

The Ocean Heat Transport

Global Heat Budget

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Lecture Outline

- 1 Introduction & Motivation
- 2 Thermodynamic Foundation
- 3 Components of Air-Sea Heat Fluxes
- 4 Meridional Heat Transport
- 5 The Ocean Heat Content
- 6 Summary & Outlook

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The Fundamental Importance of Ocean Heat Transport

- The ocean regulates Earth's climate by moving heat from the equator to the poles (meridional heat transport).
- It sequesters heat in the subsurface, buffering the effects of climate change.
- Key Fact: The ocean transports 3 PW of heat out of the tropics or more than 50% of that accomplished by the ocean-atmosphere system, (Ferrari & Ferreira, 2011).
- Without this transport, tropical regions would be much hotter, and polar regions much colder.

The Fundamental Importance of Ocean Heat Transport ...

continuation

Why should we care?

Understanding how this heat transport responds to natural variability and climate change is a first-order question in climate science.

What Drives Ocean Heat Transport? (The Traditional View)

The ocean's heat transport is attributed to two types of circulation:

- Wind-Driven Gyres (Shallow):

- 1 Confined to the upper ocean (thermocline).
- 2 Dominant in the Pacific and Indian Oceans.
- 3 Driven by surface winds (e.g., trade winds, westerlies).

What Drives Ocean Heat Transport? (The Traditional View) ...

continuation

■ Deep Overturning Circulation (Thermohaline):

- 1 Associated with high-latitude convection (e.g., North Atlantic Deep Water formation).
- 2 Involves deep water formation and abyssal mixing.
- 3 Traditionally thought to dominate heat transport in the Atlantic.

But is this separation accurate? (Ferrari & Ferreira, 2011)

What Processes Really Matter?

- The Old Question: Is heat transport dominated by wind-driven gyres or deep, mixing-driven circulations?
- A New Perspective (Holmes et al., 2019): Focusing only on the circulation of water is problematic. Much heat simply loops around in closed gyres without being exchanged with the atmosphere.

What Processes Really Matter? ... continuation

■ The Crucial New Insight:

- 1 The ocean must move heat not just from low to high latitudes, but also from WARM to COLD temperatures (diathermal heat transport).
- 2 This "downgradient" heat transfer (warm \rightarrow cold) can only be achieved by turbulent mixing.
- 3 This mixing is the missing link connecting surface heat uptake to deep ocean circulation.

Connecting the Atlantic and Indo-Pacific

The North Atlantic transports a large amount of heat northward (e.g., 0.78 PW across 50°N). **Where does this heat come from?**

- Only 40% comes from direct heat input in the tropical Atlantic + local mixing.
- 60% (0.49 PW) is supplied from the tropical Indo-Pacific!

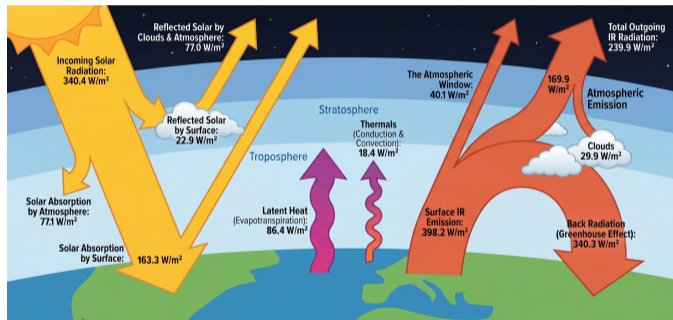
Connecting the Atlantic and Indo-Pacific ... continuation

The Pathway:

- Heat is taken up by the ocean in the eastern tropical Pacific (warm surface waters).
- Turbulent mixing transfers this heat from the warm, shallow Indo-Pacific circulation to colder, deeper water masses.
- These colder waters feed into the upper branch of the Atlantic Meridional Overturning Circulation (AMOC) and are carried northward.

The Ocean as the Planet's Thermal Reservoir

- The ocean absorbs **>90%** of excess heat from global warming
- 2.5 m of ocean depth can store as much heat as the entire atmosphere
- Air-sea heat fluxes: "valves" of the climate system



The Ocean as the Planet's Thermal Reservoir ... continuation

Key Implications:

- Heat absorbed by the ocean does not disappear; it can warm the planet for **decades** after absorption.
- This stored heat commits Earth to **at least some additional warming** in the future, even after emissions stop.
- Ocean warming raises global sea level (thermal expansion and ice melt) and threatens marine ecosystems (e.g., coral health).

Source: Climate.gov (2025). "Climate Change: Ocean Heat Content." Figure adapted from Trenberth et al. (2009), IPCC AR6, and Johnson et al. (2023).

Why Do Heat Fluxes Matter for Oceanographers?

- **Density forcing:** Controls thermohaline circulation
- **Energy for storms:** SST > 26°C fuels hurricanes

- **Water column stability:** Affects mixing and stratification
- **Meridional heat transport:** Connects tropics to poles, surface to bottom

Heat flux anomalies → Long-term climate changes

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First Law of Thermodynamics for a Fluid Parcel

$$\frac{De}{Dt} = q - w$$

■ Internal energy: $e = c_p T$

■ Heat received: $q = \frac{k_T}{\rho} \nabla^2 T$

■ Work done: $w = p \frac{DV}{Dt}$

■ c_p : specific heat at constant pressure

■ T : absolute temperature

■ k_T : thermal conductivity

■ ρ : density

■ p : pressure

■ V : volume

Equations per unit mass. For the ocean, we rewrite in terms of temperature.

Heat Conservation Equation

$$\frac{DT}{Dt} = k\nabla^2 T + F_{ext} \quad (\times \rho c_p)$$

$$\frac{\partial(\rho c_p T)}{\partial t} + \frac{\partial(u\rho c_p T)}{\partial x} + \frac{\partial(v\rho c_p T)}{\partial y} + \frac{\partial(w\rho c_p T)}{\partial z} = \mathcal{Q} + \frac{\partial}{\partial z} \left(\kappa \frac{\partial(\rho c_p T)}{\partial z} \right) + \nabla_H \cdot (A \nabla_H (\rho c_p T)) \quad (1)$$

Defining: $h = \rho c_p T$, and integrating vertically, we have $H = \int_{-D}^0 h dz$:

$$\frac{\partial H}{\partial t} + \nabla_H \cdot \int_{-D}^0 \vec{V}_H h dz = \int_{-D}^0 \mathcal{Q} dz + \kappa \frac{\partial h}{\partial z} \Big|_{z=0} - \kappa \frac{\partial h}{\partial z} \Big|_{z=D} + \nabla_H \cdot (A \nabla_H H) \quad (2)$$

Each term has an important physical meaning...

Physical Meaning of Each Term

- $\frac{\partial H}{\partial t}$ (Heat storage: the ocean's "memory")
- $Q_v = \nabla_H \cdot \left[\int_{-D}^0 \vec{V}_H h - A \nabla_H H \right] dz$ (Horiz. heat Conv.: ocean circulation)
- $\int_{-D}^0 Q dz = \int \frac{\partial F_s}{\partial z} dz = F_s(0) - \cancel{F_s(-D)}$ (Shortwave radiation: solar heating)
- $Q_{sf} = k \frac{\partial h}{\partial z} \Big|_{z=0} = -Q_{LW} - Q_{LH} - Q_{SH}$ (Losses: longwave + latent + sensible)
- $Q_D = k \frac{\partial h}{\partial z} \Big|_{-D} \leq \frac{1}{20} W m^{-2} \approx 0$ (Geothermal sources (negligible))

The Integrated Surface Budget

$$Q_{net} = Q_{SW} - Q_{LW} - Q_{LH} - Q_{SH} = \frac{\partial H}{\partial t} + \nabla_H \cdot \mathbf{F}_o$$

- Q_{net} : net flux through the air-sea interface
- $\partial H / \partial t$: local heating/cooling rate
- $\nabla_H \cdot \mathbf{F}_o$: divergence of horizontal heat transport

Important: Surface flux is not just local heating! It balances storage and transport.

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The Four Components of Surface Flux

$$Q_{net} = Q_{SW} - Q_{LW} - Q_{LH} - Q_{SH}$$

Radiative Fluxes

- Q_{SW} Shortwave (Solar) → **WARMS**
- Q_{LW} Longwave (Infrared) → **COOLS**

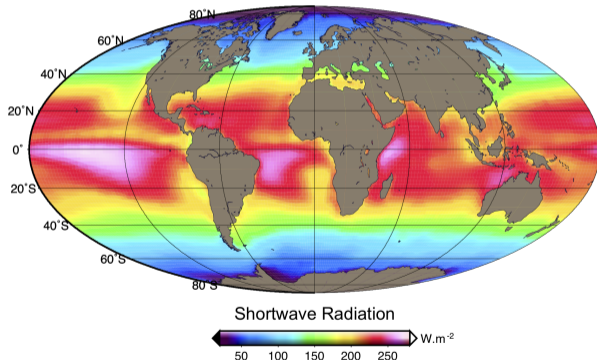
Turbulent Fluxes

- Q_{LH} Latent Heat (Evaporation) → **COOLS**
- Q_{SH} Sensible Heat (Conduction/Convection) → **CAN WARM OR COOL**

Shortwave Radiation (Q_{SW}) - The Engine

Parametric equation: A Simplistic Approach

$$Q_{SW} = Q_c(1 - \alpha)(1 - 0.62C + 0.0019\theta_N)$$



Factors affecting Q_{SW} :

- α (albedo): depends on sun angle and sea state
- Cloud cover: dominant effect
- Atmospheric absorption: water vapor, aerosols

↓ Maximum in tropics, minimum at poles

Shortwave Radiation (Q_{SW})

What affects the radiation that reaches the surface?

- Q_c : rate of solar energy arriving above the atmosphere: solar constant (1365 to 1372 $W \cdot m^{-2}$).
- Radiation scattered or absorbed and re-irradiated by the atmosphere, clouds, etc. This contribution becomes more important at high latitudes;
- Albedo or reflectivity: surface characteristic. Ratio between reflected and incident radiation (%). Depends on solar elevation and sea state;
- Clouds: scatter, reflect, and absorb radiation;
- Absorption in the atmosphere (without clouds): depends on solar elevation and atmospheric composition (gas molecules, dust, water vapor, etc).

Shortwave Radiation (Q_{SW}): From Satellite to Surface

How is it calculated by global datasets (e.g., ISCCP, ERA5)?

- Not a simple equation. It is derived by combining **satellite measurements** with a **radiative transfer model**
- **Inputs:** Satellite-observed cloud properties
- **Method:** The radiative transfer model simulates how radiation passes through this "virtual atmosphere" to calculate the final surface flux
- **Reanalysis (e.g., ERA5):** Integrates satellite-derived fluxes with a weather model to produce a globally consistent, gap-free dataset

Source: Adapted from Laszlo & Masuda (2006, EGU); Pinker & Laszlo (1992, U. Maryland); ERA5 documentation.

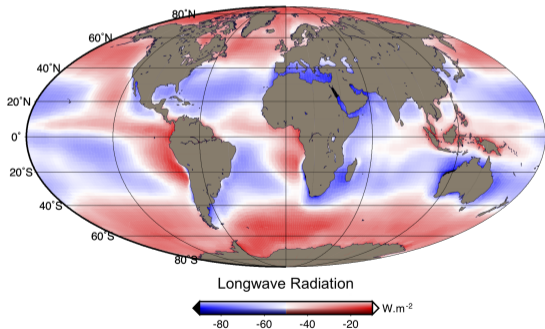
Longwave Radiation (Q_{LW}) - Radiative Cooling

Important processes:

Stefan-Boltzmann law:

$$Q_{LW} = \epsilon\sigma T_w^4(0.39 - 0.05\sqrt{e})(1 - kc^2) + 4\epsilon\sigma T_w^3(T_w - T_a)$$

- Ocean emits as a gray body ($\epsilon \approx 0.98$)
- Atmosphere re-emits back \rightarrow greenhouse effect
- Clouds reduce radiative loss
- Atmospheric humidity is critical



Non-linear effect: Increasing SST can reduce Q_{LW} !


Longwave radiation flux (Q_{LW})

What affects this radiation? (Physical Controls)

- ϵ : emissivity. Stefan's Law and the definition of a black (gray) body.
- e : vapor pressure. Reduces Q_{LW} loss \rightarrow water vapor radiates longwave radiation back to the sea;
- Clouds: reduce Q_{LW} ;
- Temperature difference: atmosphere's response to sea radiation due to humidity. Important in western boundary current regions;

Longwave Radiation (Q_{LW})

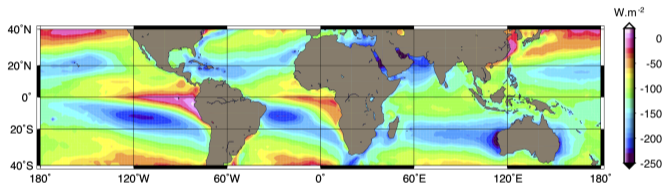
How is it calculated by global datasets (e.g., ISCCP, ERA5)?

- Derived using a **radiative transfer model** (RTM).
- **Inputs:** Satellite-retrieved cloud properties
- **Method:** The RTM simulates how longwave energy from the surface is absorbed and re-emitted by gases and clouds to calculate the net flux at the surface
- **Reanalysis (e.g., ERA5):** Integrates these satellite-derived observations with a forecast model to produce a dynamically consistent, gap-free dataset 

Latent Heat Flux (Q_{LH})

Bulk formula:

$$Q_{LH} = \rho_a C_E L u (q_s - q_a)$$



OAFLUX

Largest heat loss at all latitudes

Maximum in subtropics (trade winds + dry air)

Controlled by three factors:

- **Wind (u):** boundary layer turbulence
- **Humidity ($q_s - q_a$):** saturation deficit
- **Transfer coefficient (C_E):** depends on stability

Evaporative Cooling of the Oceans

What drives latent heat flux?

Latent heat flux represents the energy loss from the ocean due to evaporation.

- $Q_e = \rho_a L_e C_e U (q_s - q_a)$: Sensitive to wind speed and humidity gradient.
- ρ_a : air density; L_e : latent heat of vaporization; C_e : exchange coefficient; U : wind speed; q_s : saturation specific humidity at sea surface (strong function of SST); q_a : specific humidity of air just above surface.

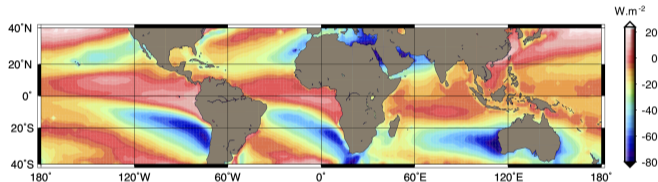
Evaporative Cooling of the Oceans ... continuation

- **Wind speed (U):** Higher winds increase turbulent mixing, removes moist air, enhance evaporation: mid-latitude storm tracks are regions of high Q_{LW} .
- **SST:** Warmer water increases q_s exponentially (Clausius-Clapeyron). The tropics are the primary source of latent heat loss despite low winds.
- **Air-sea humidity difference ($q_s - q_a$):** Dry air advected over warm water (e.g., western boundary currents like the Gulf Stream and Kuroshio) drives intense latent heat loss.

Sensible Heat Flux (Q_{SH})

Bulk formula:

$$Q_{SH} = \rho_a C_p C_H u (T_w - T_a)$$



OAFUX

■ **Positive flux ($T_w < T_a$):** ocean gains heat

■ Equatorial upwelling regions

■ **Negative flux ($T_w > T_a$):**
ocean loses heat

- Western boundary currents (Gulf Stream, Kuroshio)
- Mid-latitudes with strong winds

Sensible Heat Flux (Q_s): Direct Atmospheric Heating

What drives sensible heat flux?

Sensible heat flux represents the direct transfer of heat between the ocean surface and the atmosphere due to temperature differences. Unlike latent heat, no phase change occurs.

- $Q_s = \rho_a c_p C_h U (T_s - T_a)$: Proportional to the temperature gradient and wind speed.

Sensible Heat Flux (Q_s): Direct Atmospheric Heating ... continuation

- ρ_a : air density; c_p : specific heat capacity of air at constant pressure; C_h : exchange coefficient; U : wind speed; T_s : sea surface temperature; T_a : air temperature just above the surface.

Major controlling factors:

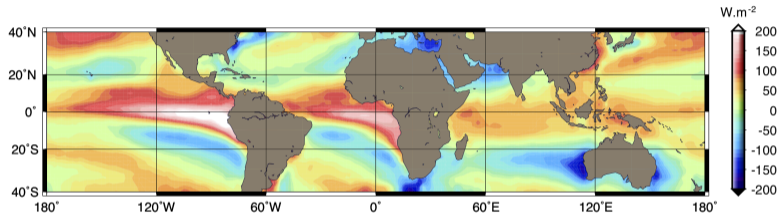
- **Sea-air temperature difference ($T_s - T_a$):** The primary driver. Ocean heats the atmosphere when $T_s > T_a$ (upward flux) and cools when $T_s < T_a$ (downward flux).

Sensible Heat Flux (Q_s): Direct Atmospheric Heating ... continuation

- **Wind speed (U):** Stronger winds enhance turbulent mixing, increasing the efficiency of heat transfer. This is why storms and cold air outbreaks produce large Q_s .
- **Synoptic weather patterns:** Advection of cold continental air over warm ocean waters (e.g., wintertime off the east coasts of continents) creates the largest positive sensible heat fluxes.

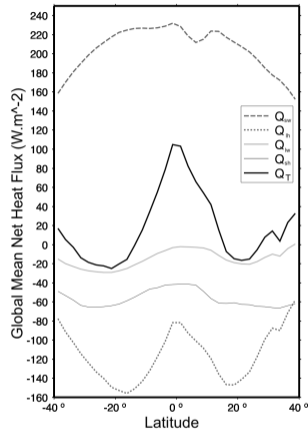
Summing the Components: Net Flux

$$Q_{net} = Q_{SW} + Q_{LW} + Q_{LH} + Q_{SH}$$



Physical Meaning

Over long time scales and globally, $Q_{net} \approx 0$ (steady state), but regional imbalances drive ocean circulation.



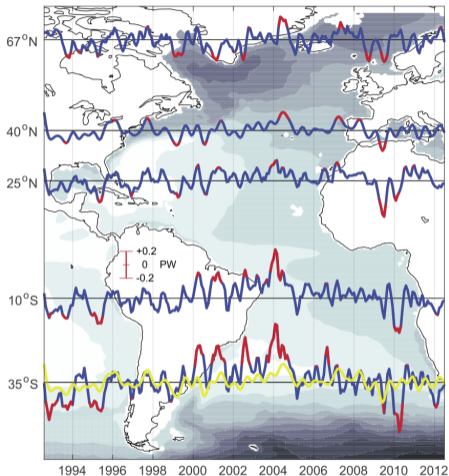
Tropics: $Q_{net} > 0$

Poles: $Q_{net} < 0$.

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Oceanic Meridional Heat Transport



Kelly et al. (2016)

- Heat transports from tropics to poles
- Peak of 2 PW (2×10^{15} W) at 20°N
- Comparable to atmospheric transport at low latitudes
- Atlantic is special: net northward transport across equator

Meridional Heat Transport Across a Latitude

The MHT Formula

$$\text{MHT}(\phi) = \rho_0 c_p \int_x \int_{z=-H}^0 v(x, y, z) \theta(x, y, z) dz dx$$

- ρ_0 : Density of seawater
- c_p : Specific heat capacity
- $v(x, y, z)$: Meridional velocity
- $\theta(x, y, z)$: Potential temperature ($^{\circ}\text{C}$ or K)
- H : Ocean depth (from bottom to surface)
- x : Zonal integration across the entire basin

MHT Decomposition

Decomposition (Bryden and Imawaki 2001)

$$\text{MHT} = \text{MHT}_{\text{mean}} + \text{MHT}_{\text{overturning}} + \text{MHT}_{\text{gyre}}$$

$$\text{MHT}_{\text{overturning}} = \rho_0 c_p \int \langle \mathbf{v} \rangle \langle \theta \rangle dz$$

$$\text{MHT}_{\text{gyre}} = \rho_0 c_p \int_x \int \langle \mathbf{v}' \theta' \rangle dz$$

mean represents the section average,

$\langle \cdot \rangle$ denotes zonal average,

primes denote deviations from the zonal average.

Interpretation:

- **Overtuning component:** Northward flow at warm temperatures, southward return flow at cold temperatures. Captures the thermohaline circulation (e.g., AMOC).
- **Gyre component:** Horizontal recirculations (subtropical and subpolar gyres) that also transport heat because northward and southward flows occur at different temperatures.

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Estimating Heat Storage (HS): Traditional

Why estimate Heat Storage?

Heat storage is fundamental to the ocean heat budget:

$$Q_{net} = \nabla_H F_o + \frac{\partial HS}{\partial t}$$

where Q_{net} is the net air-sea heat flux, $\nabla_H F_o$ is the divergence of advective ocean heat transport, and $\partial HS / \partial t$ is the rate of change of heat storage.

Traditional Method (In situ)

Based on vertical integration of temperature profiles:

$$HS = \rho C_p \int_{-h}^0 T(z) dz$$

- $T(z)$: vertical temperature profile
- h : integration depth

Limitation: Sparse data (hydrographic cruises, buoys, XBTs). Limited global coverage.

HS Estimation via Altimetry

Physical principle

Sea surface height anomalies (η) are correlated with changes in ocean heat content because:

- Thermal expansion/contraction of the water column (steric effect) dominates sea level variability on seasonal to interannual timescales.
- Surface displacement is proportional to isopycnal displacement.

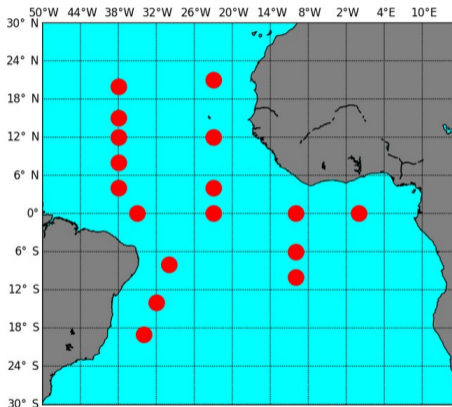
HS from Altimeter Equation (Chambers et al. 1997)

$$HS' = \frac{\rho C_p}{\alpha} \eta$$

- HS' : heat storage anomaly (J m^{-2})
- η : sea surface height anomaly (altimeter)
- α : thermal expansion coefficient (averaged over the integration layer)

Advantage: Global coverage, high spatial and temporal resolution.

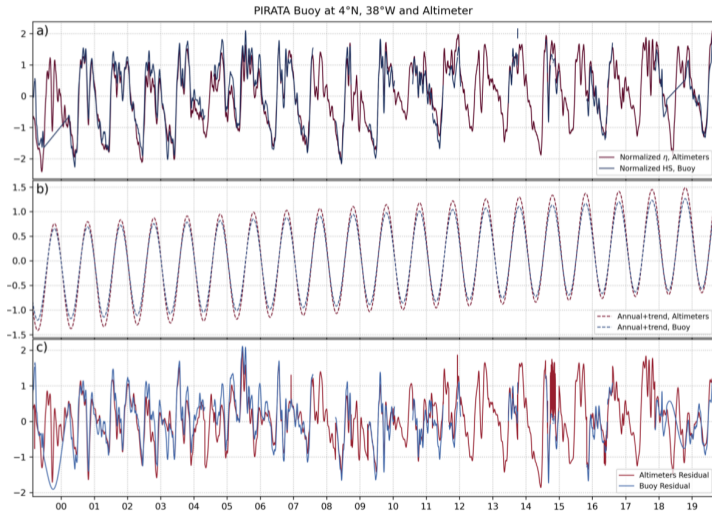
HS calculations at the Tropical Atlantic



The PIRATA Array

- Upper ocean monitoring system since 1997
- Air-sea interaction processes study
- Understand the weather variability of NE Brazil
- Temperature measurements up to 500 m, salinity up to 120.

Altimeter \times Buoy at 4°N 38°W (Best case)



■ Blue = Buoy, Red = Sat.

top : normalized, 17-point
Blackman low-passed

mid : Least Sq. Fit of
(annual + trend)

bot : Residual

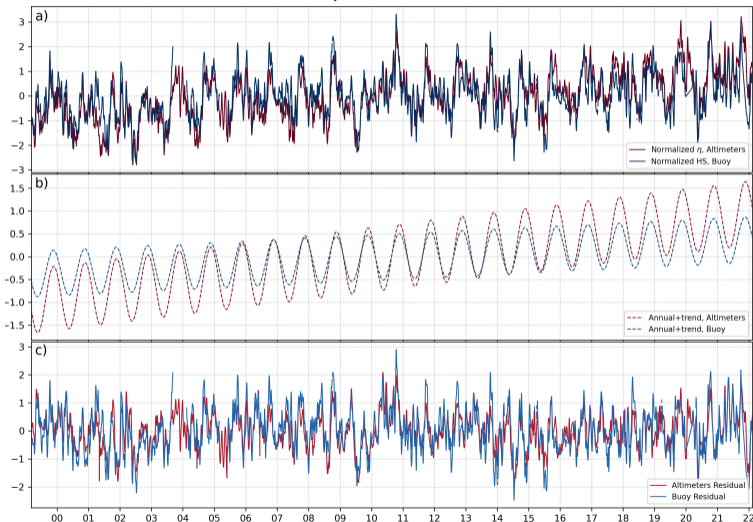
■ Corr.: Total = 88%,

Annual = 100%,

Residual = 75%.

Altimeter \times Buoy at 0°N 23°W (Few Gaps)

PIRATA Buoy at 0°, 23°W and Altimeter



■ Blue = Buoy, Red = Sat.

■ Corr.: Total = 80%,
Annual = 96%,
Residual = 81%.

■ η trend $\sim 2 \times HS'$
trend.

■ Noticeably different
spectrum from 4°N,
38°W.

Validation: Altimeter vs. In situ Data (PIRATA)

Comparison in the Tropical Atlantic (Sato & Polito, 2008)

Location	Correlation (c)	rms ($\times 10^9 \text{ J m}^{-2}$)
Equator (0°, 10°W to 35°W)	0.73 – 0.87	0.40 – 0.51
2°N, 38°W	0.80	0.52
6°S, 10°W	0.67	0.62
15°N, 38°W (Amazon region)	0.34 – 0.50	0.68

Validation: Altimeter vs. In situ Data (PIRATA) ... continuation

Important Limitation

Correlation is lower in regions where salinity plays a dominant role (e.g., 15°N, 38°W — Amazon River and ITCZ influence). Even with haline correction, local dynamics weaken the relationship between η and HS .

Conclusion: Altimetry is a reliable tool for estimating HS , especially where large-scale processes dominate (tropics, subtropical gyres). Haline correction improves results, but in situ salinity data are superior to climatology.

Haline Correction (Salinity Effect)

Sea level also responds to salinity changes (halosteric effect):

$$\eta_h = \int_{-h}^0 (-\beta) \Delta S dz$$

where β is the haline contraction coefficient and ΔS is the salinity anomaly.

Without haline correction:

- Assumes η is purely thermosteric.
- Error in regions influenced by river discharge (e.g., Amazon), intense precipitation (ITCZ), or deep convection (Labrador Sea).

With haline correction:

- From a reanalysis or climatology.
- **Result:** Correlation between altimeter and in situ HS increases by up to 22%; rms decreases by up to 12% (Sato et al., 2000).

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Key Takeaways

- 1 **Air-sea heat fluxes** are the ocean's energy lifeline
- 2 **Latent heat** is the dominant term, controlled by winds and humidity
- 3 **Net heat gain in tropics, loss at poles** → drives meridional heat transport
- 4 **Long-term trends** show a changing budget, with ocean circulation playing a key redistribution role
- 5 **Satellites + in situ** = powerful monitoring system

Outlook & Open Questions

- **How will the balance shift in a warming world?**
- **Can we close the budget at finer scales (mesoscale eddies)?**
- **New missions:** SWOT (surface water and ocean topography), future altimetry
- **Salinity from space:** Linking water cycle to heat fluxes

The future of air-sea interaction science is bright!

Acknowledgments

Thank you!

Questions?