MSE:
Transforming the Future of Engineering & IT

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Ocean Waves as a Missing Link Between Atmosphere and Ocean

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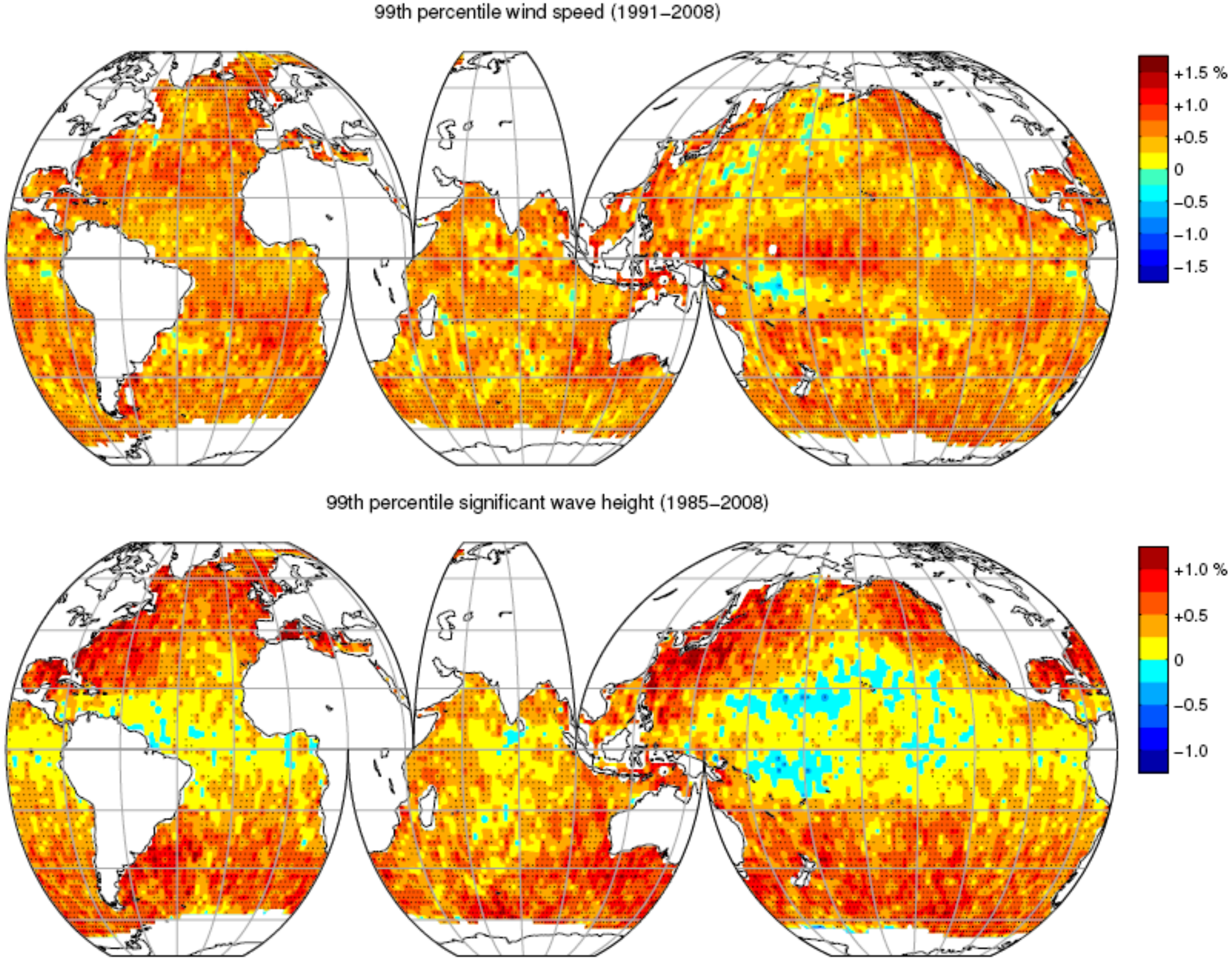
IWMO, Santos, Brazil
26 June 2018

Motivation

Waves influences the climate, climate affects the waves



Young, Zieger, Babanin, Science, 2011

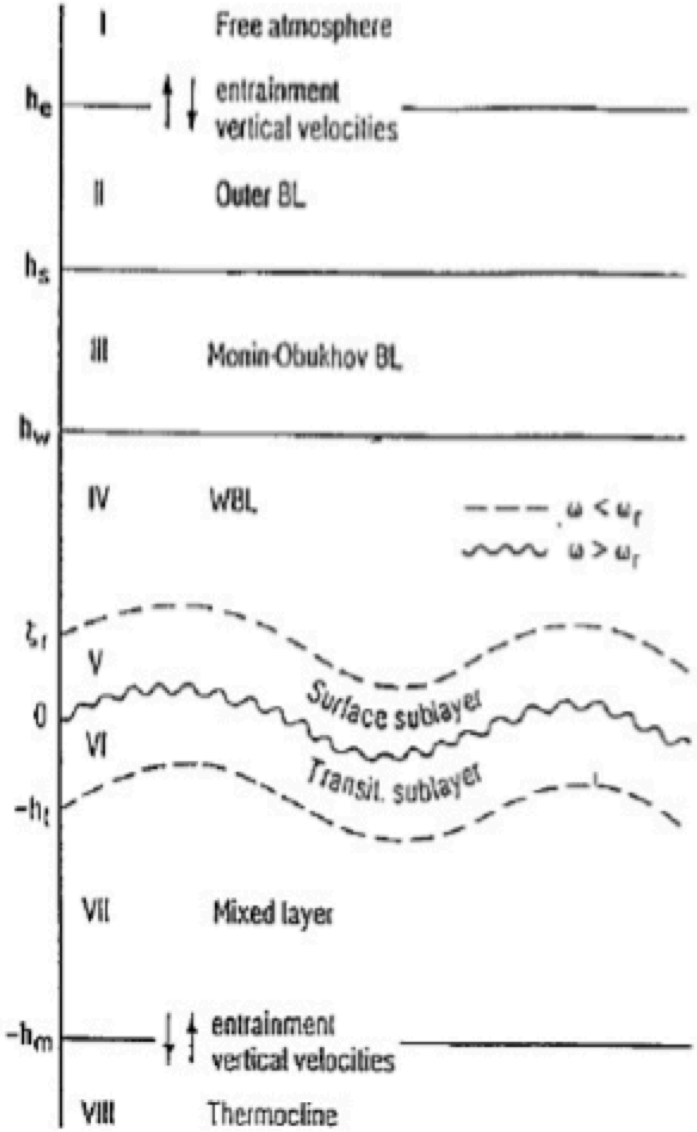
Waves as Atmosphere/Ocean Link

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

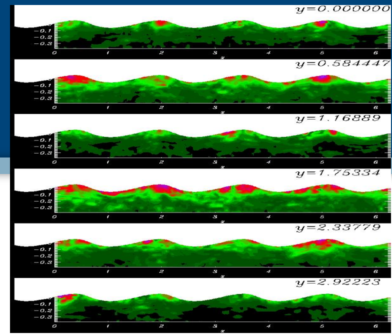
- Atmospheric boundary layer
  - winds generate waves
  - waves provide surface roughness and change the winds
  - waves evolve, fluxes change
- Upper ocean mixed layer
  - waves generate currents
  - produce turbulence
  - turbulence: moderates and facilitates mixing
  - changes the circulation, SST

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

- new research field
- sediment suspension, tropical cyclones, weather, climate

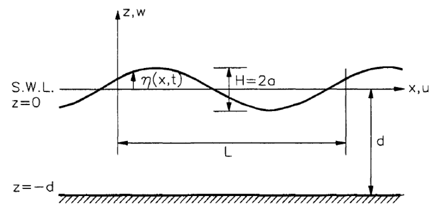
Waves and ocean turbulence

- 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 and momentum across the interface
- the turbulence is then diffused down and the mixing is achieved, momentum goes to the mean currents
- 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

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)

Navier-Stokes equation

linearised boundary conditions, with surface tension *T*

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

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

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

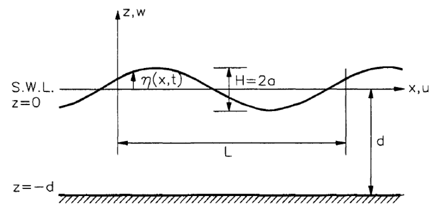
$$\frac{\partial \eta}{\partial t} = w_{z=0}$$

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

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

Solutions

vorticity



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

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

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

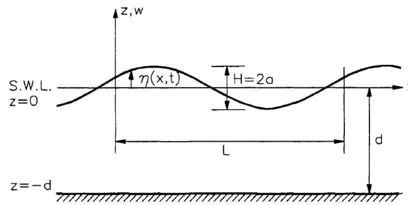
$$\frac{\delta_z}{\lambda} = \frac{1}{\lambda}\sqrt{\frac{2\nu}{\sigma_{real}}} = \frac{1}{2\pi}\sqrt{\frac{2\nu k^2}{\sigma_{real}}} = \frac{\sqrt{2}}{2\pi}\frac{1}{\sqrt{\mathrm{Re}_w}}$$

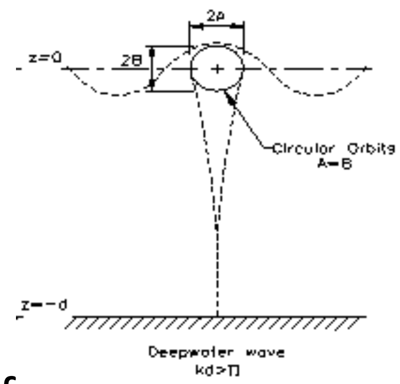
- exponential decay in *z* and *t*
- oscillations in *x*, *z* and *t*
- ‘length’ of vertical vorticity oscillation is much smaller than *λ*

Hypothesis of the Wave Reynolds Number

$\eta(x,t) = a_0 \cos(\omega t + kx)$

$a(z) = a_0 \exp(-kz)$





It is the hypothesis that the a-based Reynolds number

$$\mathrm{Re} = \frac{aV}{\nu} = \frac{a^2\omega}{\nu}$$

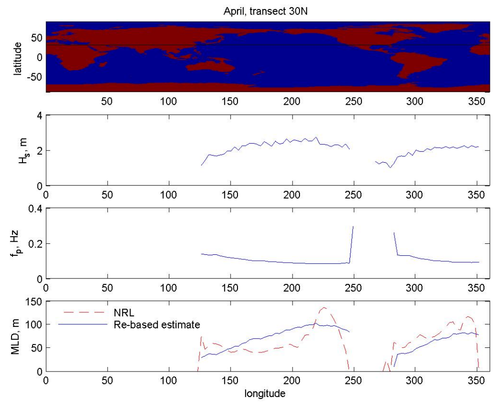
where $V=\omega a$ is orbital velocity, and ν is kinematic viscosity of the ocean water, indicates transition from laminar orbital motion to turbulent

Critical Reynolds Number for the Wave-Induced Motion, and Depth of the Mixed Layer

$Re_{cr}=3000$

$$\mathrm{Re}(z) = \frac{\omega}{\nu} a_0^2 \exp(-2kz) = \frac{\omega}{\nu} a_0^2 \exp\left(-2\frac{\omega^2}{g} z\right)$$

$$z_{cr} = -\frac{1}{2k} \ln\left(\frac{\mathrm{Re}_{cr}\,\nu}{a_0^2\omega}\right) = \frac{g}{2\omega^2} \ln\left(\frac{a_0^2\omega}{\mathrm{Re}_{cr}\,\nu}\right)$$



Babanin, GRL, 2006

Dai et al., JPO, 2010

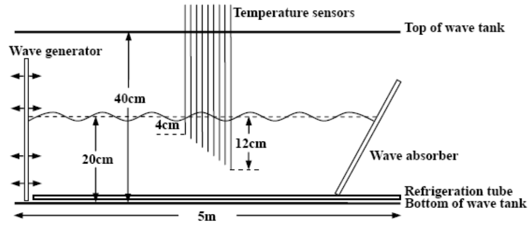


Figure 1. Sketch of the laboratory setup.

Mixing the stratified fluid
experiment (left), model (right)

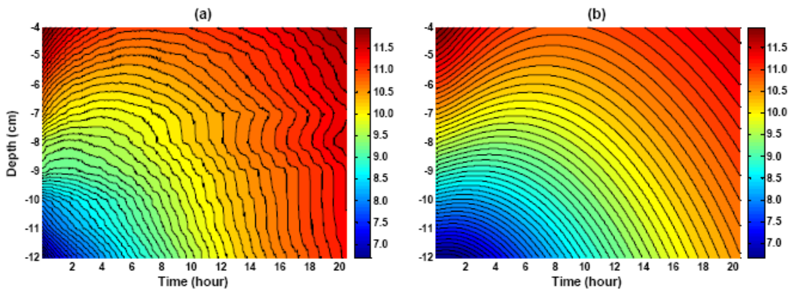
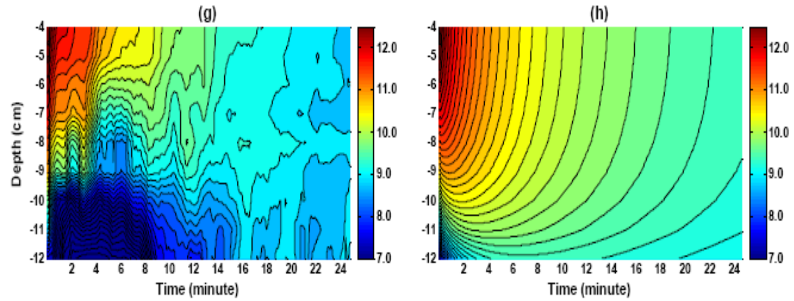


Figure 2. Evolution of the water-temperature profile without waves. (a) observations; (b) numerical simulation with the one-dimensional model. The time is in hours.



no waves
time scale: hours

non-breaking waves
time scale: minutes

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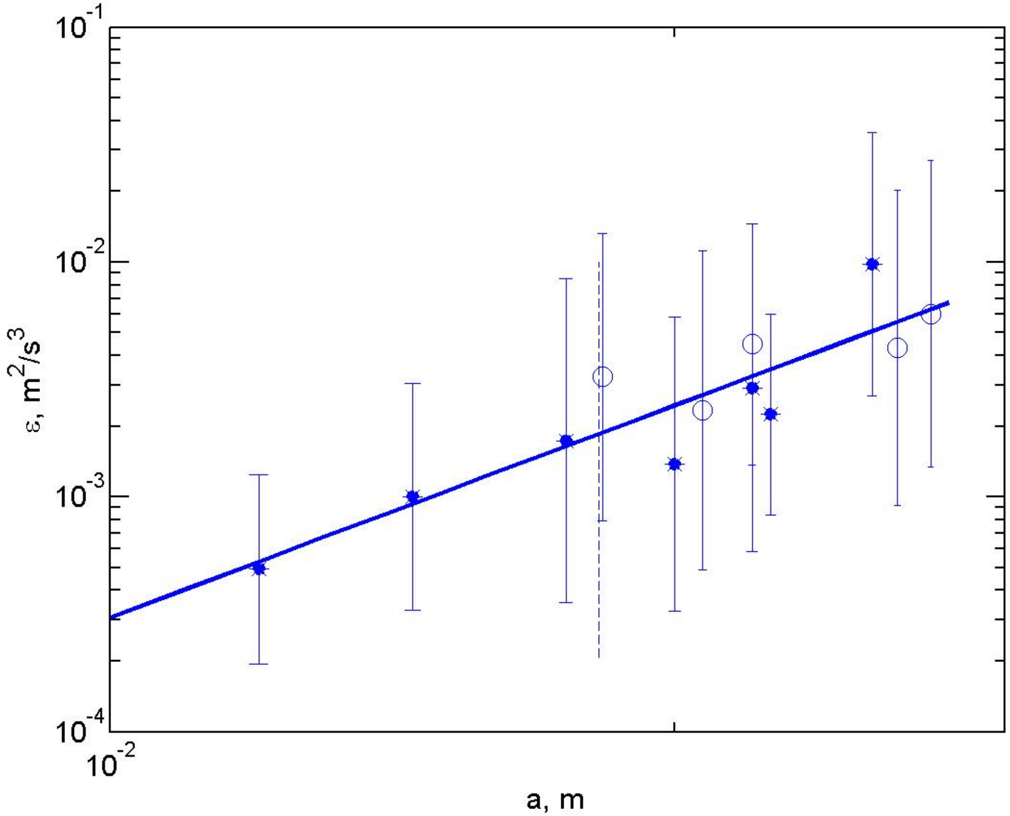
Laboratory Experiment, ASIST, RSMAS, University of Miami



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

This is close to the expectation: since the force due to the turbulent stresses is proportional to a^2, the energy dissipation rate should be ~a^3.

Babanin & Haus, JPO, 2009

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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

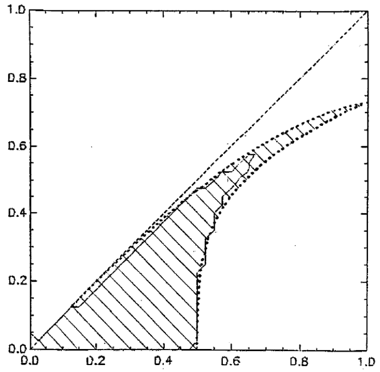


Figure 1. Diagram of stability of solutions of equation set (22) on the (ε_1, ε_2) plane. Here ε_1 is the horizontal axis and ε_2 is the vertical axis. F1 corresponds to $\varepsilon_2 = \varepsilon_1$; F2$_1$ corresponds to $\varepsilon_2 = \varepsilon_1 - 0.3\ \varepsilon_1^3 + 0.03\ \varepsilon_1^4$, $0 \le \varepsilon_1 \le 1$; 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.

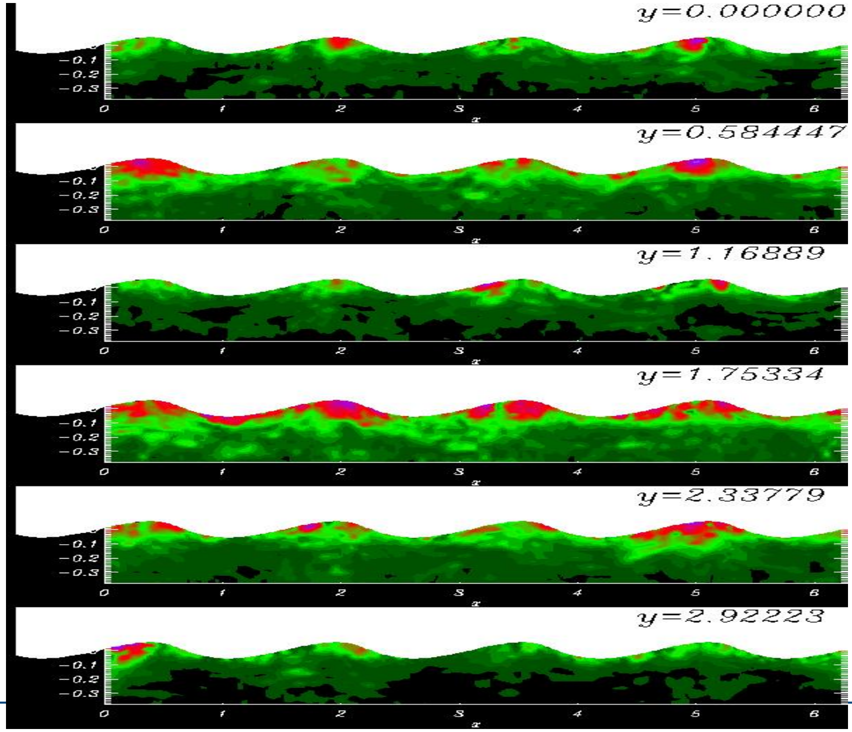
[1] The turbulence (the random vortex motions) of the upper ocean is nourished by the energy and momentum of the surface waves (the potential motion). The statistical characteristics of the turbulence (turbulent kinetic energy, dissipation rate, and Reynolds stresses) depend on the state of the ocean 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.

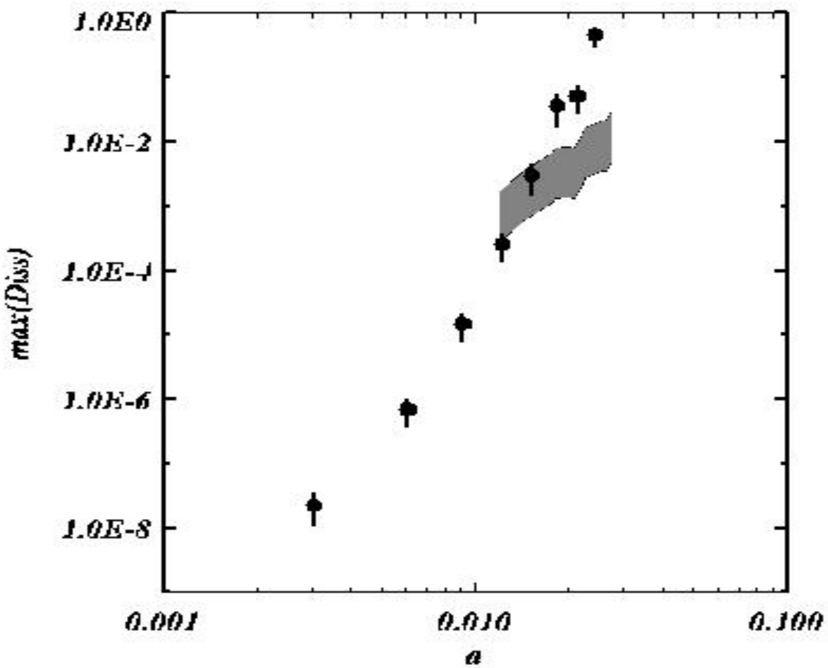
Citation: Benilov, A. Y. (2012), On the turbulence generated by the potential surface waves, *J. Geophys. Res.*, *117*, C00J30, doi:10.1029/2012JC007948.

Benilov, JGR, 2012

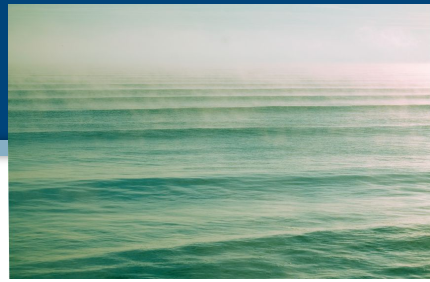
Model of generation of turbulence by nonlinear waves

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





Swell attenuation



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

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

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

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

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

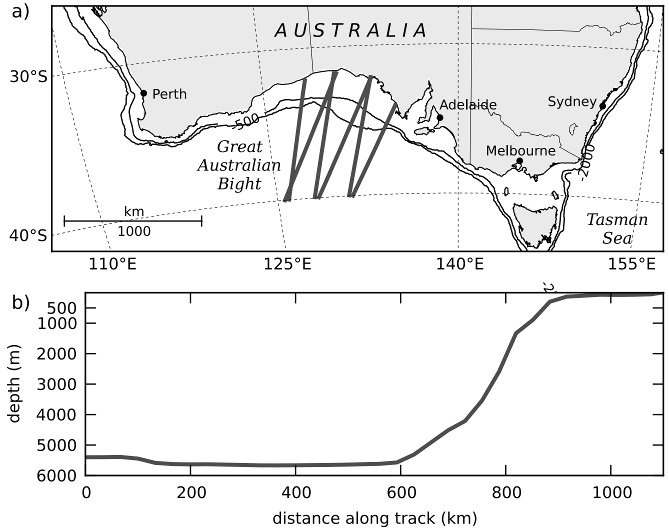
$$a_0(x)^2 = \frac{4}{B^2} x^{-2} = \frac{9}{4 \cdot b_1^2 k^4} x^{-2} = \frac{9}{64} 10^6 k^{-4} x^{-2}.$$

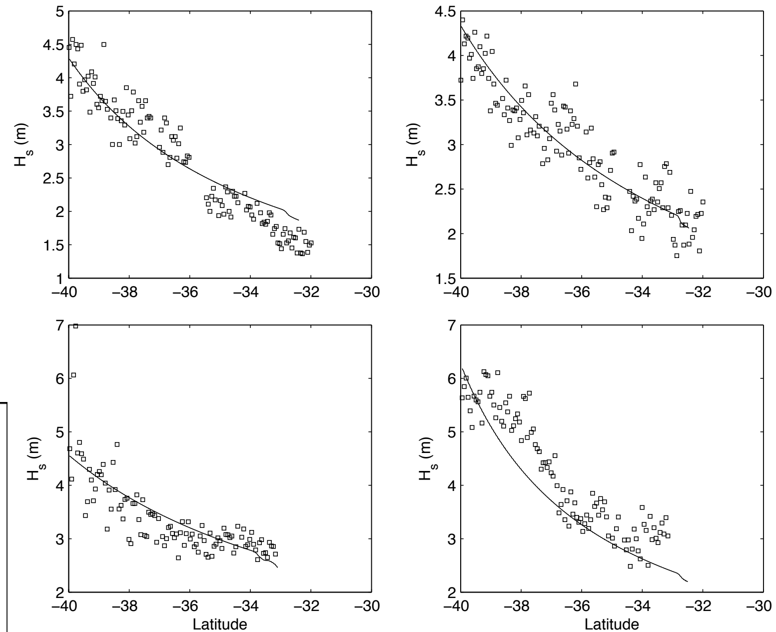
Dissipation

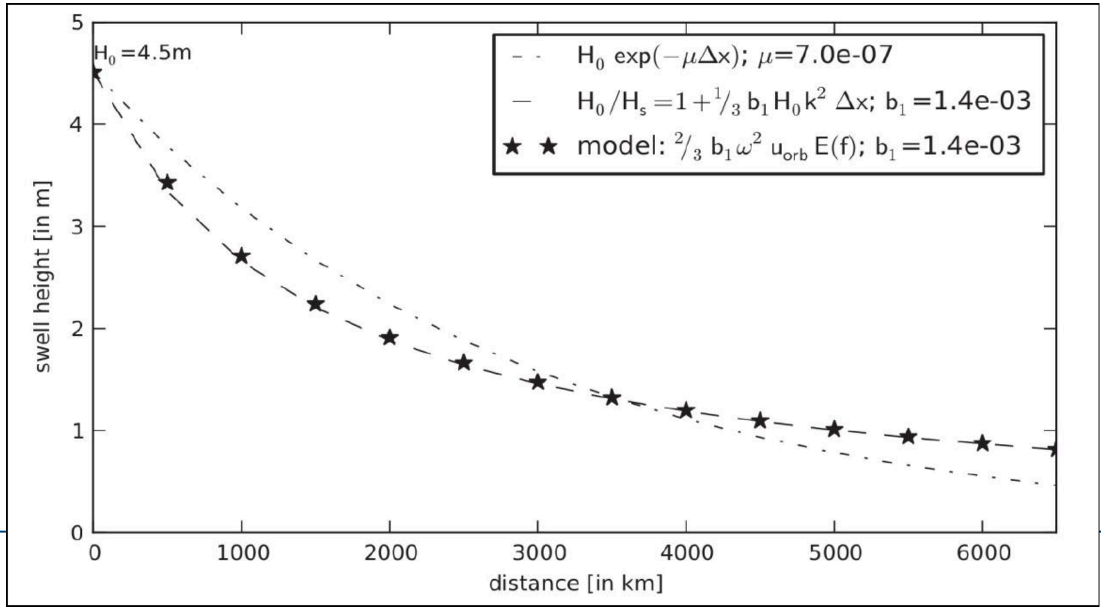
- volumetric
- per unit of surface
- per unit of propagation distance

Babanin, 2011, CUP

Swell attenuation

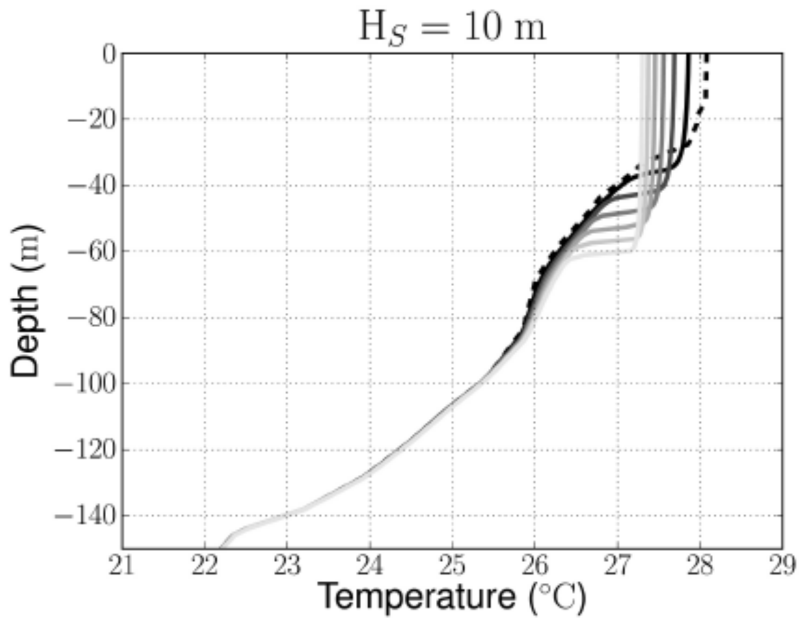


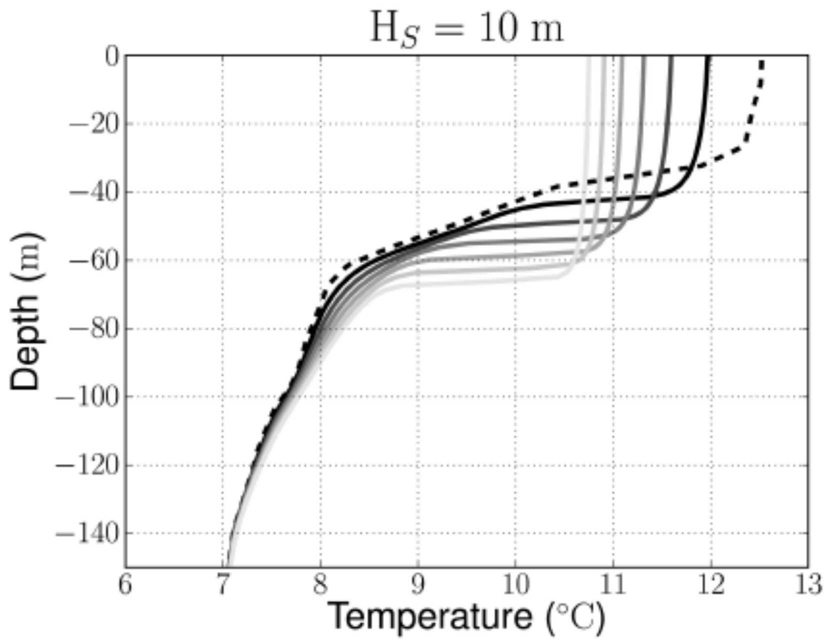




Young, Babanin, Zieger, JPO, 2013

Modelling SST and MLD at the scale of tropical cyclone





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

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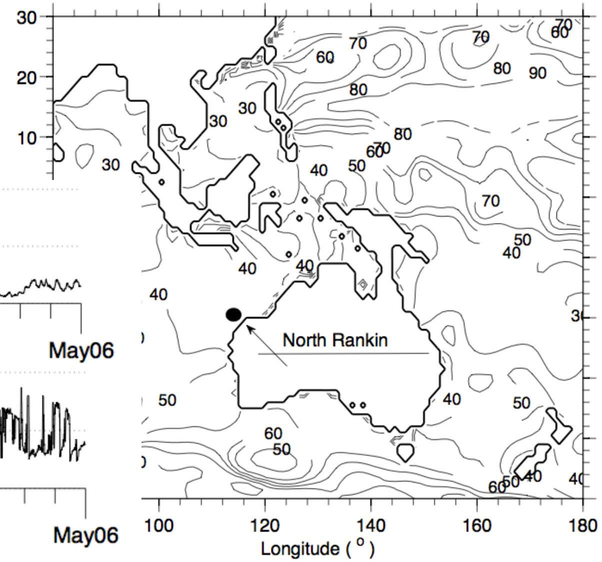
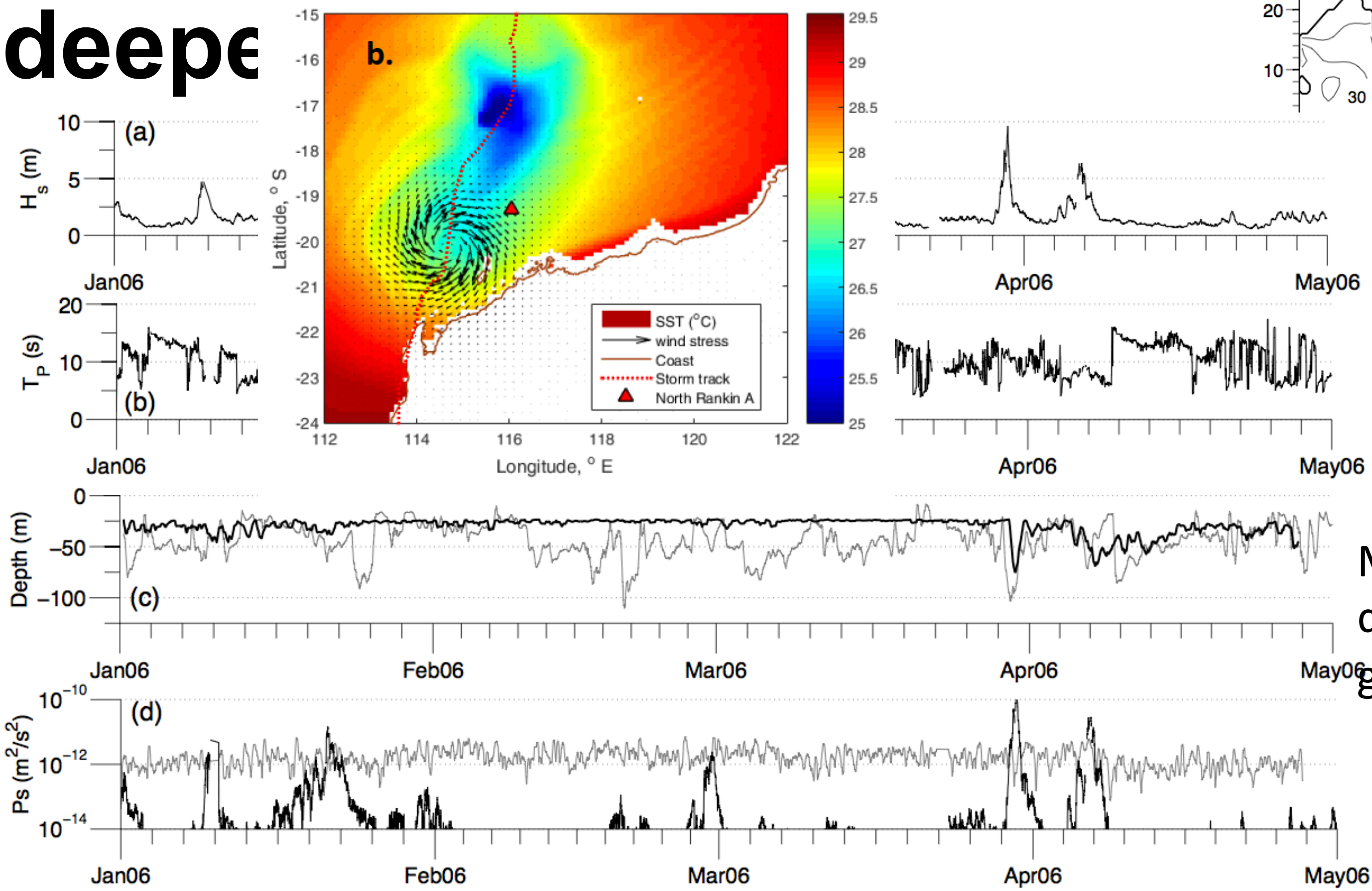
Field observations, North Rankin mixed layer deepe





MLD:
dark – wave induced
grey - measured

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

Field observations, North Sea, sediment suspension

Pleskachevski et al., JPO, 2011

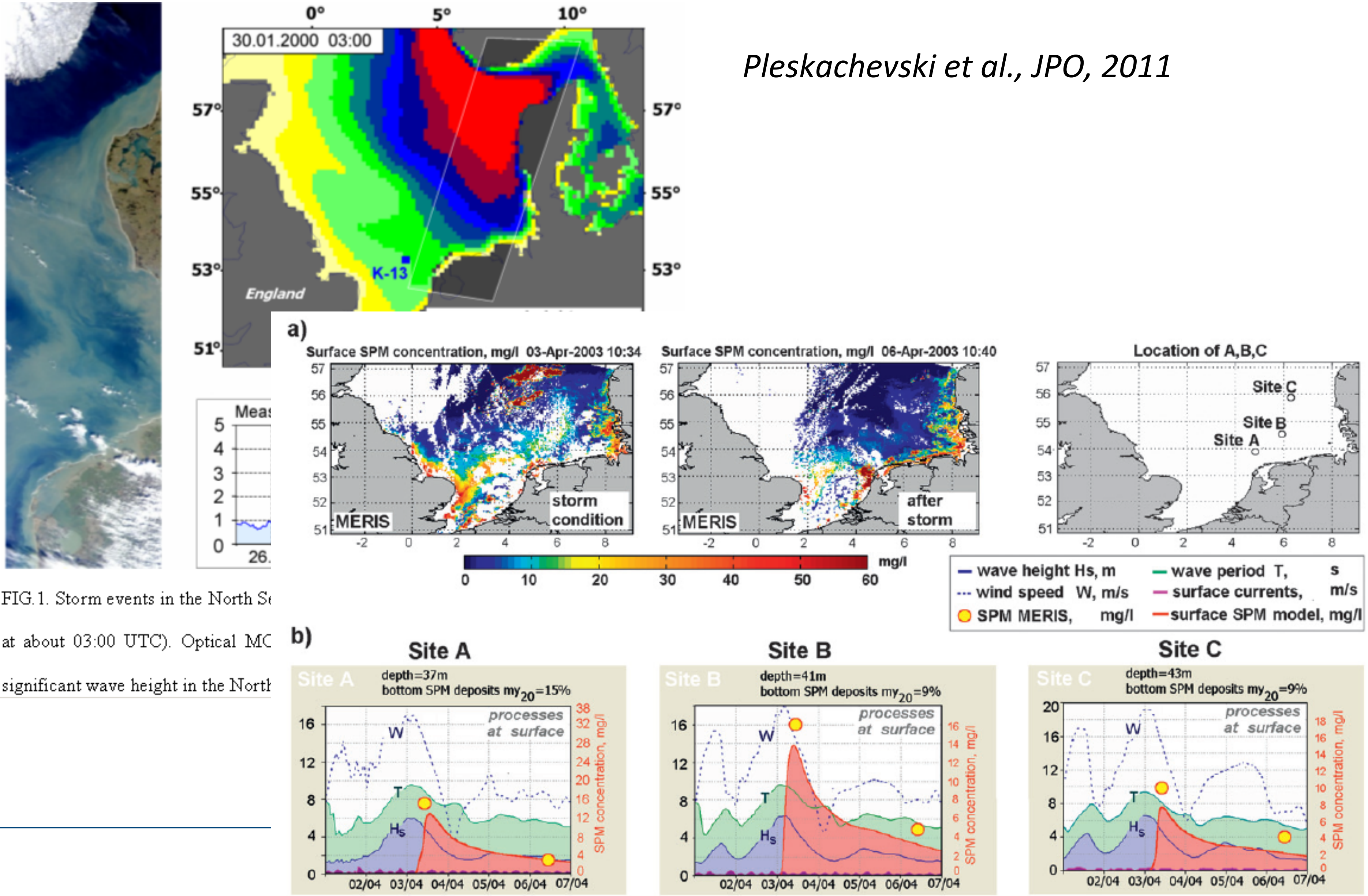
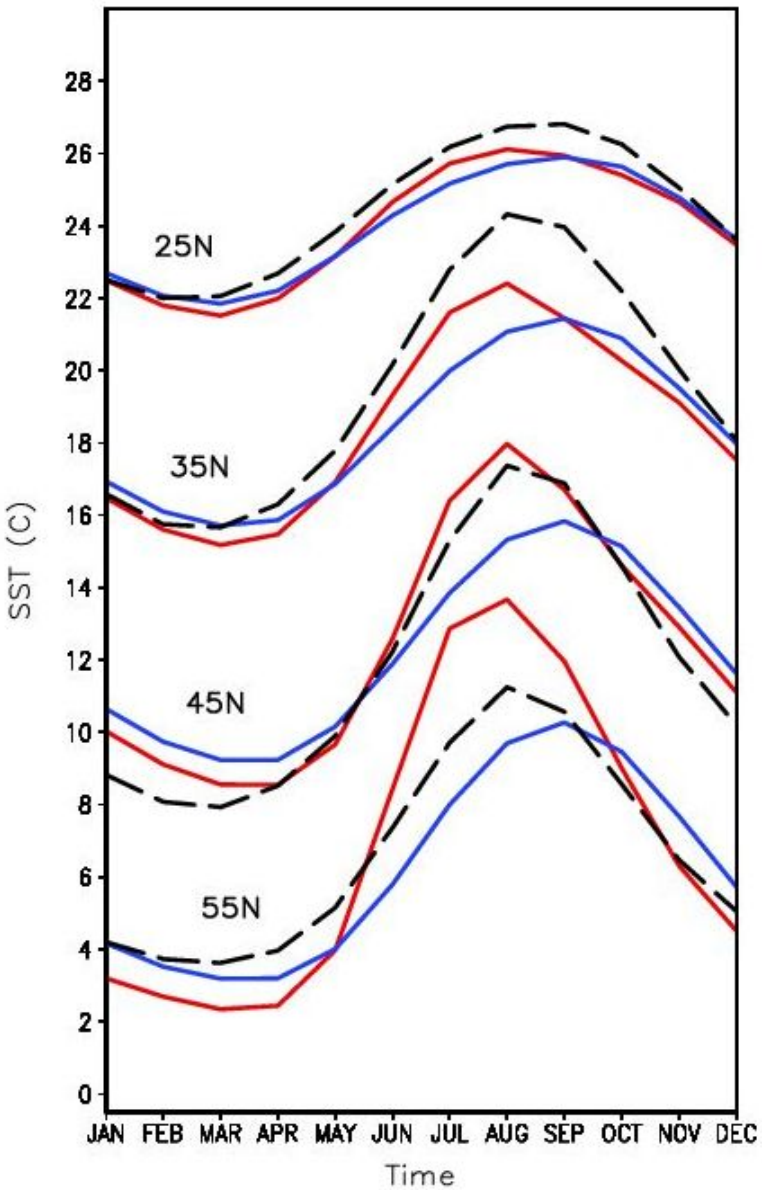


FIG. 1. Storm events in the North Se
at about 03:00 UTC). Optical MC
significant wave height in the Nort

Implementing wave-induced mixing in CLIMBER



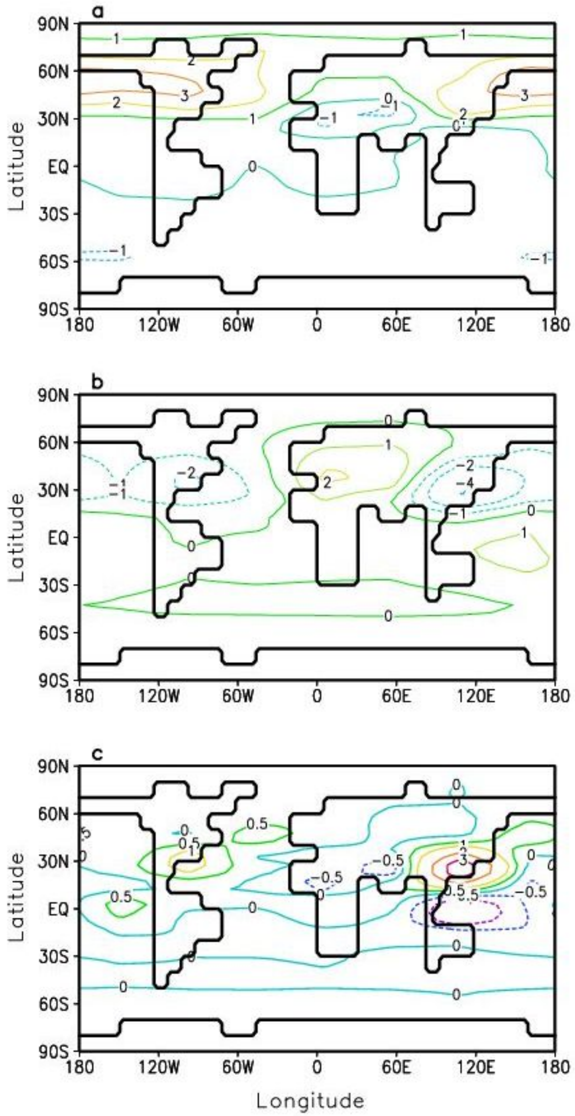
- Seasonal trend of the global zonally averaged SST. Panels shown: 25, 35, 45 and 55 degrees North (from top to bottom). Lines shown: default version of CLIMBER (blue), variable MLD (red) and observations based on Levitus data (black).

- effect is essential outside the tropical areas

- both magnitudes and phases of SST are imporved

Babanin, Ganopolski, Phillips, 2009, OM

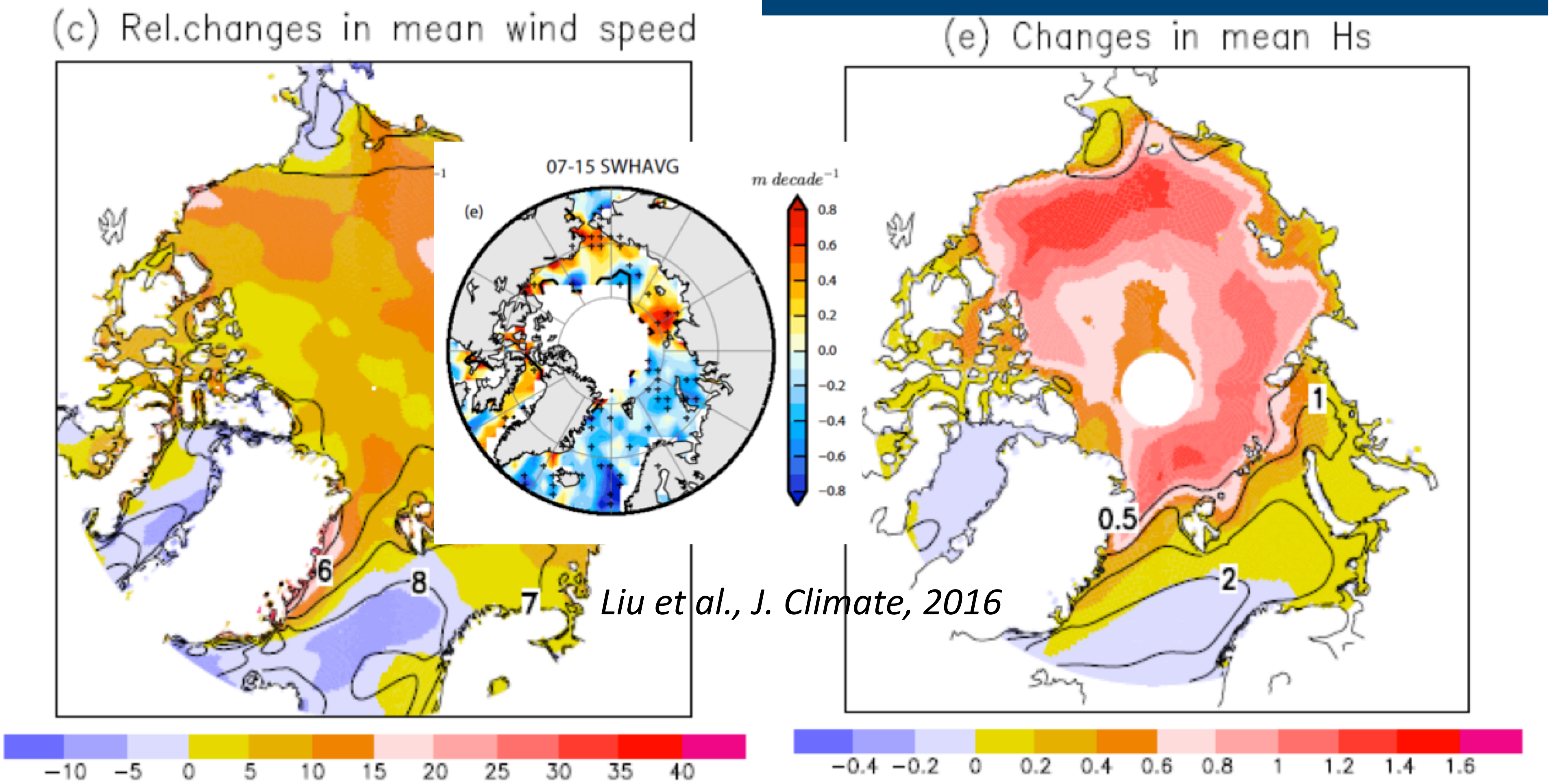
Implementing wave-induced mixing in CLIMBER



Global distribution (Northern summer)

- temperature (*degrees*)
- pressure (*mbar*)
- precipitation (*mm* per day)

Comparison of climate modelling (2046-2065 versus 1980-1999, in %, with satellite observations (insert))



Liu et al., J. Climate, 2016

Wind **Waves**

September. **Coupled climate, ice and wave models**

Khon et al., Geophys. Res. Lett., 2013

Wave-induced mixing

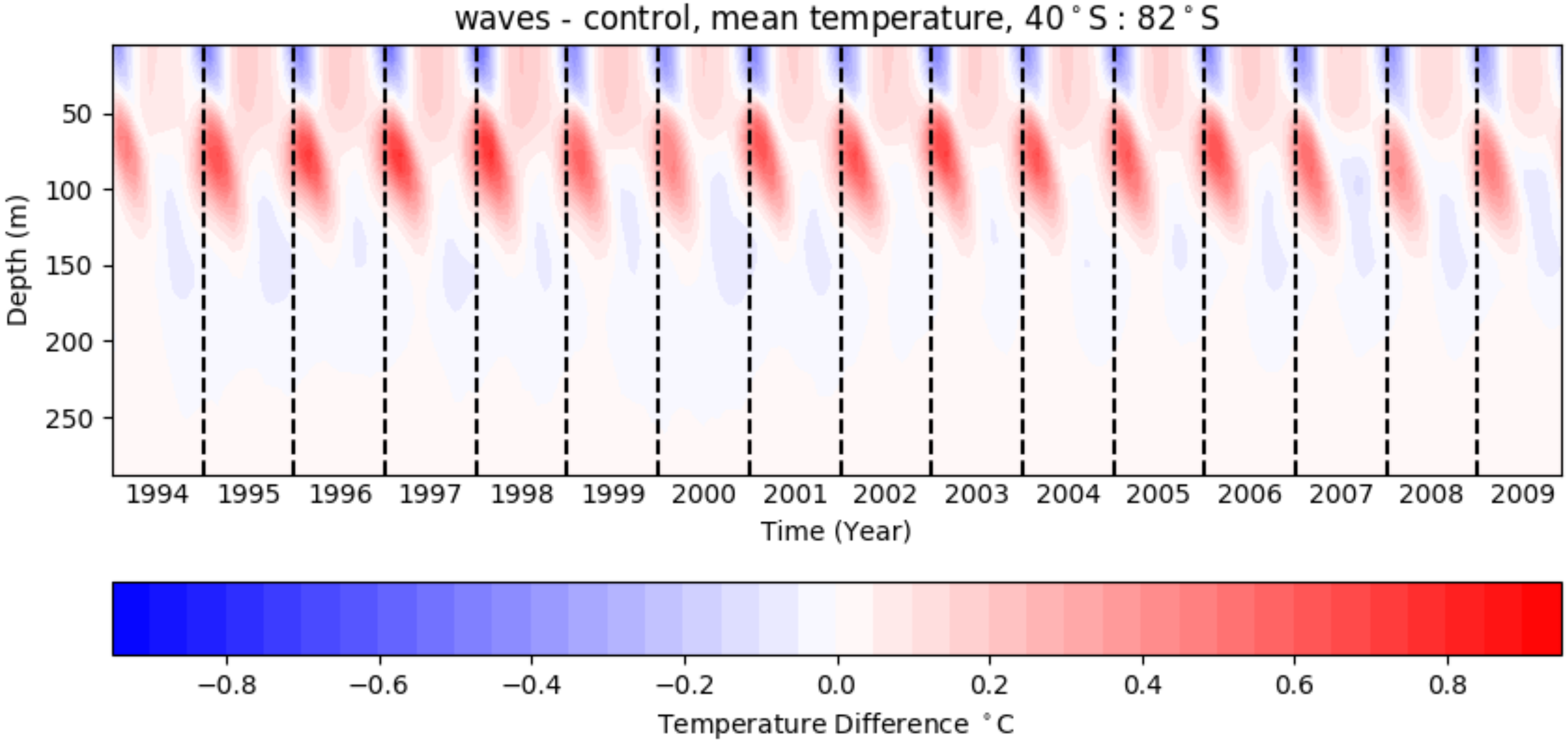
- Wave-induced mixing (MOM5)
- One-way coupling with WW3

In the figures (subsequent slides):

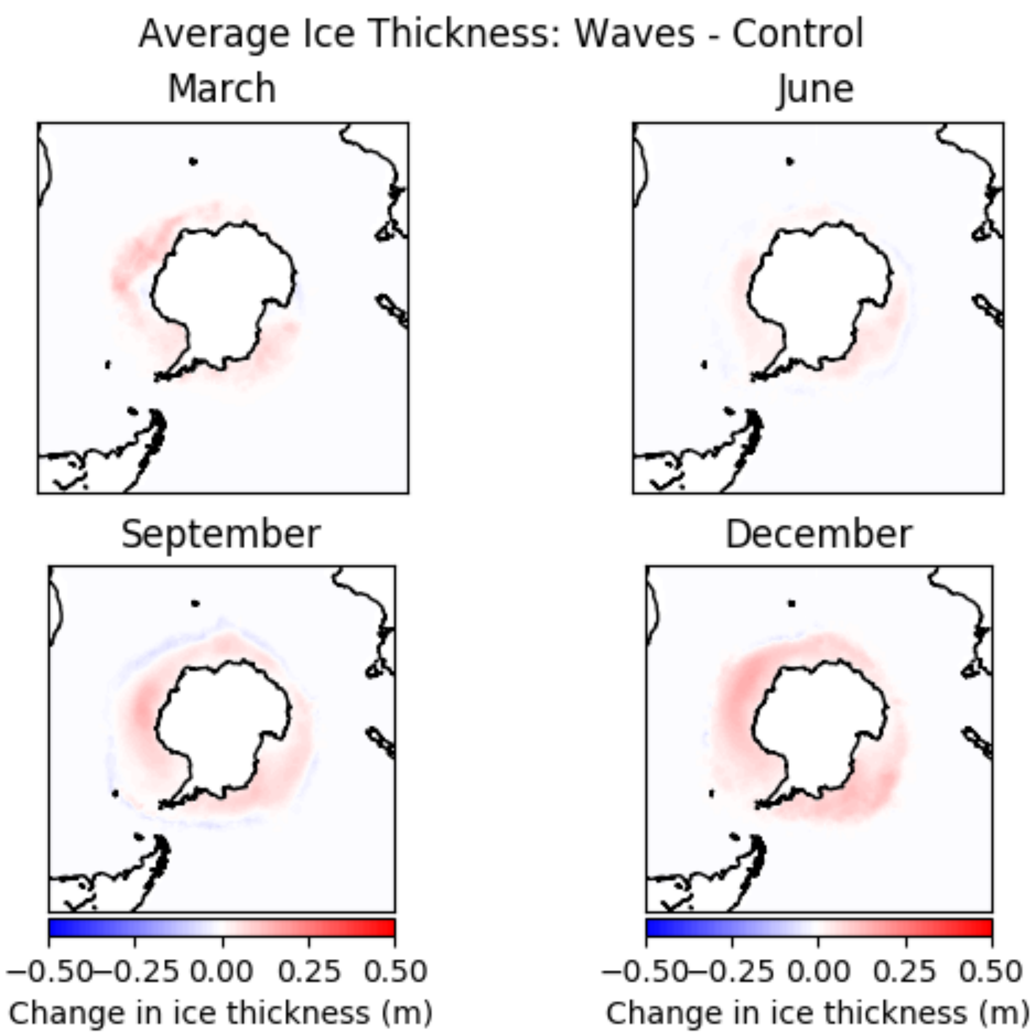
- the anomalies are calculated by taking the difference between the simulation with waves and the simulation without waves
- the seasonal values come from a 10 year mean

Results from an Ocean model with and without the inclusion of an extra wave mixing term.

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.



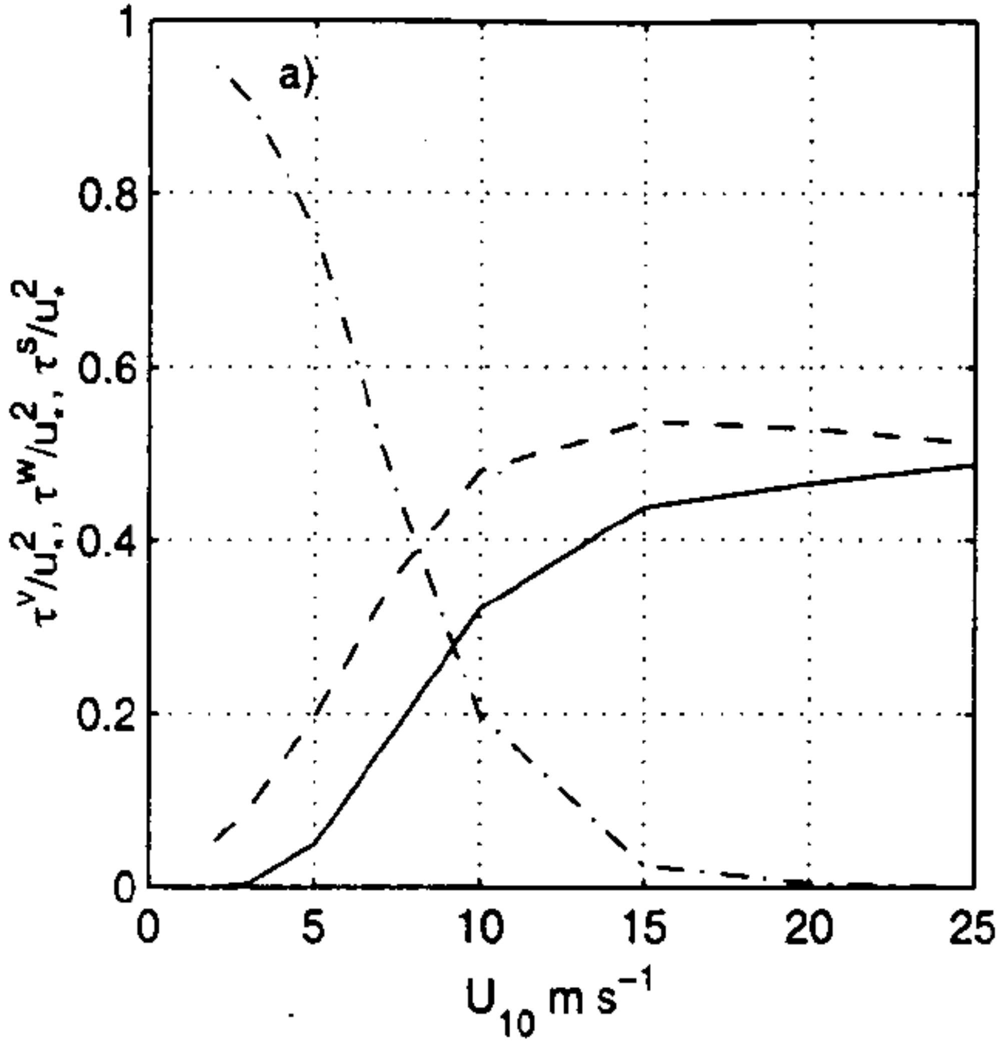
Difference in ice thickness between ocean models with and without the extra wave-mixing term (wave and control, respectively).

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

- coupling of small-scale models (waves, turbulence) with large-scale models (weather, climate) is necessary
  - physics is continuous
  - computing capabilities allow the coupling
- waves provide feedback
  - to the atmospheric boundary layer
  - to the upper ocean (usually overlooked)
  - to the large-scale air-sea interactions
- wave climate also changes

Momentum flux to currents and waves (through slope-coherent pressure and breaking)



Kudryavtsev-Makin, 2011, BLM