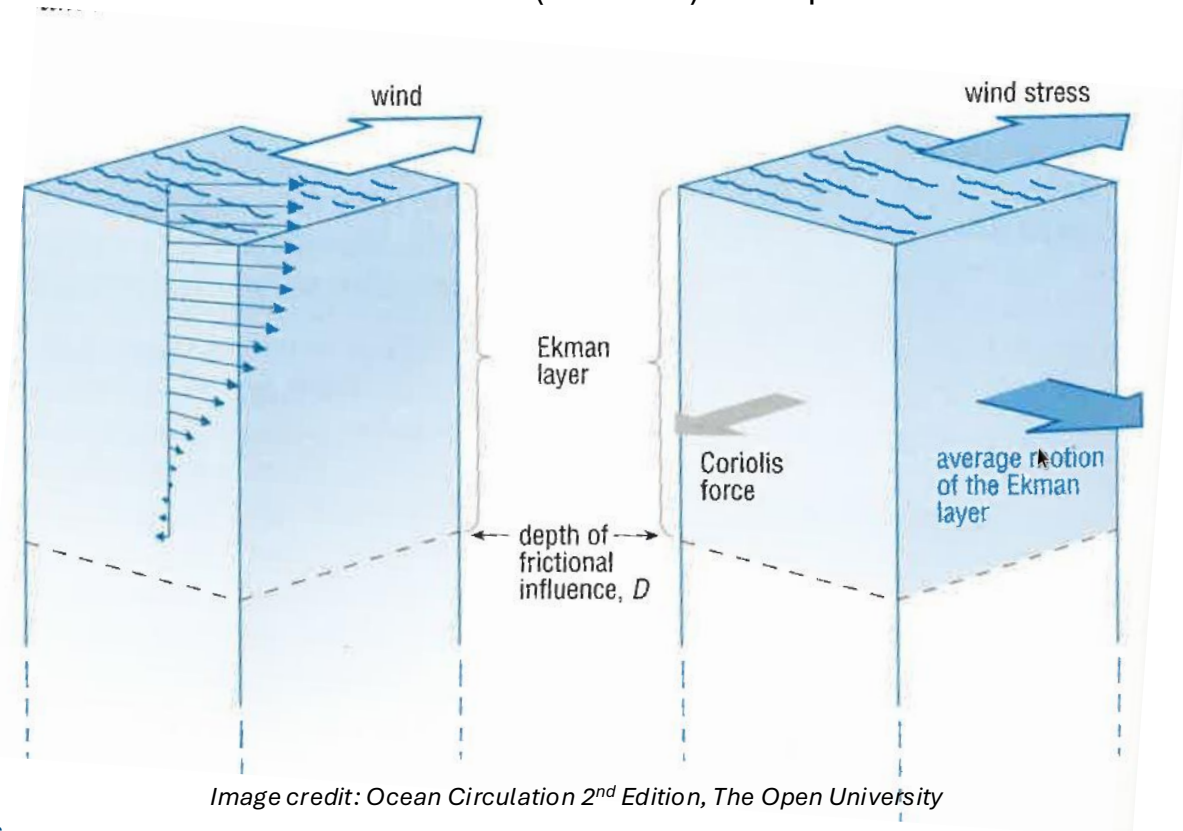


Image credit: Ocean Circulation 2nd Edition, The Open University

Extra slide 2: Ekman pumping

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

$$\nabla_h \cdot \mathbf{u} \simeq \nabla_h \cdot \mathbf{u}_{ek} = -\frac{\partial w}{\partial z}$$
$$w_{ek} = \frac{1}{\rho_0} \nabla_h \cdot \mathbf{M}_{ek} = \frac{1}{\rho_0} \nabla_h \times \frac{\tau_{wind}}{f}$$

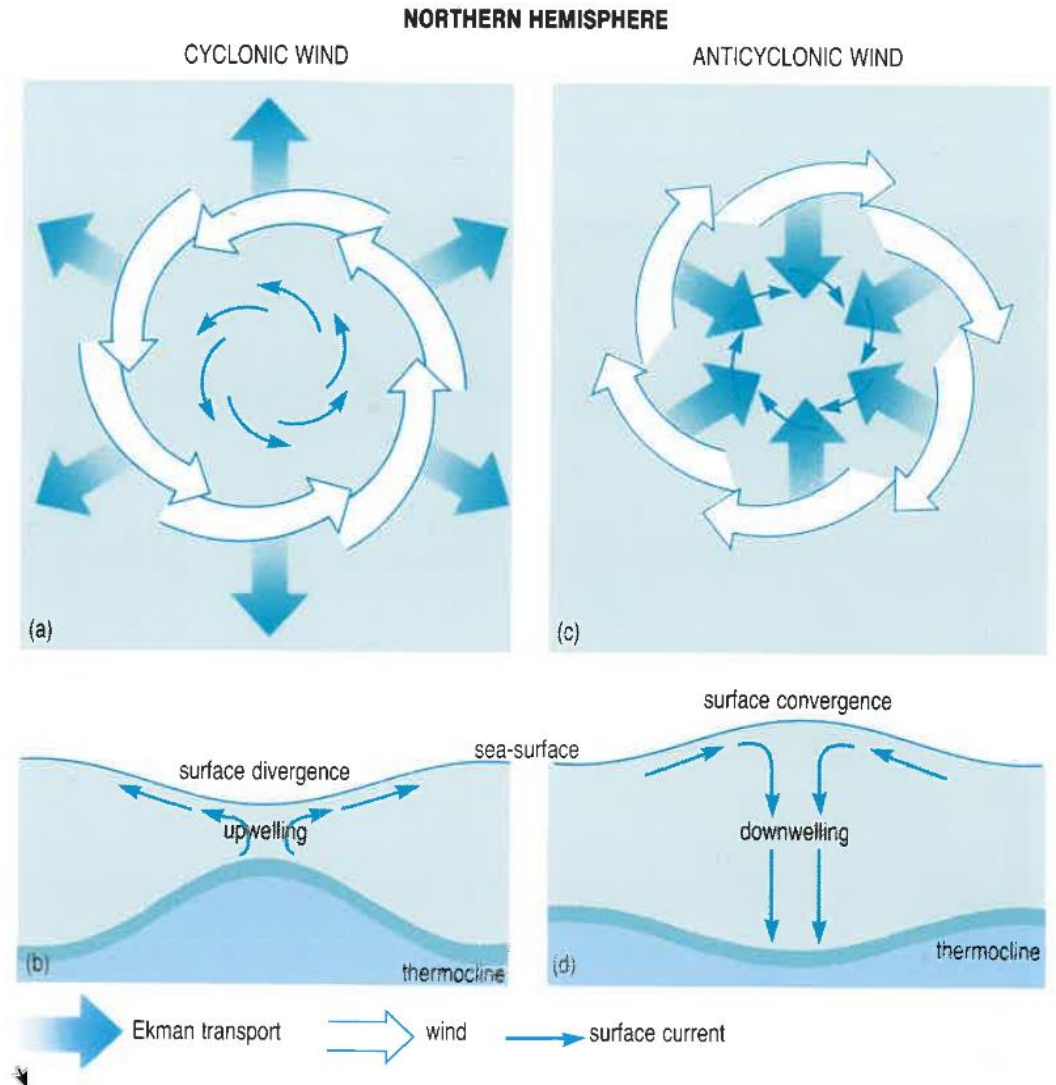
Vertical velocities are a result of wind-stress curl (and variations in f near the equator). In both hemispheres:

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

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

Assumptions:

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



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):

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

At planetary scale (i.e. $Ro \ll 1$) $f \gg \zeta$ and $PV = \frac{f}{H}$

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).

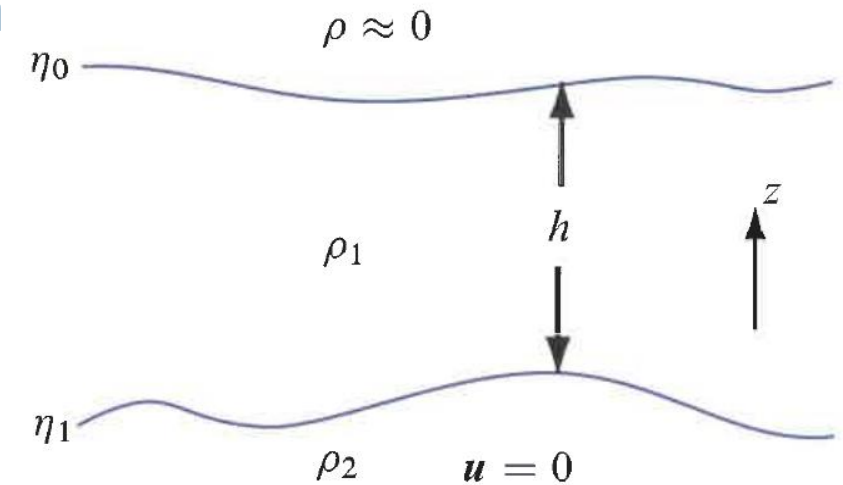
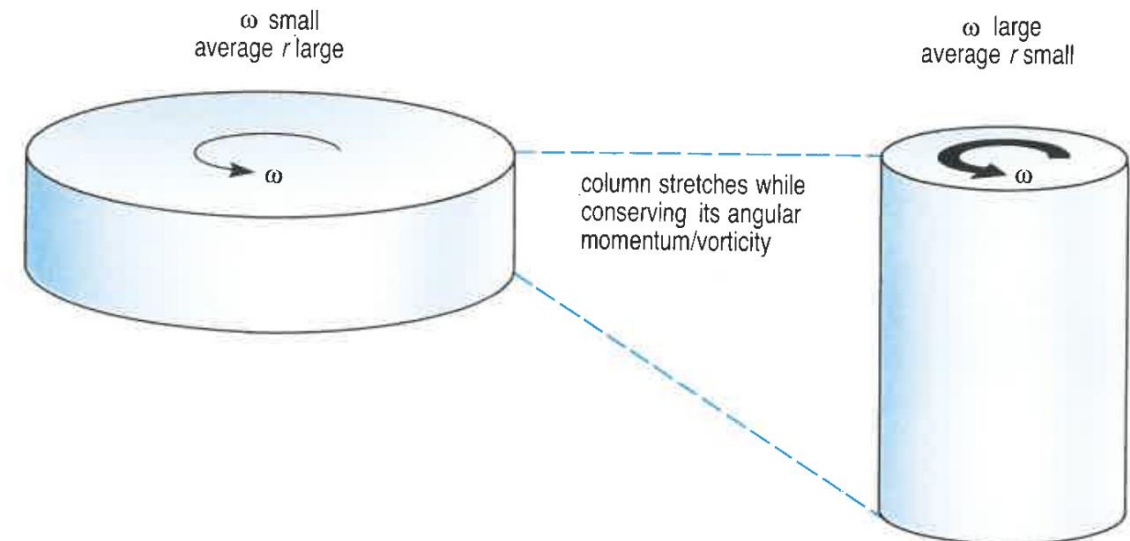


Image credit: Atmospheric and Oceanic Fluid Dynamics 2nd Edition, Vallis



(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:

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

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

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

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):

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

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

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

Substitute w_{ek} and manipulate:

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

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

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:

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

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

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

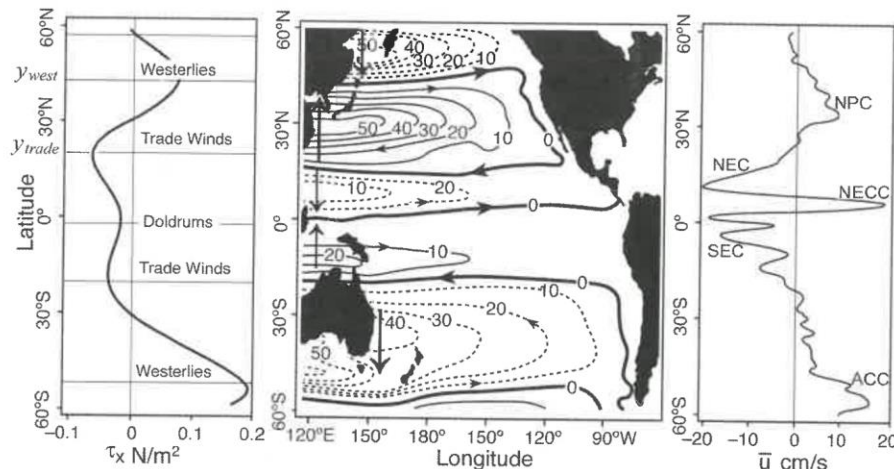


FIGURE 10.21. (left) The zonal-average of the zonal wind stress over the Pacific Ocean. (middle) The Sverdrup transport stream function (in $Sv = 10^6 \text{ m}^3 \text{ s}^{-1}$) obtained by evaluation of Eq. 10-20 using climatological wind stresses. (right) The zonal wind speed over the Pacific Ocean. The streamfunction is in $10^6 \text{ m}^3 \text{ s}^{-1}$. The wind speed is in cm/s . The streamfunction is in $10^6 \text{ m}^3 \text{ s}^{-1}$. The wind speed is in cm/s . The streamfunction is in $10^6 \text{ m}^3 \text{ s}^{-1}$. The wind speed is in cm/s .

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.

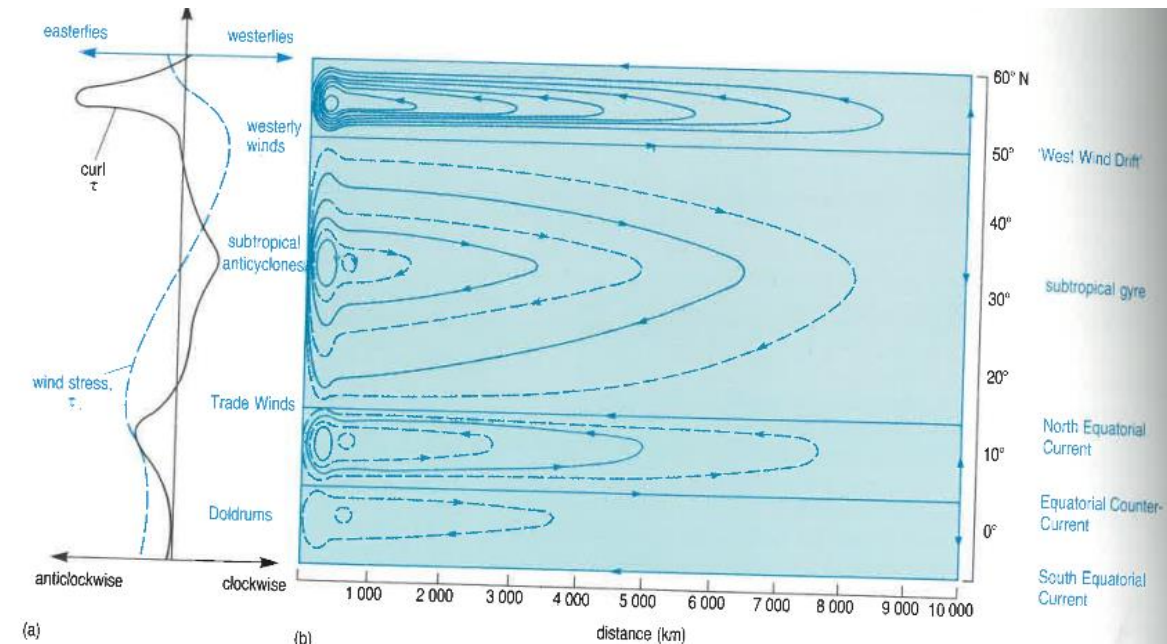


Image credit: Ocean Circulation 2nd Edition, The Open University

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 ($Ro \ll 1$), 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):

$$q = \nabla^2 \Psi + \beta y - \frac{1}{L_d^2} \Psi = \begin{array}{l} \text{relative vorticity} + \\ \text{planetary vorticity} + \\ \text{vortex stretching} \end{array}$$

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

Assumptions:

- Scale of flow similar to deformation length scale ($L_d = \sqrt{gH}/f$).
- Close to geostrophic balance ($Ro \ll 1$).
- Variations in Coriolis parameter are small ($f = f_0 + \beta y$)
- 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. $f + \zeta = \text{constant}$)

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.

