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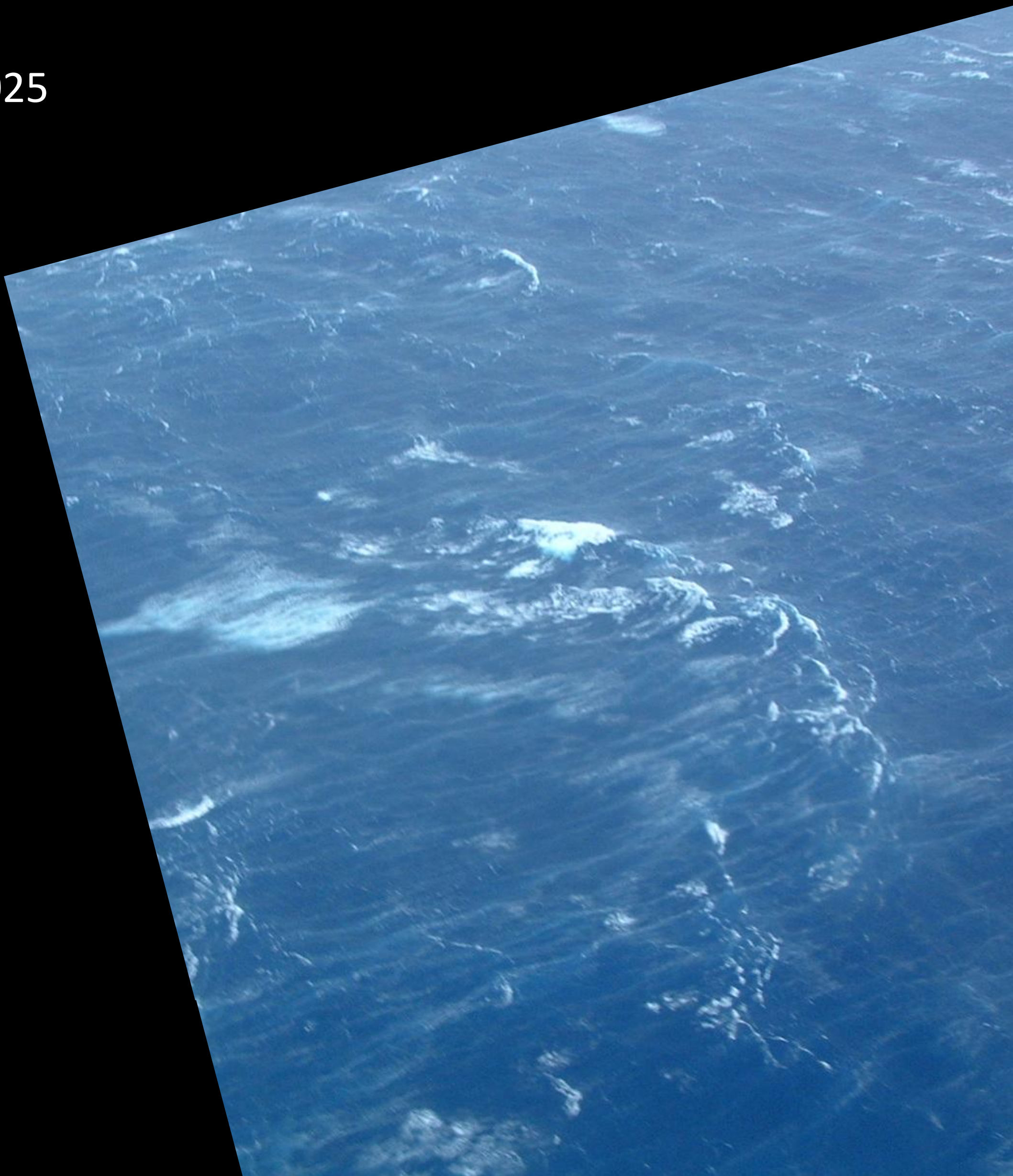
Alex Babanin, *a.babanin@unimelb.edu.au*

Wave Hindcast and Forecast Meeting
Santander, Spain
22 September 2025

Ocean Waves in the Large-Scale Air-Sea System



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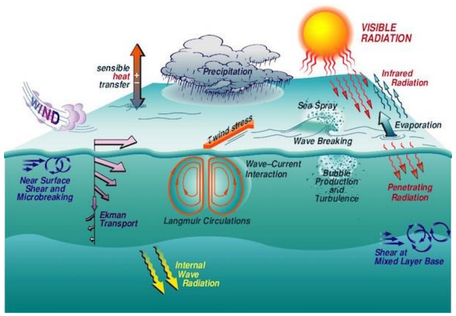
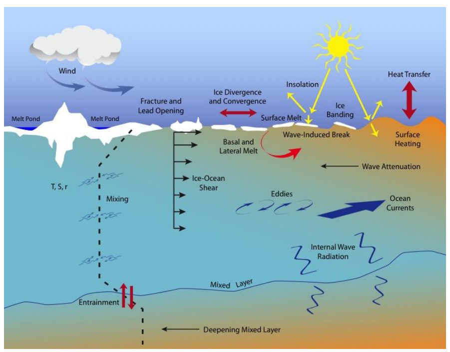
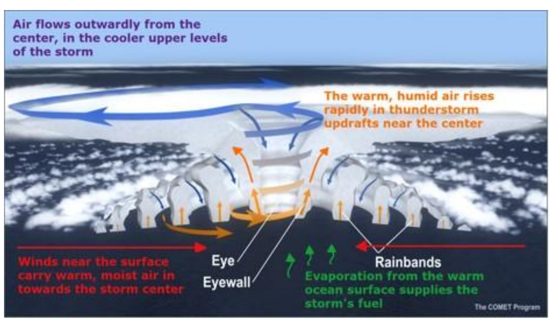
Small/large-scale air-sea processes are essentially coupled in nature, but not in the models







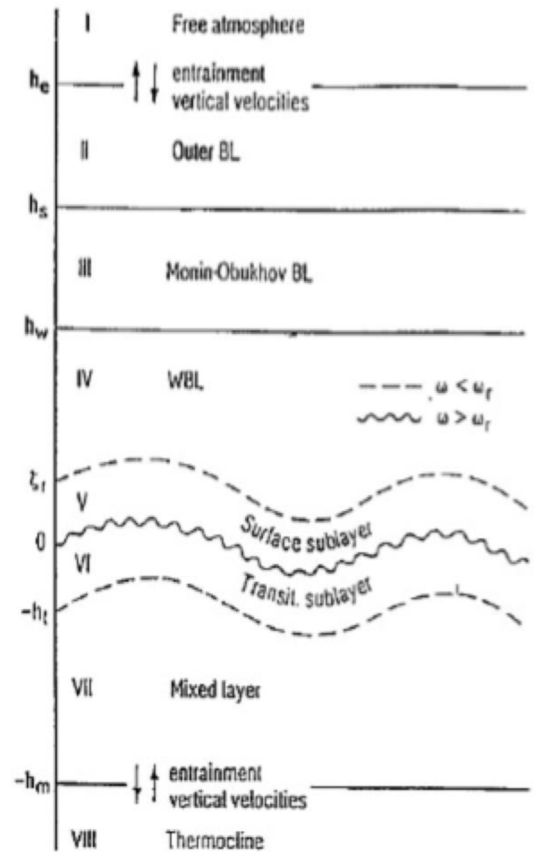
> Atmospheric boundary layer
  - winds generate waves
  - waves provide surface roughness and change the winds
  - waves evolve, fluxes change
  - waves generate spray

> Upper ocean mixed layer
  - waves generate currents
  - produce turbulence
  - turbulence: facilitates mixing
  - changes the circulation, heat content, SST, nutrient transport
  - facilitate gas exchange

Tradition and future

> Small scales and large scales are separated. Models reach saturation in their performance

> They need to be coupled, from turbulence to climate. Understanding exists, computer capacity exists



Chalikov & Belevich, 1993, BLM

wave mixing in the upper ocean

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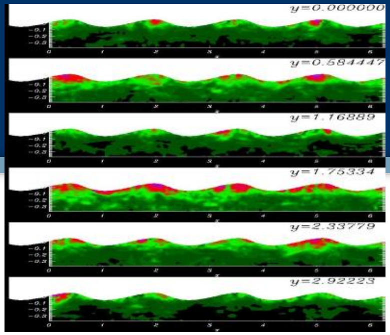
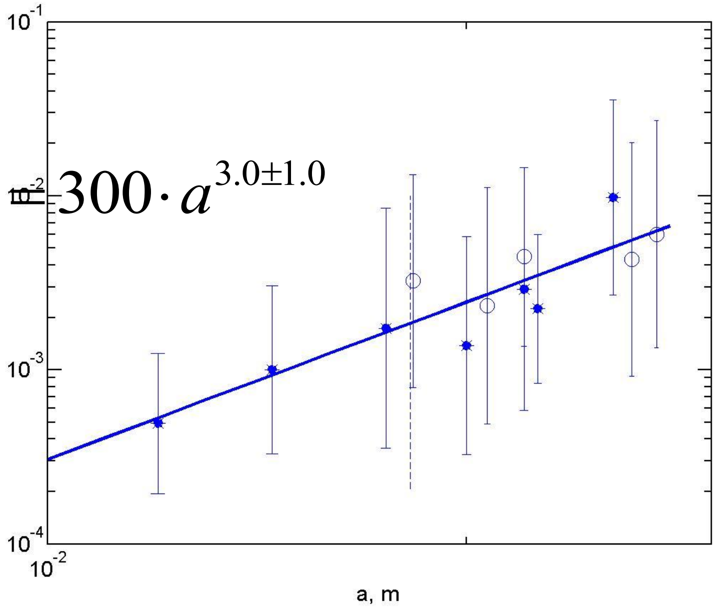


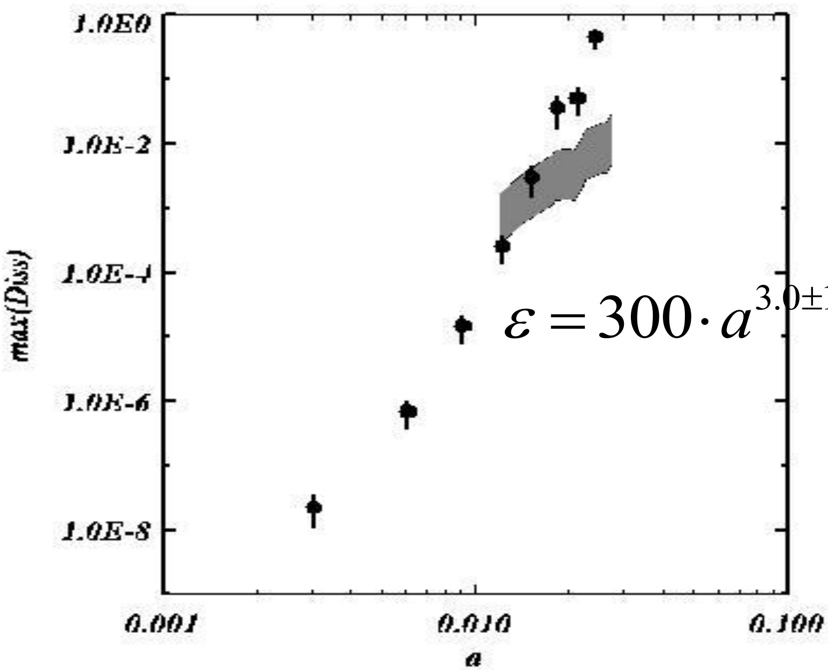
Model of generation of turbulence by nonlinear waves

Model is based on exact 2-D (x-z) model of surface waves coupled with 3-D LES (x-y-z) model of vortical motion based on Reynolds equation with parameterised subgrid turbulence



$$\varepsilon = 300 \cdot a^{3.0 \pm 1.0}$$

Lab experiment,



$$\varepsilon = 300 \cdot a^{3.0 \pm 1.0}$$

Babanin & Haus, JPO, 2009

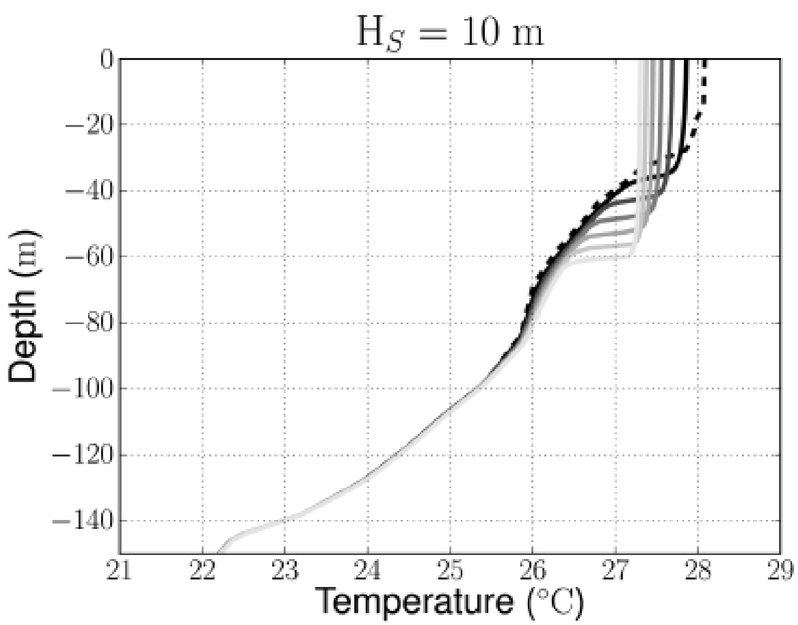
Babanin & Chalikov, JGR, 2012

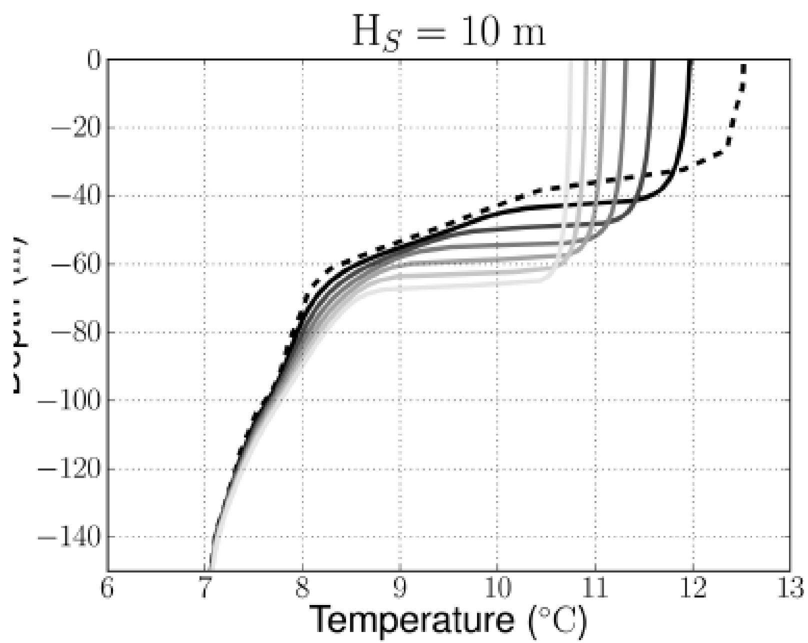
Modelling SST and MLD at the scale of tropical cyclone

Ghantous and Babanin, Nonlin. Proc. in Geophysics, 2014

$$\epsilon_{dis} = b_1 k \omega^3 a_0^3$$

$$P_{\mathrm{w}} = b_1 \kappa\, \omega_p^3 \frac{H^3}{8} e^{3\kappa z},$$





$$\frac{\partial K}{\partial T_{mean}} + U_i \frac{\partial K}{\partial X_i} = D_K + P_S + G - E_K$$

$$b_1 = 5\left(\kappa \frac{H}{2}\right)^2.$$

Earth and Space Science

RESEARCH ARTICLE

Wave-Coupled Effects on Oceanic Biogeochemistry: Insights
iogeochemical Model in the Southern

ander V. Babanin[1], and Mihir Kumar Dash[2]

ıe University of Melbourne, Parkville, VIC, Australia, [2]Centre for Ocean,
Institute of Technology Kharagpur, Kharagpur, India

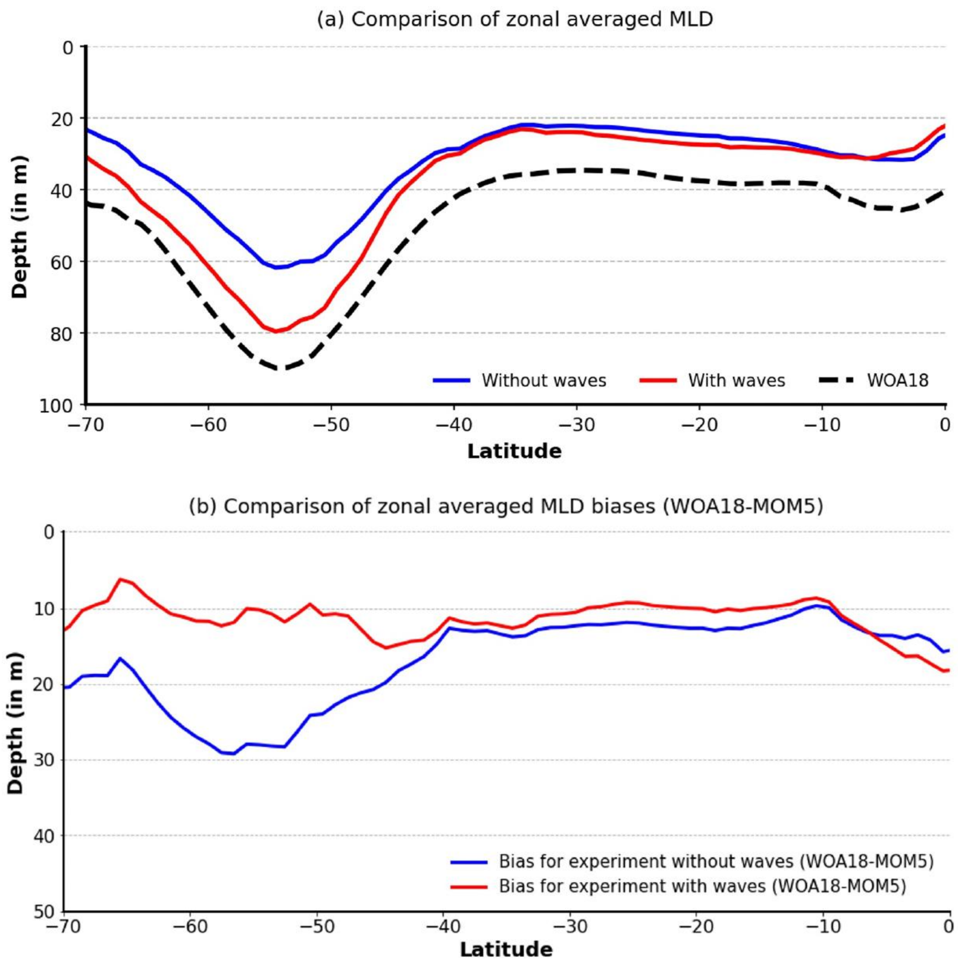


Figure 5. (a) Comparison of zonal averaged mixed layer depths (MLDs) during January to March season and (b) Comparison of zonal averaged MLD biases for "with waves" and "without waves" experiments for the same season.

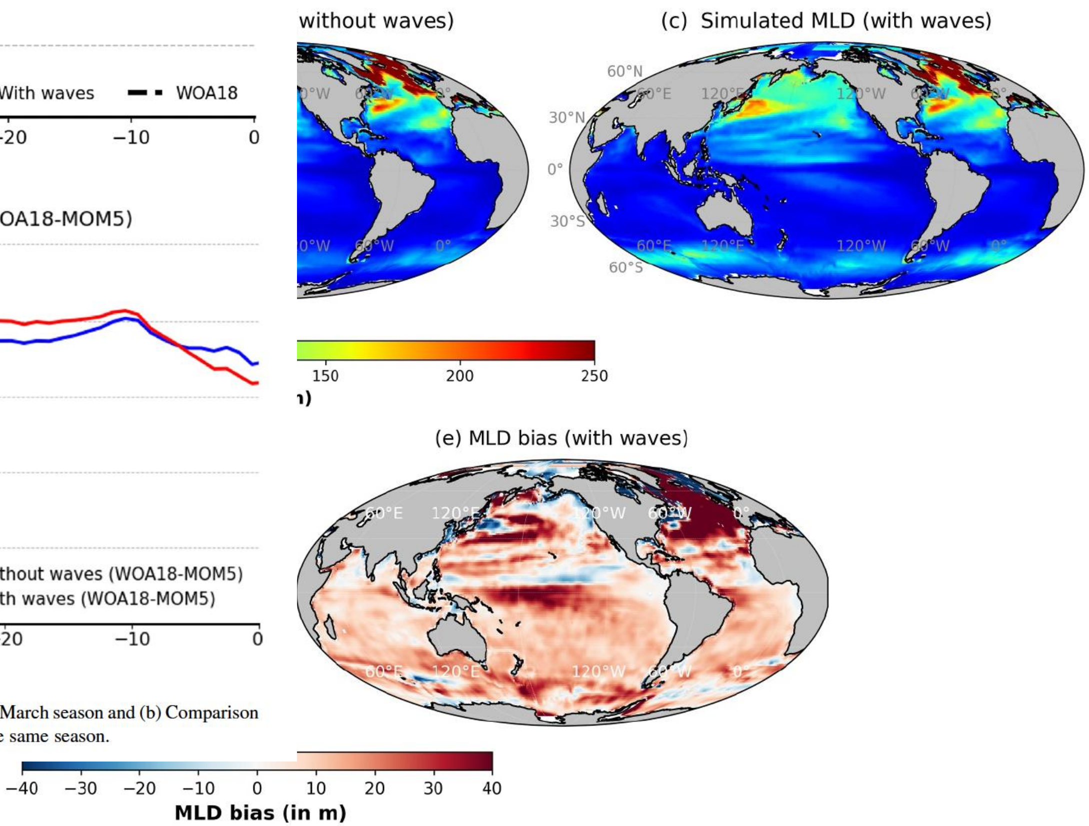


Figure 4. A comparison of simulated winter (January–March) mixed layer depth (MLD) with observation. (a) Observed MLD (WOA18), (b) simulated MLD (without waves), (c) simulated MLD (with waves), (d) MLD bias (without waves), and (e) MLD bias (with waves).

Journal of Advances in Modeling Earth Systems

RESEARCH ARTICLE
10.1002/2016MS000878

Simulated ocean response to tropical cyclones: The effect of a novel parameterization of mixing from unbroken surface waves

Lachlan Stoney[1], Kevin Walsh[2], Alexander V. Babanin[1], Malek Ghantous[3], Pallavi Govekar[4], and Ian Young[1]

Key Points:

- Mixing from unbroken surface waves under tropical cyclones can modify ocean surface temperatures by up to 0.5°C

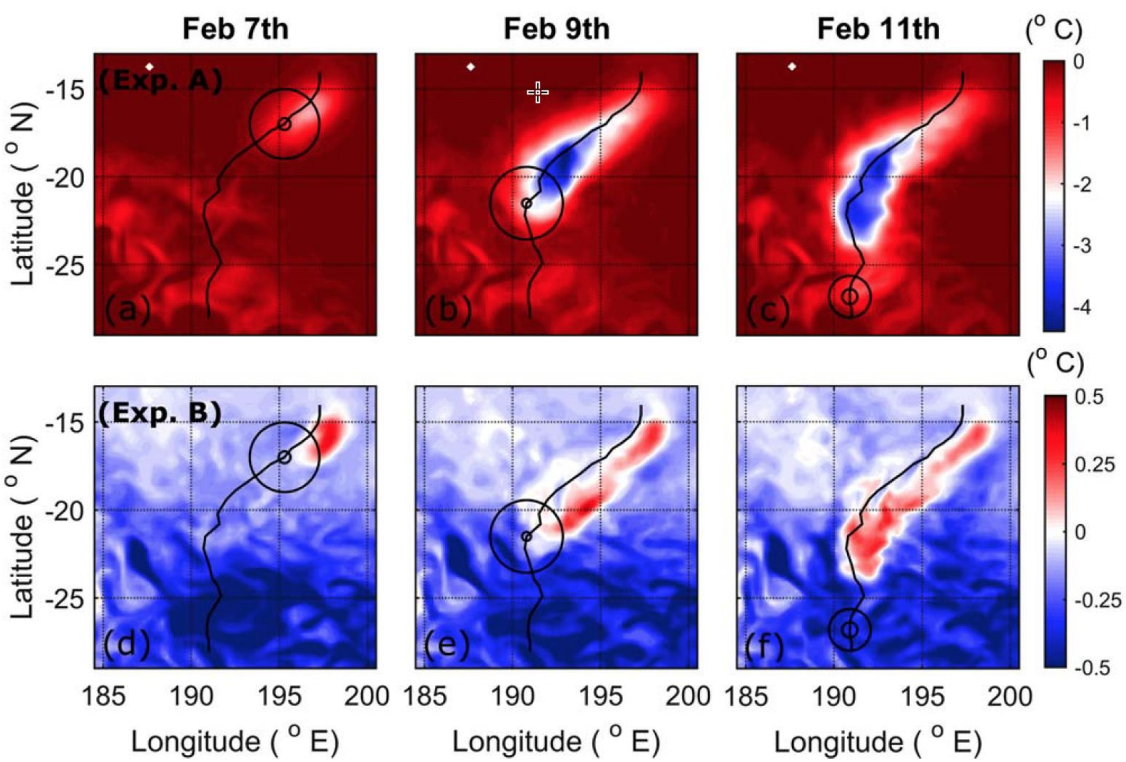


Figure 3. The development of the cold wake from Tropical Cyclone Dovi, 2003. The top plots (a–c) show the potential temperatures at 5 m depth from Experiment A, while the bottom plots (d–f) show the wave-induced anomalies thereof from Experiment B. Values are daily averages. The storm moved along the black line from top right to bottom left, with the inner and outer circles representing the *RMW* and R34, respectively, for each day at 12:00 UTC. Note the different color scales.

- Southern Hemisphere TC
- MOM5
- (top) no waves
- (bottom) warm anomalies on the side with the strongest winds and cool anomalies in other regions
- initial wave-induced deepening of the mixed layer, which can modify the subsequent shear-induced entrainment and upwelling
- could potentially influence tropical cyclone intensity and structure

Journal of Advances in Modeling Earth Systems

RESEARCH ARTICLE
10.1002/2016MS000707

The effect on simulated ocean climate of a parameterization of unbroken wave-induced mixing incorporated into the k-epsilon mixing scheme

Kevin Walsh[1], Pallavi Govekar[1,2], Alexander V. Babanin[3], Malek Ghantous[4], Paul Spence[5,6], and Enrico Scoccimarro[7,8]

Key Points:
- The inclusion of a process that is missing from most climate models causes noticeable changes in Southern Ocean regional ocean

Table 1. Biases Versus the World Ocean Atlas (2009) Data (http://data.nodc.noaa.gov/woa/WOA09/) for the Two Simulated Years, Without and With Wave-Mixing Included, at Specified Ocean Depths for Selected Latitude Bands[a]

	1980				2007			
	December Without	December With	July Without	July With	December Without	December With	July Without	July With
5 m Temperature								
Global	0.11	0.03	0.17	0.05	0.31	0.22	0.18	0.07
70°S–50°S	−0.11	−0.25	−0.18	−0.06	0.26	0.05	−0.43	−0.25
50°S–30°S	0.61	0.20	0.08	0.14	0.38	0.00	−0.21	−0.16
30°S–30°N	0.18	0.17	0.29	0.22	0.55	0.54	0.52	0.45
30°N–50°N	−0.15	0.18	1.02	−0.15	−0.46	−0.17	1.45	0.29
50°N–70°N	−0.56	−0.22	0.99	−0.19	−0.74	−0.35	0.78	−0.37
25 m Temperature								
Global	0.06	0.08	−0.05	0.09	0.27	0.29	−0.01	−0.11
70°S–50°S	−0.15	−0.18	−0.16	−0.04	−0.03	0.01	−0.41	−0.48
50°S–30°S	0.27	0.15	0.14	0.20	0.22	0.03	−0.15	−0.42
30°S–30°N	0.20	0.24	0.13	0.21	0.65	0.68	0.46	0.27
30°N–50°N	−0.08	0.25	−0.65	−0.20	−0.40	−0.11	−1.23	−0.30
50°N–70°N	−0.49	−0.16	−0.73	0.01	−0.67	−0.29	−0.71	−0.33
55 m Temperature								
Global	−0.02	0.17	−0.02	0.12	0.18	0.39	0.07	0.19
70°S–50°S	−0.22	−0.04	−0.14	−0.04	−0.39	−0.09	−0.29	−0.21
50°S–30°S	−0.43	−0.06	0.20	0.26	−0.31	0.01	−0.10	−0.05
30°S–30°N	0.31	0.42	0.11	0.23	0.77	0.89	0.51	0.60
30°N–50°N	−0.02	0.39	−0.48	0.15	−0.26	0.07	−0.92	−0.39
50°N–70°N	−0.27	0.07	−0.71	−0.17	−0.45	−0.07	−0.81	−0.14

[a]Regions where the introduction of wave-induced mixing reduces the bias are shaded.

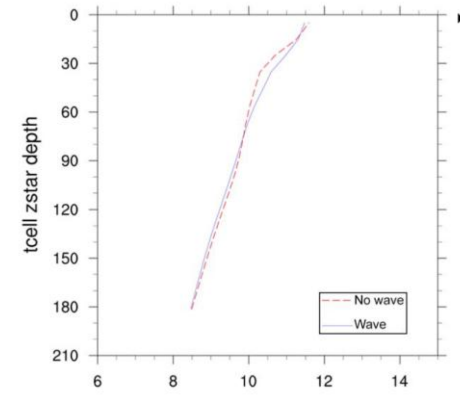
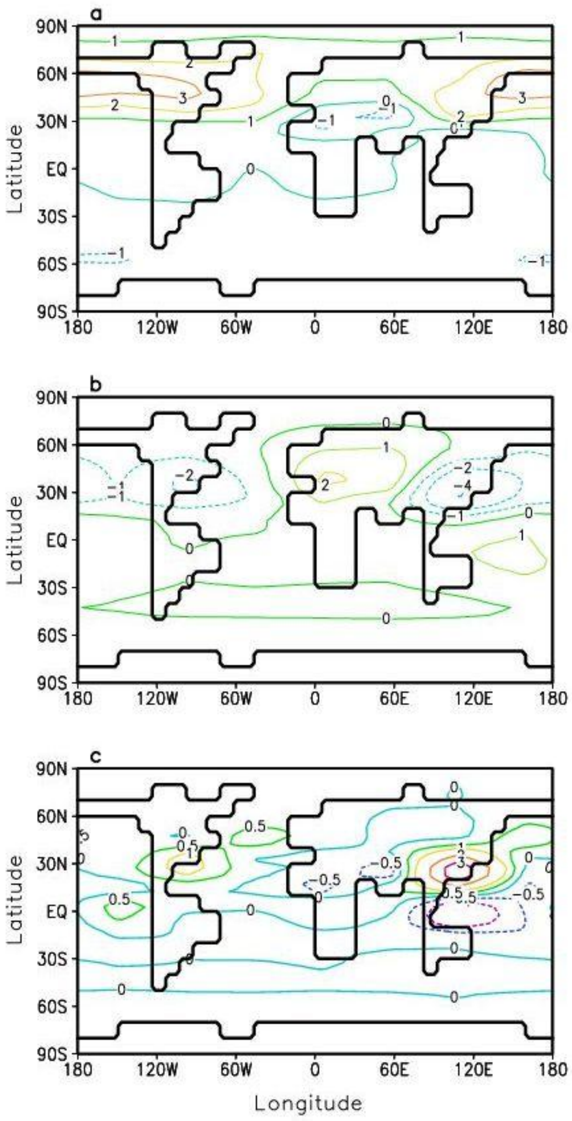


Figure 14. Comparison of global average ocean temperature profiles (°C) for July 1980, wave-mixing included (solid line) versus no wave (dashed line). Vertical axis is in meters and refers to the temperature levels in the model.

MOM5 global modelling

Implementing wave-induced mixing in CLIMBER



Global distribution (Northern summer)

- temperature (*degrees*)

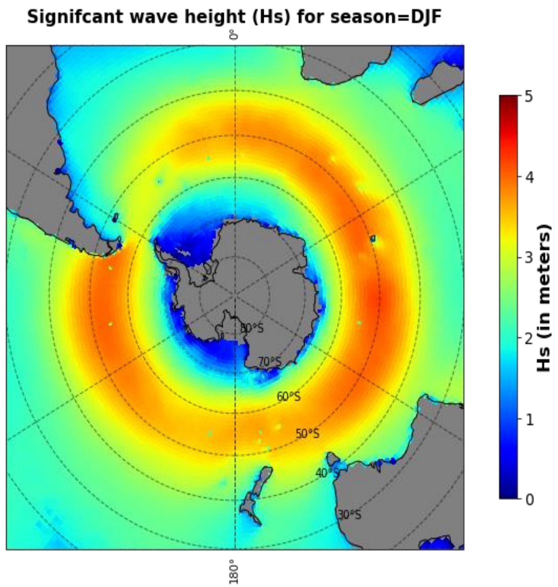
- pressure (*mbar*)

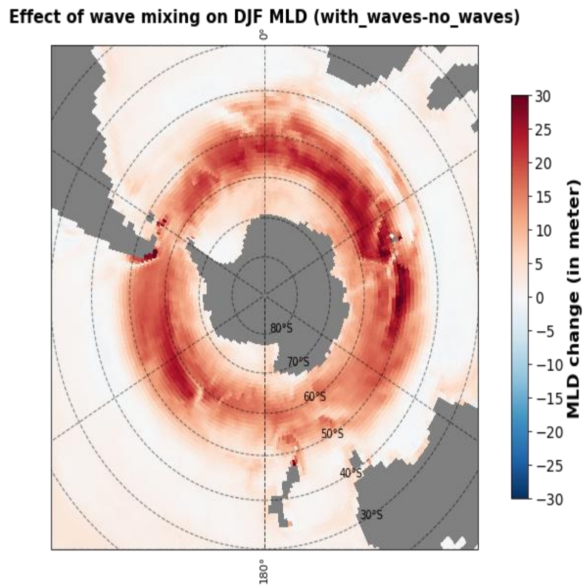
- precipitation (*mm* per day)

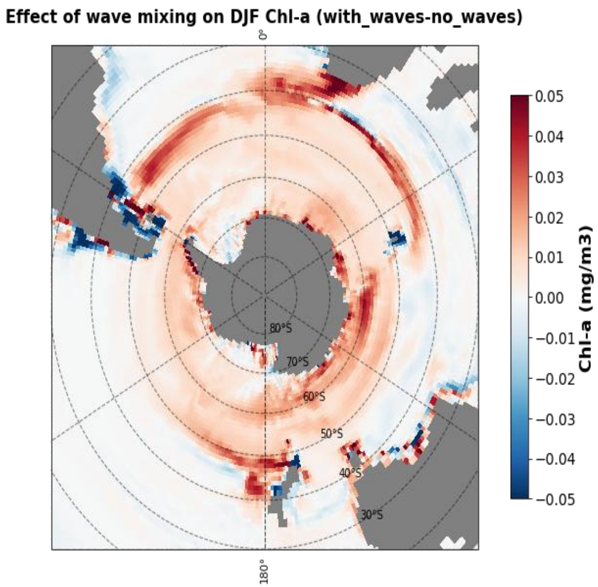
Babanin, Ganopolski, Phillips, 2009, OM

Bigeochemistry

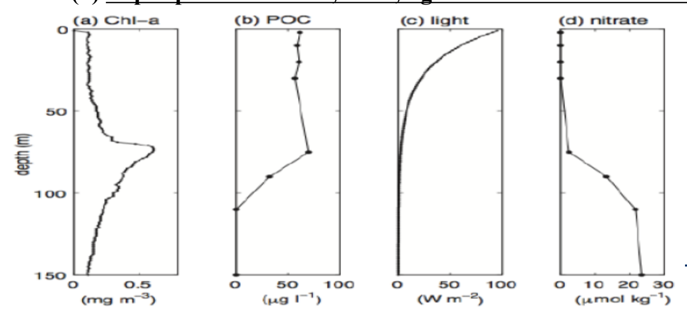
Effect of surface wave mixing on MLD and Chl-a concentration during DJF season in the Southern Ocean







(b) Depth profile of Chl-a, POC, light and nitrate in the ocean



Tensubam, Babanin, Dash, 2024, *STE*

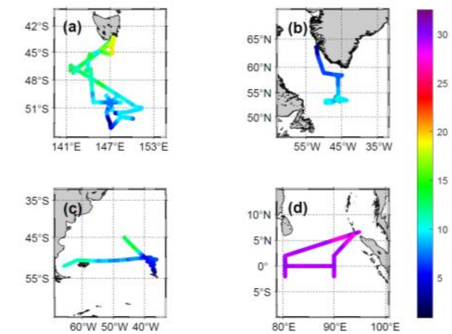
CO_2 transfer due to waves

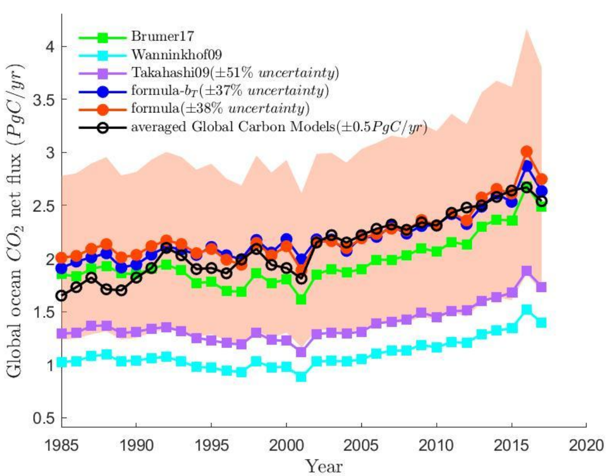
$$\widetilde{K} = \frac{K_{660}}{U_{wm}},\ R_{HM} = \frac{H_s \cdot U_{wm}}{\nu},$$
$$\widetilde{U} = \frac{U_*}{\sqrt{g \cdot H_s}},\ \widetilde{V_b} = \frac{V_b}{U_{wm}}$$

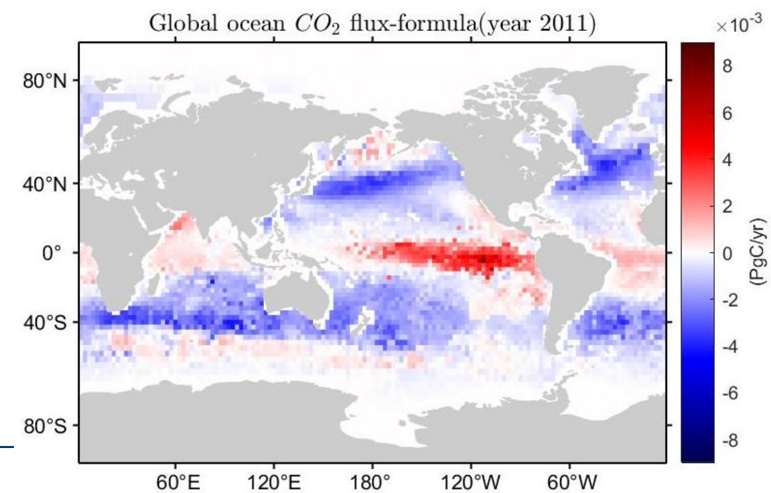
Suggested Formula

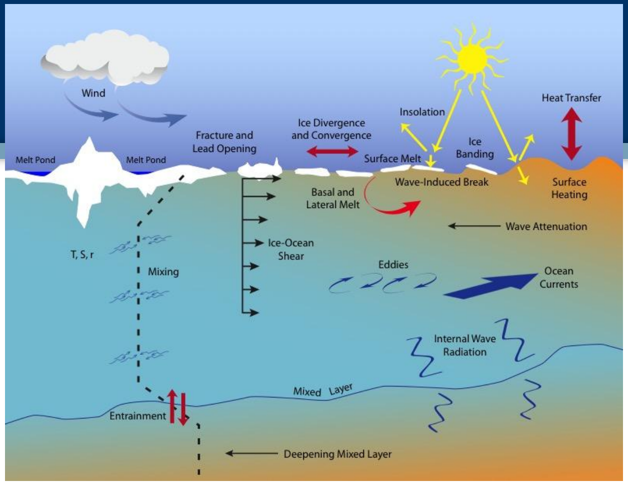
A combined formula is proposed for breaking/non-breaking wave conditions which can be determined by spectral wave steepness ε (Babanin *et al* (2001)).

$$\widetilde{K} = \begin{cases} 9.57 \cdot 10^{-11} \cdot \left[R_{HM} \cdot \left(1 + \widetilde{U}\right)\right]^{0.876}, & \varepsilon \leq 0.055 \\ 2.82 \cdot 10^{-11} \cdot \left[b_T \cdot R_{HM}^{4} \cdot \left(1 + \widetilde{U}\right)\right]^{0.260}, & \varepsilon \geq 0.055 \end{cases}$$









Wave-coupled effects in Marginal Ice Zone

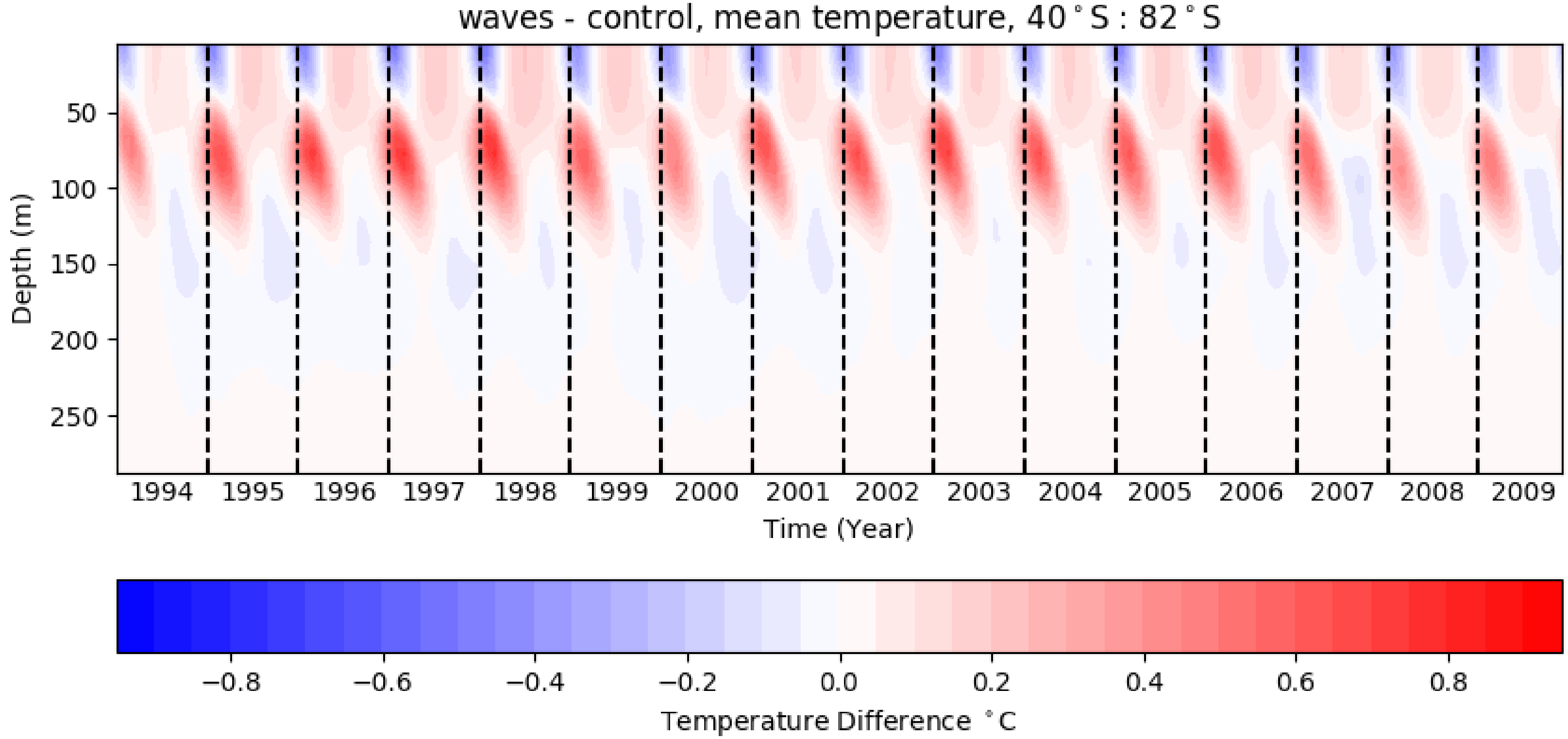
Mixing: wave-induced mixing $P = K_M M^2 + P_W$

Breakup: wave-induced sea ice breakup

When sea ice is broken, we reduce albedo by 40% in summer and increase the rate of frazil ice growth by 10% in winter.

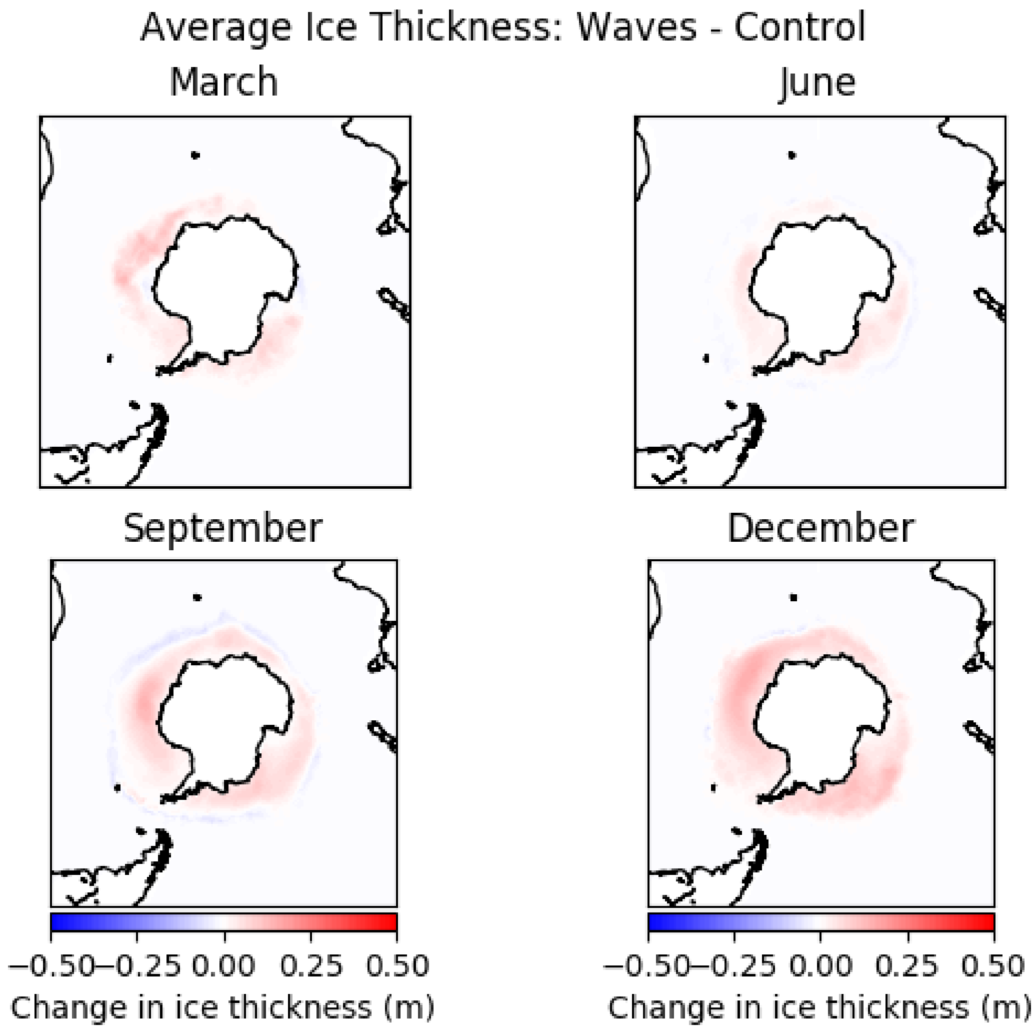
Results from MOM5 Ocean model with and without the inclusion of an extra wave mixing term (from WW3)

Temperature difference between ocean models with and without the extra wave-mixing term (wave and control, respectively).



In the Southern Ocean an increase in wave mixing captures more heat and transports it into deeper waters over the (southern) summer. While producing a net increase in ocean temperature this results in the waters surface being colder during the summer.

Thomas, Babanin, Walsh, Stoney, Heil, 2019, OD



Difference in ice thickness between ocean models with and without the extra wave-mixing term (wave and control, respectively).

The decrease in surface temperature despite the net increase in ocean heat content causes a reduction in the amount of ice melt during the summer. This results in a thicker Antarctic sea sheet, particularly in December (summer).

Conclusions

- wave-induced turbulence (not to be confused with wave-breaking turbulence) is produced at the vertical scale of wavelength
- important for ocean mixing, sediment suspension, tropical cyclones, weather, climate, polar oceans and ice, gas transfer, biogeochemistry
- laboratory experiments, numerical simulations, field observations – all give similar rates for wave turbulence production
- wave models need to meet requirements of correct fluxes in order to be useful for coupling with atmospheric and ocean models

The 7th workshop on waves and wave-coupled processes

24-27 March 2026, Bangkok, Thailand

Register your interest prior to 30 September 2025
https://infrastructure.eng.unimelb.edu.au/ocean/news/air-sea-interface-2026

Over the years, it has become clear that ocean surface waves play a critical role in the Earth System, modulating many surface exchanges, from tropical cyclones to marginal ice zone, as well as acting in the atmospheric boundary layer and the upper ocean. Accounting for their impacts in ocean circulation, extreme marine weather, climate and other large-scale systems has recently attracted renewed interest and requires much attention.

After the previous six successful workshops in Melbourne, Qingdao, Hangzhou, Uppsala, Reading, Melbourne, Kasetsart University will organise the 7th workshop on waves and wave-coupled processes in Bangkok, aiming to foster discussion and collaboration within this field among the wider community. This meeting will be conducted in plenary, with time reserved for discussion to identify key research and technological questions relevant for the uptake of relevant wave information in Earth System models.

Organising Committee:
Dr. Montri Maleewong
Kasetsart University, Thailand
Professor Alexander Babanin
The University of Melbourne, Australia
Professor Fangli Qiao
First Institute of Oceanography, China
Professor Lichuan Wu
Uppsala University, Sweden
Dr. Jean Bidlot
ECMWF, UK

The workshop will cover the following research themes:

- **Dynamics of ocean waves and wave breaking, wave-current interactions**
- **Spectral wave modelling**
- **Air-sea fluxes and atmospheric wave boundary layer**
- **Wave influences in the upper ocean, wave turbulence and mixing**
- **Wave-ice interactions**
- **Wave-coupled processes in extreme metocean conditions, tropical cyclones**
- **Wave-coupled effects in gas transfer, ocean biogeochemistry, ambient noise, other air-sea interface and upper ocean processes**
- **Waves in the large-scale air-sea system, metocean climatology**

Keynote speakers:
Al Osborne, USA
Norden Huang, China
Alexei Slunyaev, Russia
Joey Voermans, Australia
Zhenya Song, China

Regardless of the turbulence source, 3D turbulence is unstable to 2D wave orbital motion

JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 117, C00J30, doi:10.1029/2012JC007948, 2012

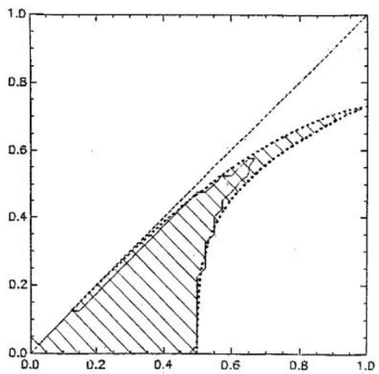


Figure 1. Diagram of stability of solutions of equation set (22) on the $(\varepsilon_1, \varepsilon_2)$ plane. Here ε_1 is the horizontal axis and ε_2 is the vertical axis. F1 corresponds to $\varepsilon_2 = \varepsilon_1$; $F2_1$ corresponds to $\varepsilon_2 = \varepsilon_1 - 0.3\,\varepsilon_1^3 + 0.03\,\varepsilon_1^4$, $0 \le \varepsilon_1 \le 1$; $F2_2$ corresponds to $\varepsilon_2 = 0.9(\varepsilon_1 - 1/2)^{0.3}$, $1/2 \le \varepsilon_1 \le 1$; and F3, the dashed curve, corresponds to the numerical calculations using the Floquet theory [*Hale*, 1969].

On the turbulence generated by the potential surface waves

A. Y. Benilov[1]

Received 31 January 2012; revised 8 May 2012; accepted 12 June 2012; published 18 August 2012.

[1] The turbulence (the random vortex motions) of the upper ocean is nourished by the energy and momentum of the surface waves (the potential motion). The statistical characteristics of the turbulence (turbulent kinetic energy, dissipation rate, and Reynolds stresses) depend on the state of the ocean surface waves. This paper discusses the possibilities of generating this turbulence using the vortex instability of the potential surface waves. The vortex component of fluctuations of velocity field and possibly the interaction between both the vortex and potential motions cause the vertical transport of the momentum. The Reynolds tensor is a linear function of the correlation tensor of vortex field. The initial small vortex perturbations always exist in the upper ocean because of the molecular viscosity influences, especially near the free surface, and the fluctuations of the seawater density. The horizontal inhomogeneities of the seawater density produce the vortex field even if the initial vorticity was zero and the initial flow was the potential flow. The evolution of the small initial vortex disturbances in the velocity field of potential linear surface waves is reduced to a coupled set of linear ordinary differential equations of the first order with periodic coefficients. The solution of this problem shows that the small initial vortex perturbations of potential linear surface waves always grow. The initial small vortex perturbations interacting with the potential surface wave produce the small-scale turbulence (Novikov's turbulence) that finally causes the viscous dissipation of the potential surface wave. The wave-induced turbulence can be considered as developed turbulence with a well distinguishable range of the turbulent wave numbers k where turbulence obeys the Kolmogorov's self-similarity law.

Citation: Benilov, A. Y. (2012), On the turbulence generated by the potential surface waves, *J. Geophys. Res.*, *117*, C00J30, doi:10.1029/2012JC007948.

Benilov, JGR, 2012

Swell attenuation

Dissipation

$$\varepsilon = 300 \square a^{3.0\pm1.0}, \quad b = b_1 k \omega^3 = 30. \quad b_1 = 0.004$$

- volumetric

$$\epsilon_{dis} = b_1 k \omega^3 a_0^3 = 0.004 k u_{orb}^3.$$

- per unit of surface

$$D_a = b_1 k \int_0^\infty u(z)^3 dz = b_1 k u_0 \int_0^\infty \exp(-3kz) dz = \frac{b_1}{3} u_0^3.$$

- per unit of propagation distance

$$D_x = \frac{1}{c_g} D_a = \frac{b_1}{3} 2 \frac{k}{\omega} u_0^3 = \frac{2}{3} b_1 k \omega^2 a_0^3 = \frac{2}{3} b_1 g k^2 a_0^3.$$

$$\frac{g}{2} \frac{\partial (a_0(x)^2)}{\partial x} = \frac{2}{3} b_1 g k^2 a_0(x)^3,$$

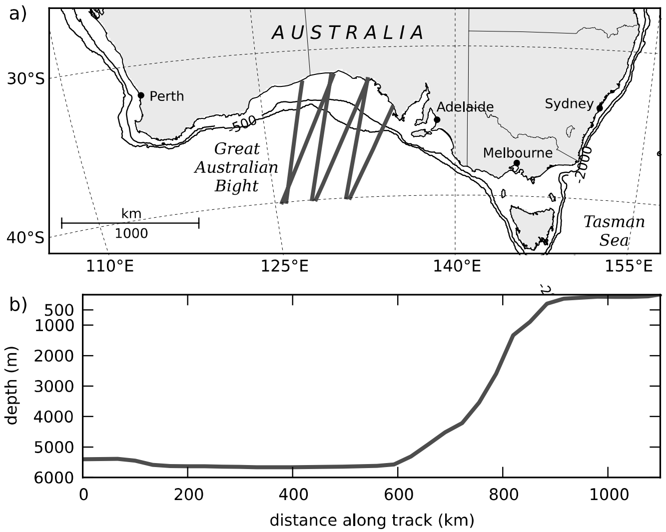
$$a_0(x)^2 = \frac{4}{B^2} x^{-2} = \frac{9}{4 \cdot b_1^2 k^4} x^{-2} = \frac{9}{64} 10^6 k^{-4} x^{-2}.$$

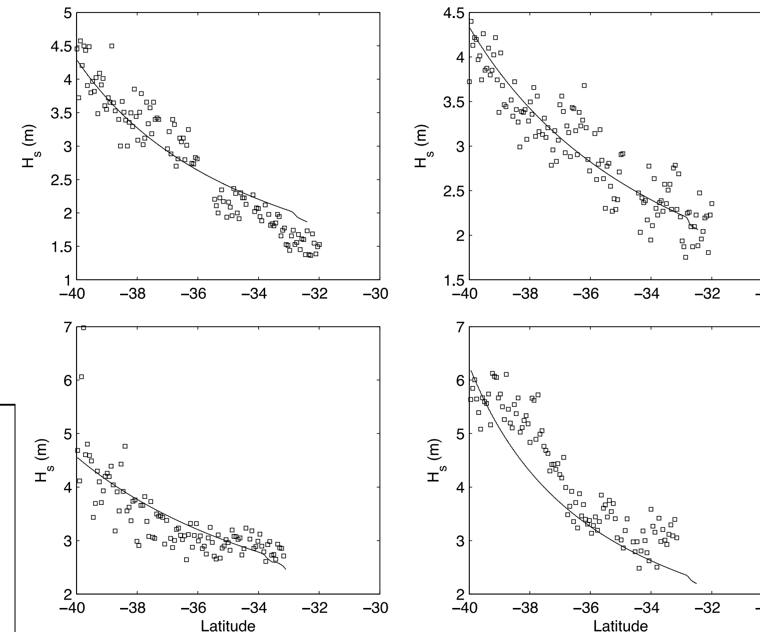
Wave turbulence production profile

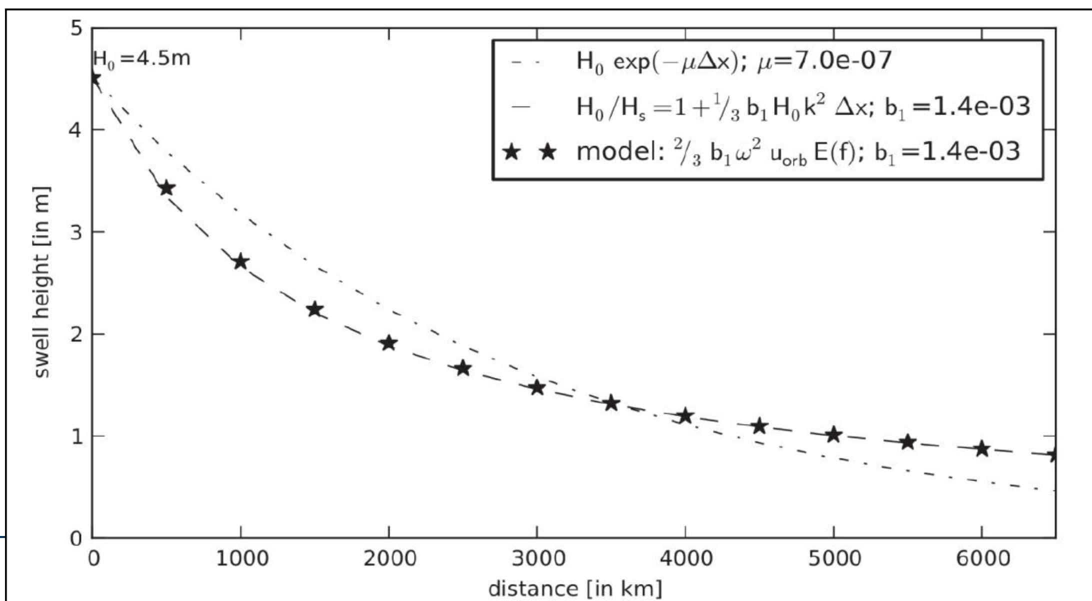
$$P(z) = \varepsilon(z) = b_1 k \omega^3 a_0^3 \exp(-kz)$$

Babanin, 2011, CUP

Swell attenuation







Young, Babanin, Zieger, JPO, 2013

Waves and ocean turbulence

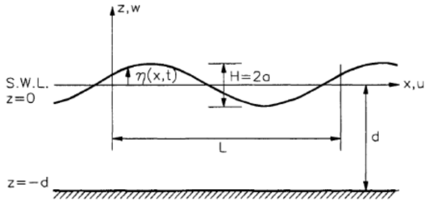
- in air-sea interaction and ocean-mixing models, the wind stress is usually parameterised to directly drive the dynamics of the upper ocean
- wind provides momentum and energy fluxes to the ocean surface and thus mixes the upper ocean
- dominant part of the wind stress, however, is supported by the flux of momentum from wind to waves
- these waves break, and the breaking is regarded as the main source of the turbulence across the interface
- it is often assumed that the turbulence is then diffused down and the mixing is achieved
- if the wave breaking was the only role of the waves in the upper-ocean mixing, such a scheme would perhaps be feasible
- there are, however, two potential problems in such approach

Waves and ocean turbulence

- there are, however, two potential problems in such approach
- first of all, time scales of the turbulence lifetime and turbulence diffusion down to some 100m should agree
- secondly, before the momentum is received by the upper ocean in the form of turbulence and mean currents, it goes through a stage of surface wave motion
- such motion can directly affect or influence the upper-ocean mixing and other processes, and thus ignoring the wave phase of momentum transformation may undermine accuracy and perhaps even validity of such parameterisations
- there are at least two processes in the upper ocean which can deliver turbulence straight to the depth of 100m or so instead of diffusing it from the top
- these are wave-induced turbulence and Langmuir circulation
- 2-3m of the ocean water have the same heat capacity as the entire atmosphere

Linear Wave Theory. Governing equations

•Most fluid mechanics problems can be solved by considering the governing Equations of conservation of mass, momentum and energy



Young, 1999, Elsevier

Define the velocity potential φ

$$u = -\frac{\partial \varphi}{\partial x}, \quad w = -\frac{\partial \varphi}{\partial z}$$

• Laplace Equation (Continuity Equation) - conservation of mass (two-dimensional case):

$$\frac{\partial^2 \varphi}{\partial x^2} + \frac{\partial^2 \varphi}{\partial z^2} = 0$$

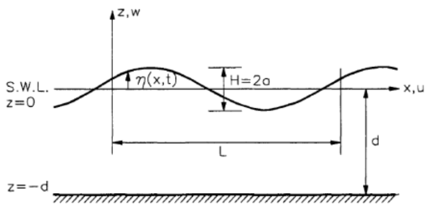
• Unsteady Bernoulli Equation – conservation of momentum:

$$\frac{p}{\rho} + gz - \frac{\partial \varphi}{\partial t} = 0$$

$$\varphi(x,z,t) = \frac{ag}{\omega} \frac{\cosh[k(d+z)]}{\cosh[kd]} \cos(kx - \omega t)$$

Kinsman, 1965: Wind Waves based on Phillips (1961)

Solutions



Navier-Stokes equation

linearised boundary conditions, with surface tension *T*

$$\frac{\partial u}{\partial t} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + \nu\nabla^2 u$$

$$\frac{\partial w}{\partial t} = -\frac{1}{\rho}\frac{\partial p}{\partial z} + \nu\nabla^2 w - g$$

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

$$\frac{\partial \eta}{\partial t} = w_{z=0}$$

$$p - 2\mu\frac{\partial w}{\partial z} = -\frac{\partial^2 \eta}{\partial x^2}\mathrm{T}_{z=\eta}$$

$$\frac{\partial w}{\partial x} + \frac{\partial u}{\partial z} = 0_{z=\eta}$$

vorticity

$$\omega = \frac{\partial w}{\partial x} - \frac{\partial u}{\partial z} = \nabla^2\Psi$$

$$\omega = \beta\frac{i\sigma}{\nu}e^{mz}e^{i(kx+\sigma t)} =$$

$$= -2\gamma k\sigma\exp\left(\sqrt{\frac{\sigma_{real}}{2\nu}}z - \frac{2\sigma_{real}}{\mathrm{Re}_w}\right)\exp\left\{i\left(kx + \sqrt{\frac{\sigma_{real}}{2\nu}}z + \sigma_{real}t\right)\right\}$$

$$\frac{\delta_z}{\lambda} = \frac{1}{\lambda}\sqrt{\frac{2\nu}{\sigma_{real}}} = \frac{1}{2\pi}\sqrt{\frac{2\nu k^2}{\sigma_{real}}} = \frac{\sqrt{2}}{2\pi}\frac{1}{\sqrt{\mathrm{Re}_w}}$$

- exponential decay in *z* and *t*
- oscillations in *x*, *z* and *t*
- 'length' of vertical vorticity oscillation is much smaller than λ

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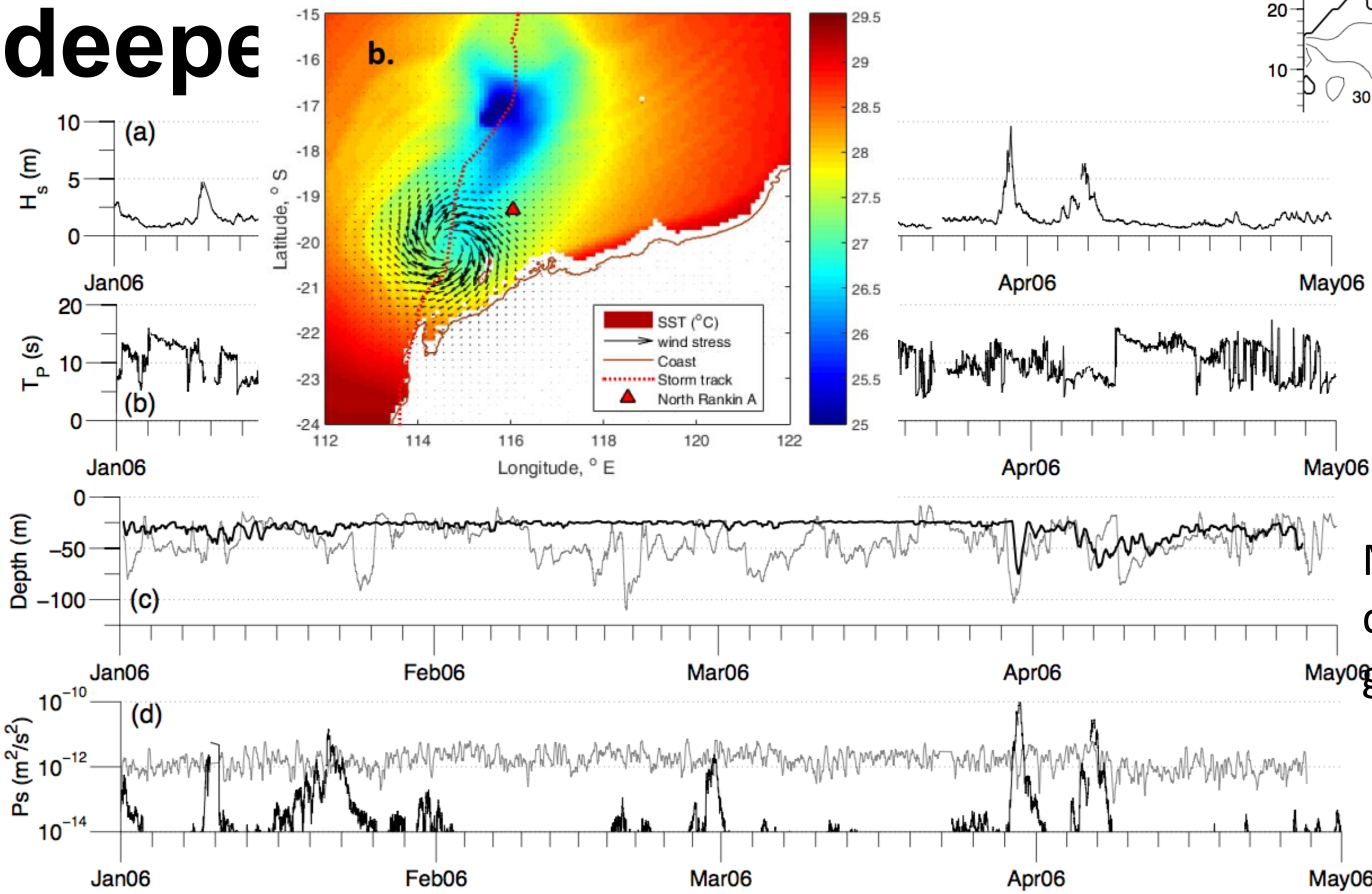
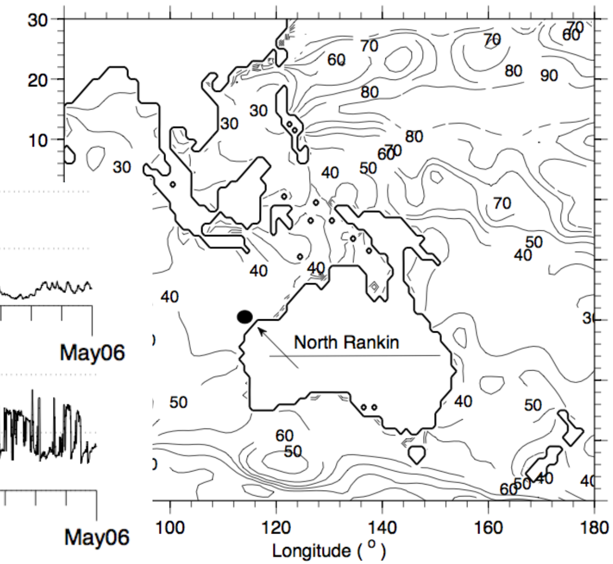
Field observations, North Rankin mixed layer deepe





MLD:
dark – wave induced
grey - measured

Toffoli et al., *JGR*, 2012, Aijaz et al., *JGR*, 2017

Sensitivity test - *wave induced mixing only*

- mixed layer deepened and a substantial amount of heat was brought down into the ocean
- most notably between the depths of 25–100 m and the latitudes 40S–20S
- Upper Ocean Heat Content (OHC) increases by approximately ~0.1%. This was three times the size of the expected climate change signal by the end of the 21st century under the RCP4.5 scenario
- the modified projection, however, showed about 3% less ocean heat uptake than the standard
- although wave-induced mixing results in reduced global ocean heat uptake, there were regions in which OHC substantially increased, particularly the North Atlantic sub-tropics, the South Atlantic around the tip of South Africa, the Tasman Sea, and the Sea of Japan
- All effects for completeness: (1) momentum budget aloft (2) wave-induced currents, (3) wave-induced mixing, (4) heat fluxes, (5) mass flux, (6) albedo, and (7) sea ice

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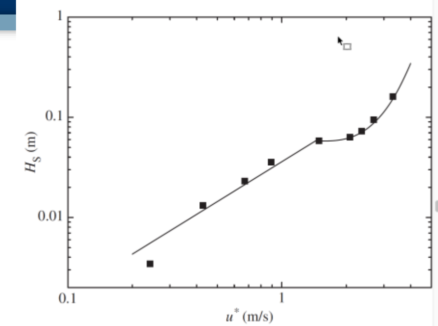
Mass transfer velocity across the breaking air-water interface at extremely high wind speeds



- *Iwano et al., 2013, Tellus B*

Generally, CO_2 flux F between atmosphere and ocean is estimated by the following bulk equation:

$$F = k_L S \Delta p CO_2, \tag{1}$$

where k_L is the mass transfer velocity of CO_2, S is the solubility of CO_2 in water, and ΔpCO_2 is the difference in the partial pressure of CO_2 between atmosphere and ocean.

in the lab tests, at U10 ~ 34 m/s

- mass transfer kL increases drastically
- volume flux of droplets changes sharply
- growth rate of significant wave height drops suddenly (top right)

Authors conclude: This change indicates that intense wave breaking occurs at extremely high wind speeds and it has significant effects on CO2 transfer

We conclude: Wind input drops too (at least in relative terms)

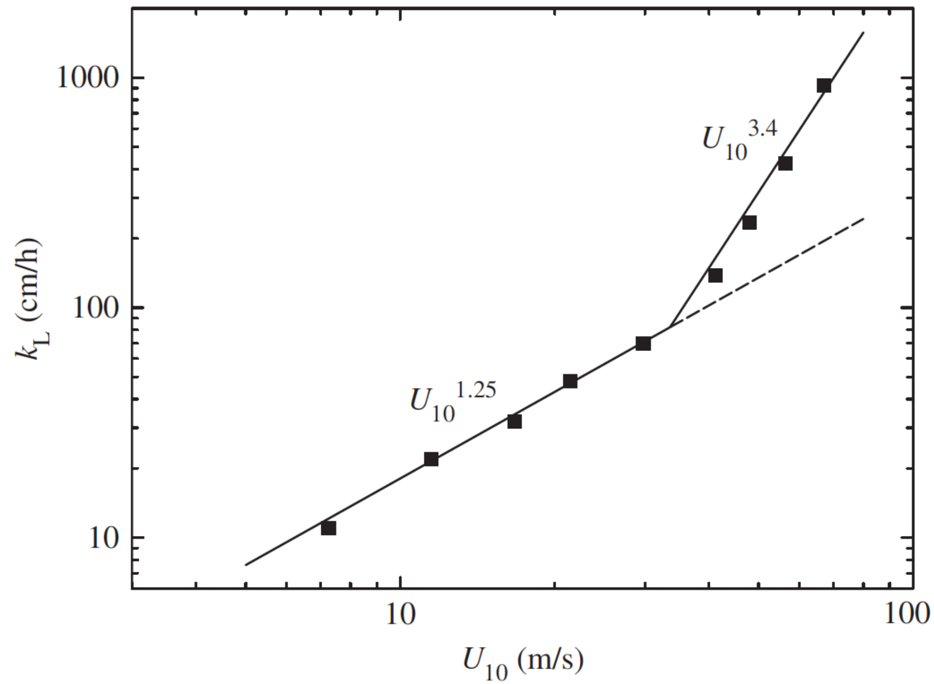


Fig. 3. Mass transfer velocity k_L against wind speed at 10 m height U_{10}.

$$k_L = \begin{cases} 1.02 U_{10}^{1.25} & (U_{10} < 33.6 \text{ m s}^{-1}) \\ 5.32 \times 10^{-4} U_{10}^{3.4} & (U_{10} \geq 33.6 \text{ m s}^{-1}) \end{cases},$$