Wave-Coupled Processes in Air-Sea Modelling: from Turbulence to Climate

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Motivation

- Large-scale air-sea processes are coupled with the ocean waves
- This includes weather, tropical cyclones, ocean currents, upper-ocean mixing, climate
- Waves, in turn, change or fluctuate at long scales
- Until recently, turbulence produced by the orbital motion of surface waves was not accounted for, and this fact limits performance of the models for atmospheric boundary layer, air-sea interactions and upper-ocean dynamics

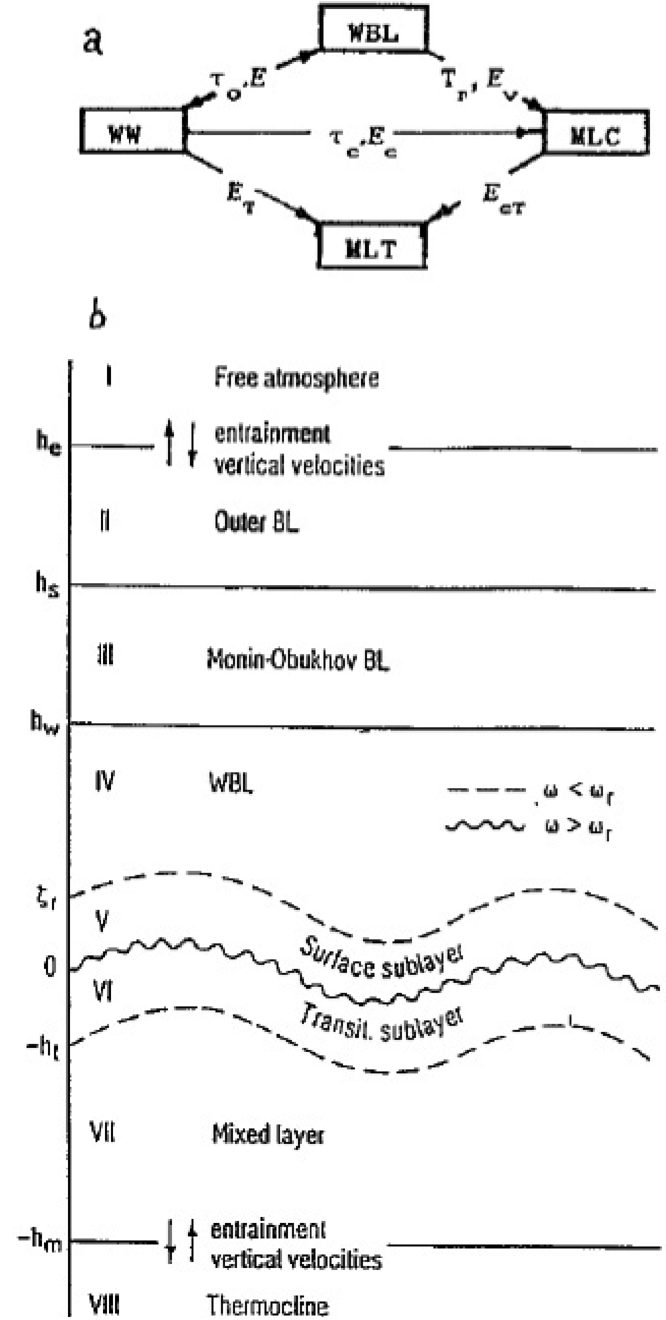
Introduction

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
  - generate currents
  - produce turbulence
  - turbulence: moderate and facilitate mixing
  - changes the circulation, SST

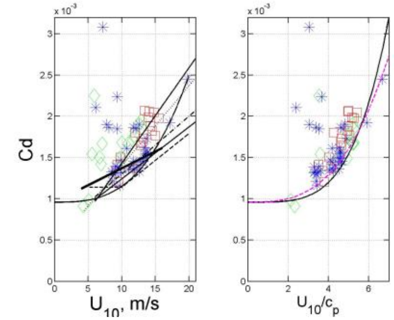
Tradition and future

- 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

Waves and air-sea interactions



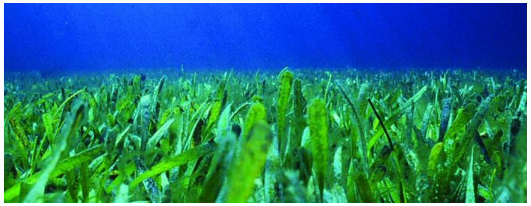
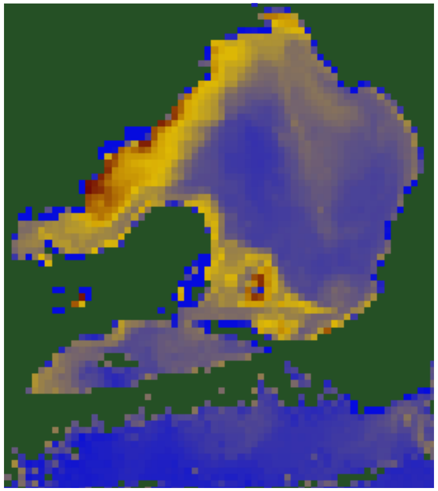
- in air-sea interaction and ocean-mixing models, the wind stress is usually parameterised to directly drive the dynamics of the upper ocean
- ~90% of the flux, however, first input into the waves
- air-sea coupling is usually parameterised in terms of the drag coefficient C_d

$$t = r_a \overline{u'w'} = r_a u_*^2 = r_a C_d U_{10}^2$$

- the parameterisation relies on the concept of constant flux layer
- C_d is routinely parameterised in terms of wind speed U_{10}
- scatter has not improved over some 30 years
- coupling with wave models is necessary

Upper-ocean mixing

- Wave-induced mixing
- dissolved gases
- nutrients
- water temperature and stratification
- Direct influence in finite-depths:
- sediment suspension
- impact on corals, sea grass



Turbulence
in the water

Dai, Qiao, Sulisz, Han, Babanin, 2010, JPO

Laboratory Experiment, First Inst. of Oceanography, China

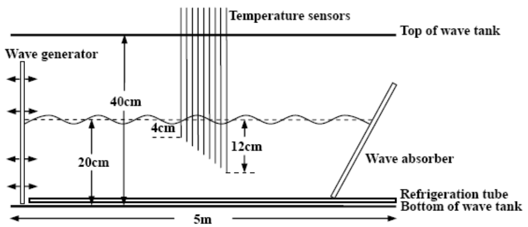


Figure 1. Sketch of the laboratory setup.

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

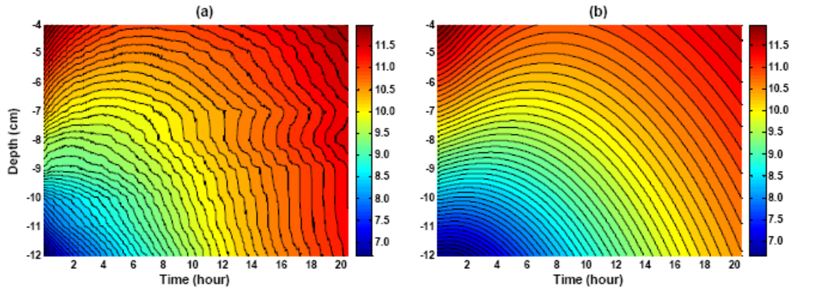
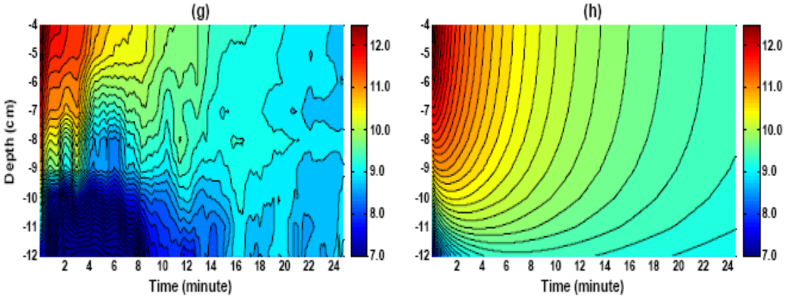


Figure 2. Evolution of the water-temperature profile without waves. (a) observations; (b) numerical simulation with the one-dimensional model. The time is in hours.

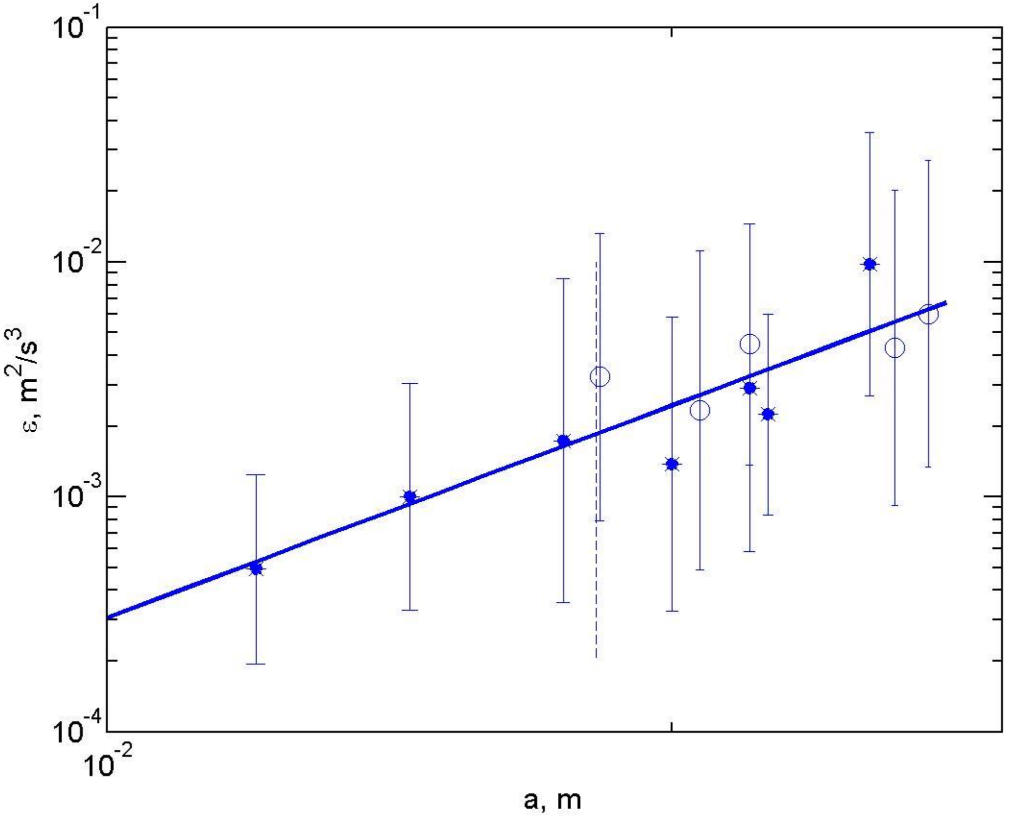


no waves
time scale: hours

non-breaking waves
time scale: minutes

Babanin & Haus, 2009, JPO

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.

Model of generation of turbulence in potential waves

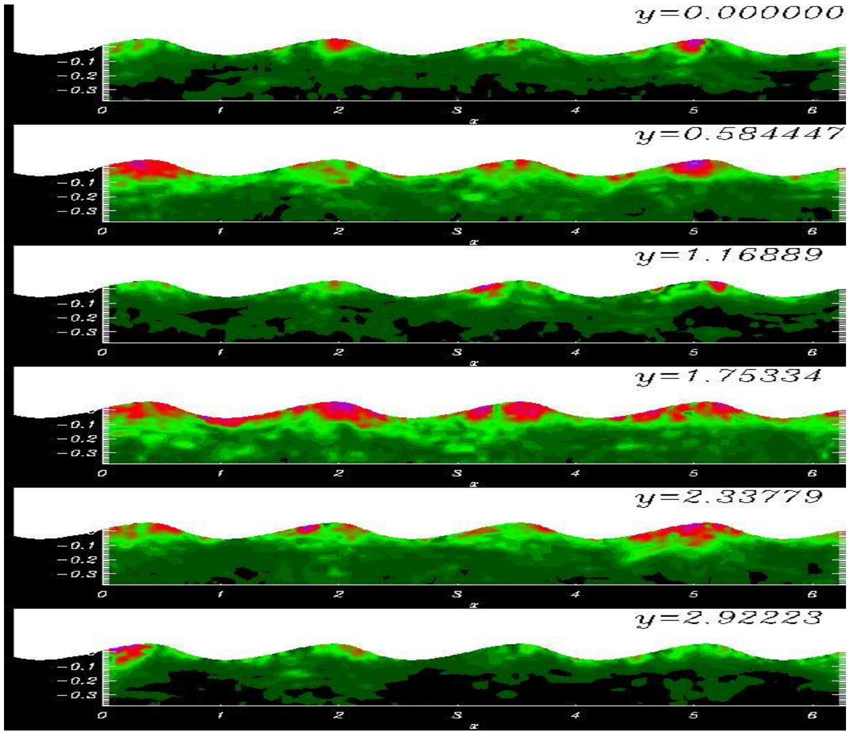
- Regardless of the turbulence source, 3D turbulence is unstable to 2D wave orbital motion *(Benilov, 2012, JGR)*
- Model is based on exact 2-D (x-z) model of surface waves coupled with 3-D LES (x-y-z) model of vortical motion based on Reynolds equation with parameterised subgrid turbulence
- Both systems of equations are written in conformal cylindrical surface-following coordinates
- The one-way coupling of models occurs through components of potential orbital velocity and vorticity components

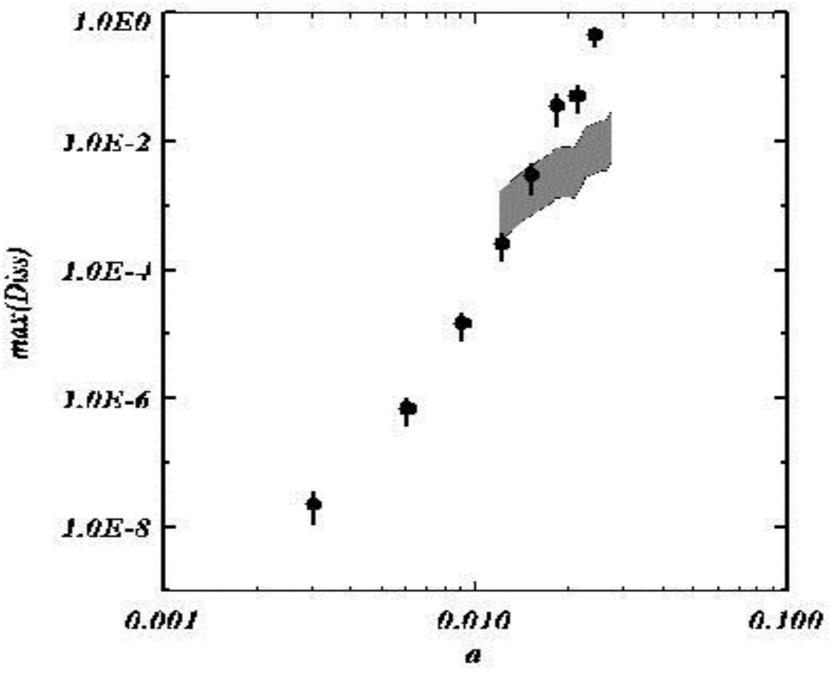
Babanin & Chalikov, 2012, JGR

Babanin & Chalikov, 2012, JGR

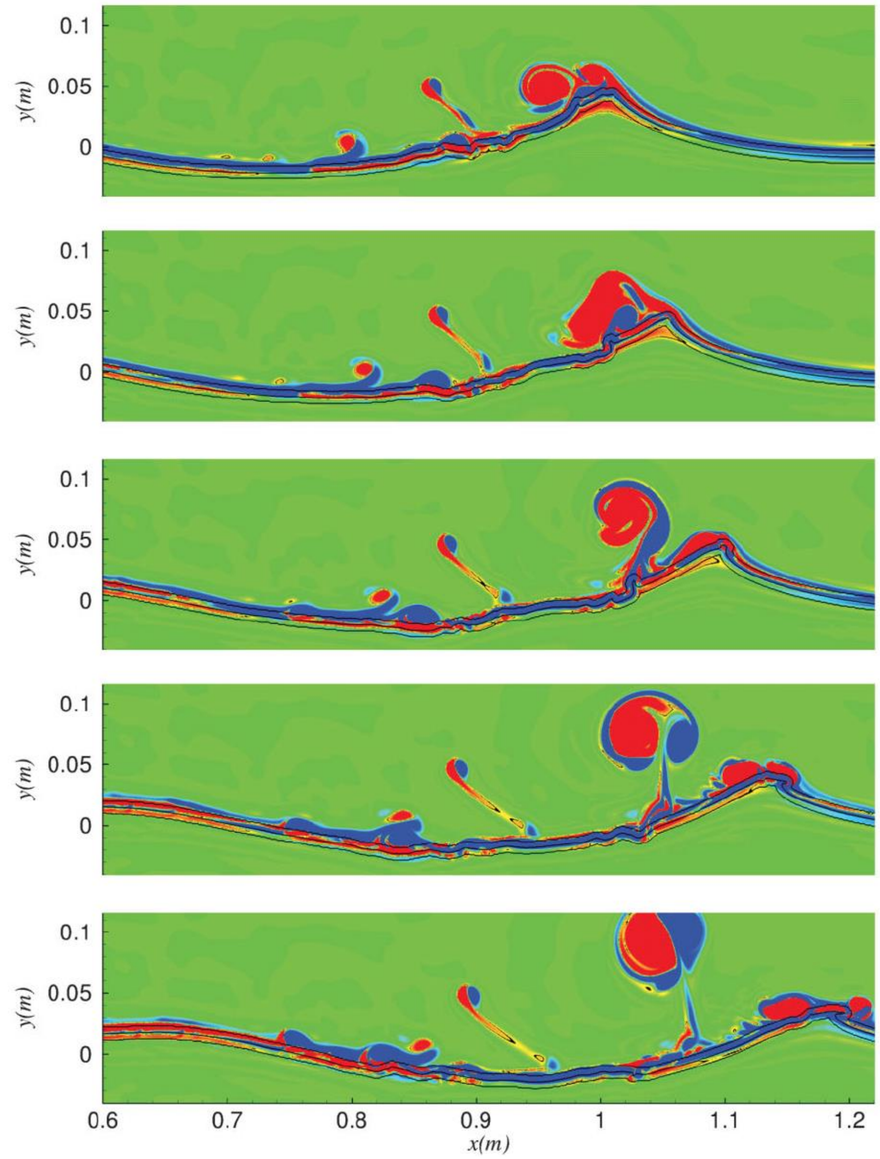
Model of generation of turbulence by nonlinear waves

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



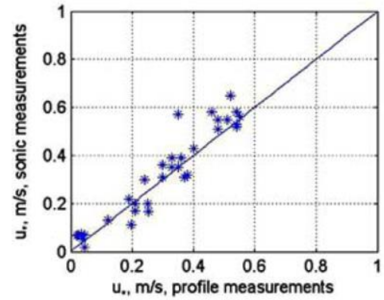


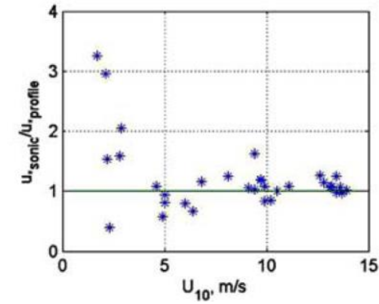
Turbulence in the air

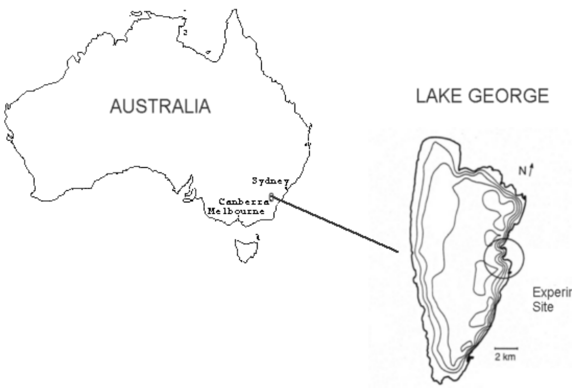


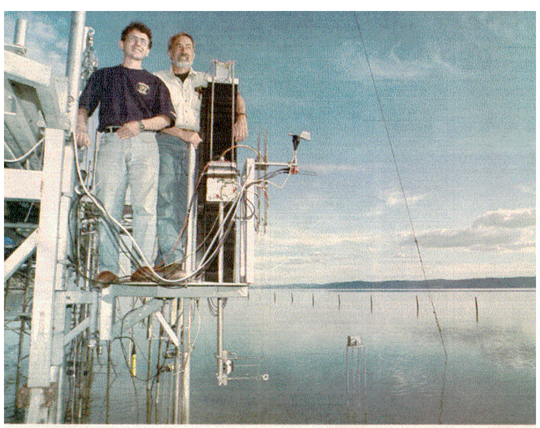
Iafrati, Babanin, Onorato, 2013, PRL

Lake George experiment









Donelan, Babanin, Young, Banner, McCormic, 2005, JTec

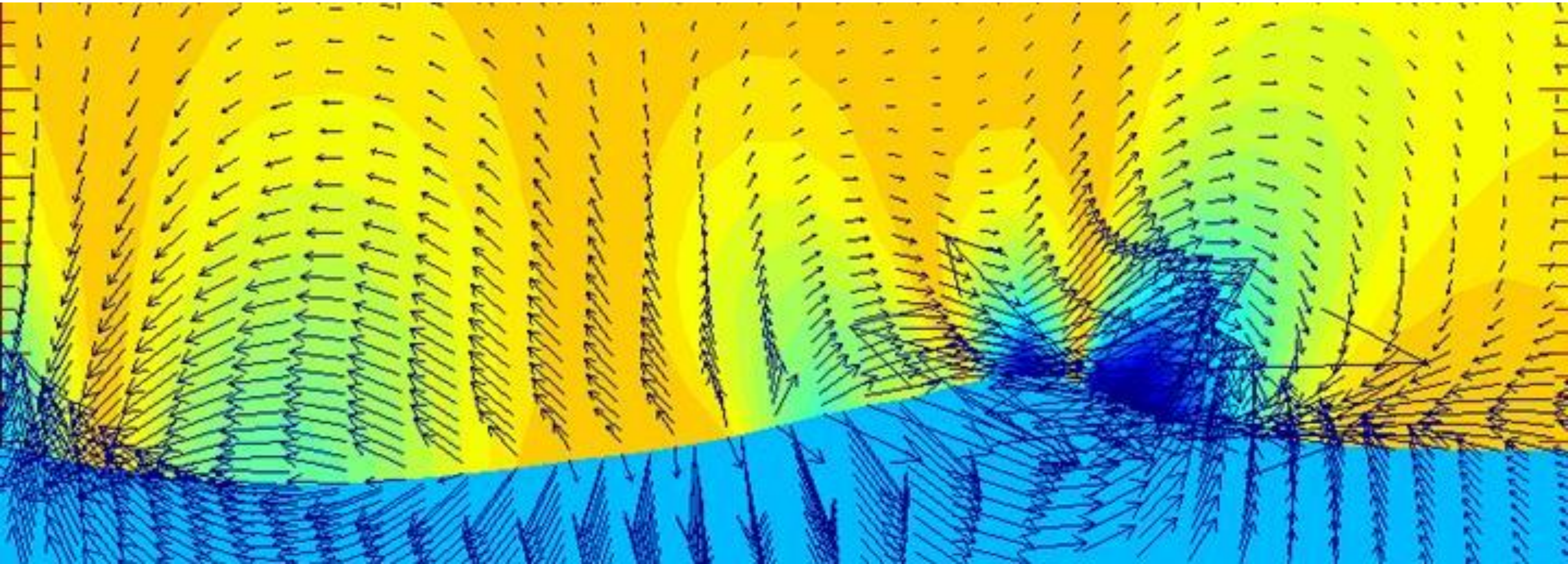
Example of the 2-D flow structure above waves

A small fragment of the reproduced field is represented

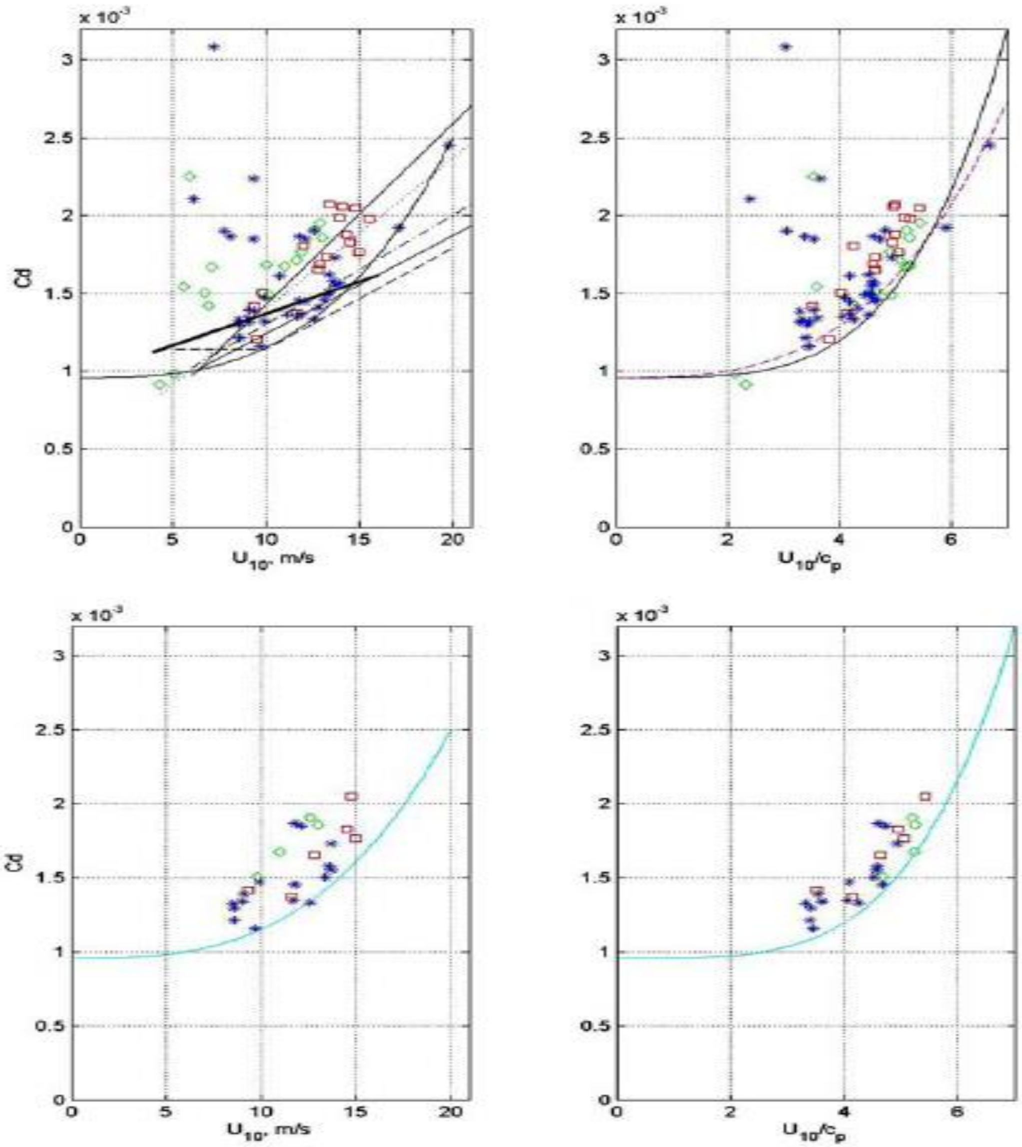
**The colours in the air show pressure distribution
(brown colour corresponds to positive values,
blue colour – to negative values);
arrows are vectors of the wave-produced velocity.**

Chalikov & Rainchik, 2011, BLM

Wave Boundary Layer Model



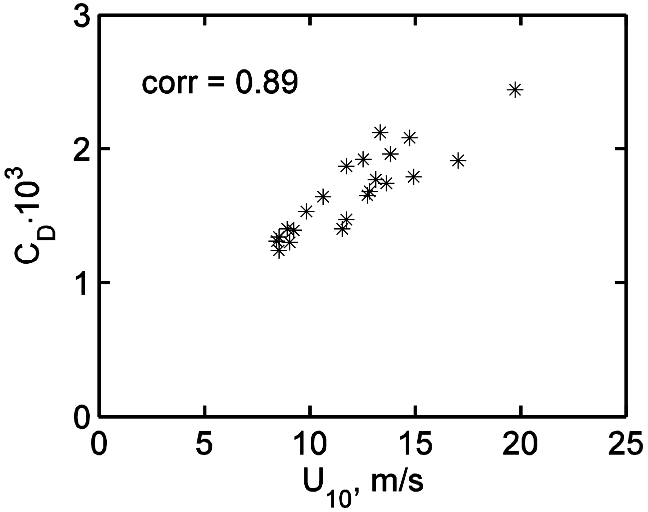
Gustiness

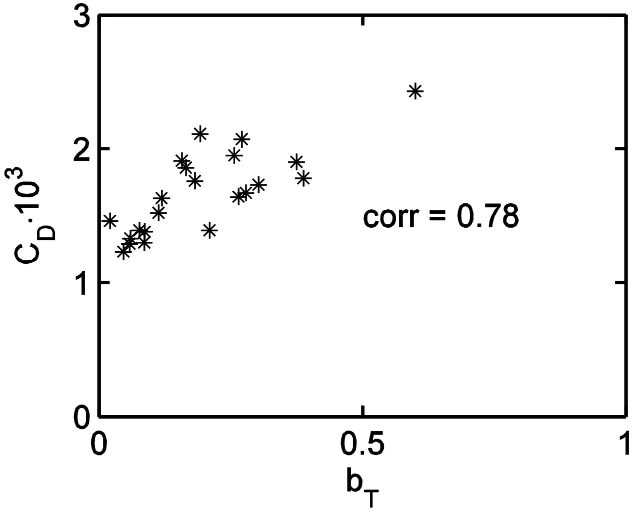


Breaking and Directional Spreading

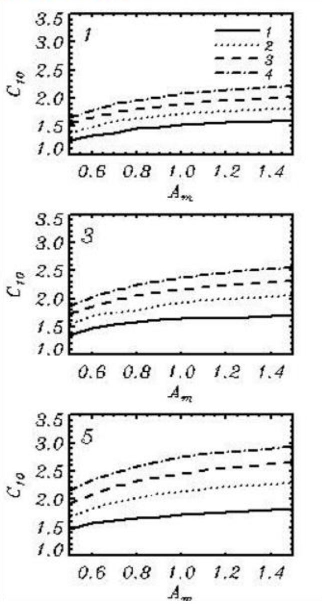
Dependence on breaking

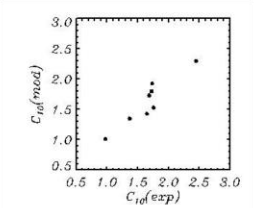
Babanin, 2011, CUP





4. Result





*Comparison of drag coefficient C_{10} (mod) calculated CR model with experimental data on spreading with experimental data on drag coefficient C_{10} (exp).

*Dependence of drag coefficient on angular spreading, characterized by parameter A_m. *Inverse wave age U_{10}/C_p is indicated in top left corner.*
1- U_{10}=10 m/s; 2- U_{10}=15 m/s;
3- U_{10}=20 m/s; 4- U_{10}=25 m/s

Dependence on directional spreading (*Chalikov & Rainchik (2011*) model)

Ting, Babanin, Chalikov, Hsu, 2012, JGR

$$K(U/c_p,f,\mu)=A(U/c_p,f)\cos^{2s(U/c_p,f)}(\mu-\mu_0/2)$$

$$U/c_p= 1, 3, 5$$

Waves and air-sea interactions, sea drag

JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 113, C02015, doi:10.1029/2007JC004233, 2008

Effects of wind trend and gustiness on the sea drag: Lake George study

Alexander V. Babanin[1] and Vladimir K. Makin[2]

Babanin & Makin, 2008, JGR

[10] We believe that a complete list of physical properties and phenomena, whose effect on the sea drag should be investigated and incorporated in the final parameterization to reduce the scatter, includes, among possible others, 1) mean wind speed; 2) sea state dependence; 3) wave steepness; 4) full flow separation for strongly forced wind waves; 5) enhancement of sea drag due to wave breaking; 6) rising and falling winds; 7) gustiness of the wind; 8) temperature stratification in the atmospheric boundary layer; 9) swell; 10) non-linear wind-wave interactions; 11) wave horizontal skewness and vertical asymmetry; 12) variation of the wavy surface properties at wave group and wavelength scales; 13) wave directionality; 14) wave short-crestedness; 15) coupled effects in the air/sea boundary layers. The 16th and separate item would be that due to peculiarities of air-sea interaction at extreme wind-forcing conditions which include an entire set of new features irrelevant at moderate winds as mentioned above. In this list, we do not mention properties and processes which breach validity of the constant-flux-layer approximation, as in such circumstances the notion of the drag coefficient (1) becomes

swells

Babanin, 2011, CUP

Swell attenuation

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

Dissipation

- 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

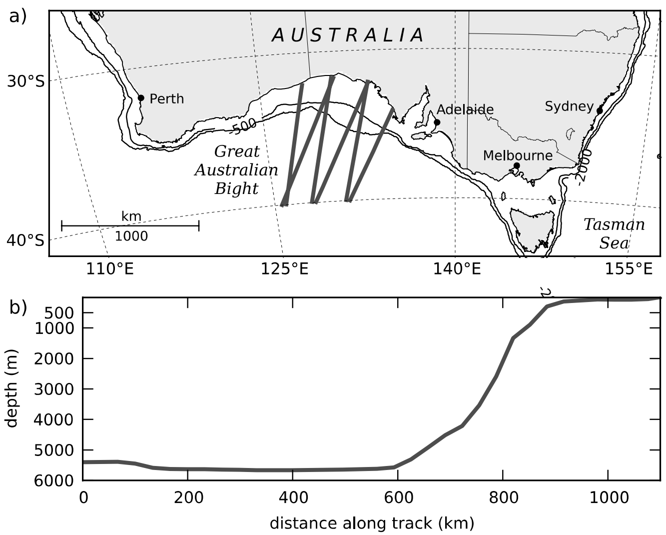
$$D_x = \frac{1}{c_g} D_a = \frac{b_1}{3} 2 \frac{k}{\omega} u_0^3 = \frac{2}{3} b_1 k \omega^2 a_0^3 = \frac{2}{3} b_1 g k^2 a_0^3.$$

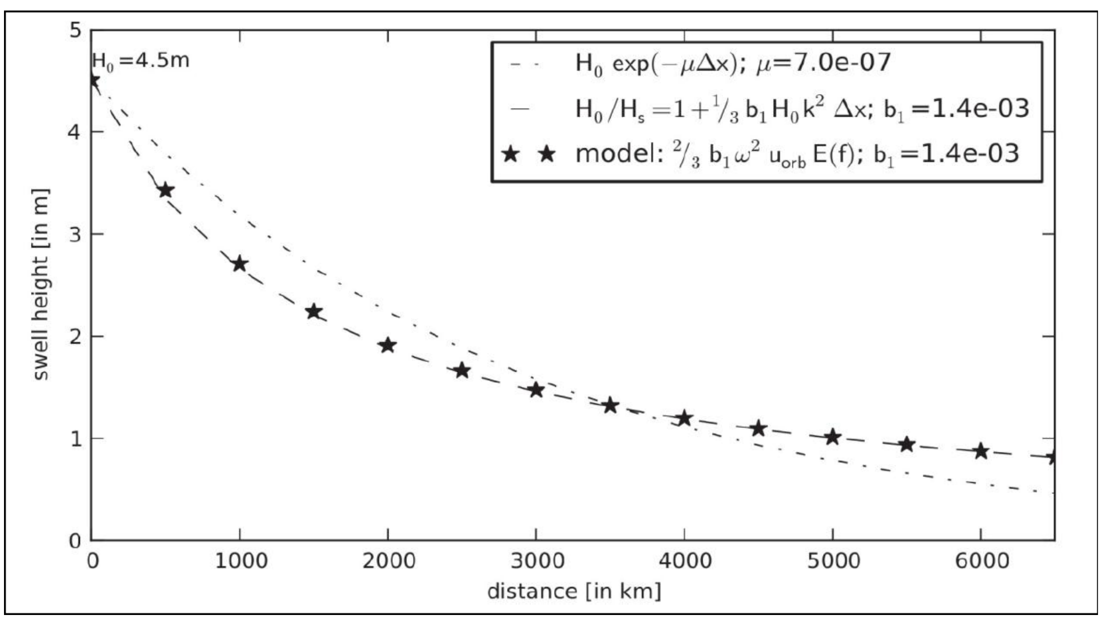
$$\frac{g}{2} \frac{\partial (a_0(x)^2)}{\partial x} = \frac{2}{3} b_1 g k^2 a_0(x)^3,$$

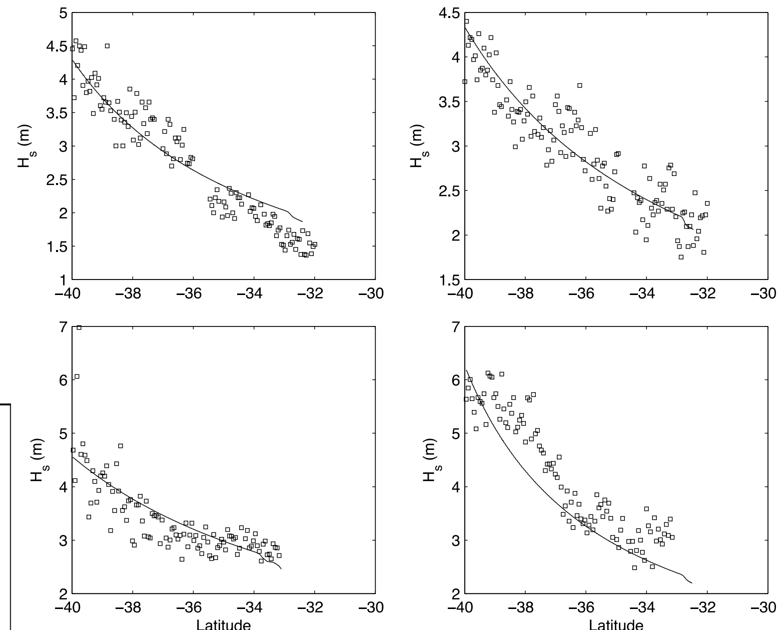
$$a_0(x)^2 = \frac{4}{B^2} x^{-2} = \frac{9}{4 \cdot b_1^2 k^4} x^{-2} = \frac{9}{64} 10^6 k^{-4} x^{-2}.$$

Young, Babanin, Zieger, 2013, JPO

Swell attenuation







A Global Climatology of Wind–Wave Interaction

KIRSTY E. HANLEY AND STEPHEN E. BELCHER

Department of Meteorology, University of Reading, Reading, United Kingdom

PETER P. SULLIVAN

National Center for Atmospheric Research, Boulder, Colorado

2010, JPO

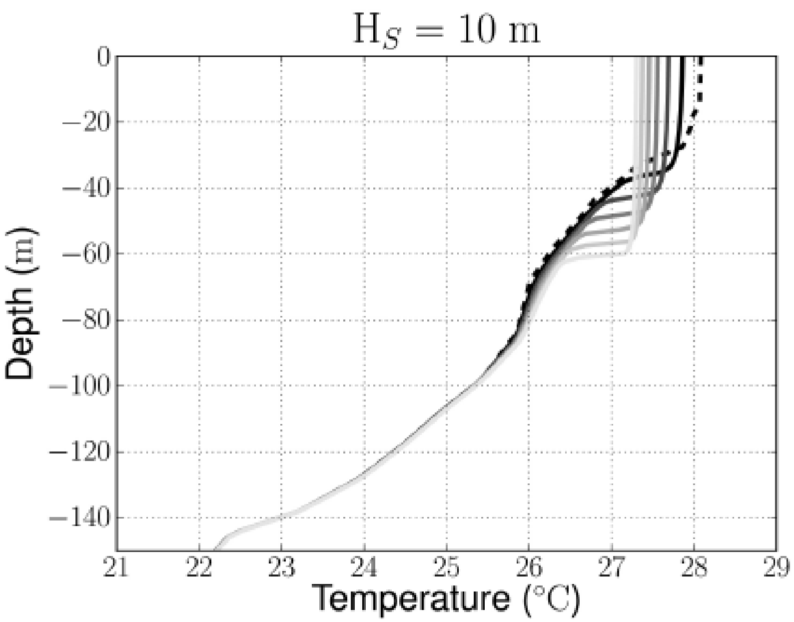
(Manuscript received 1 October 2009, in final form 6 January 2010)

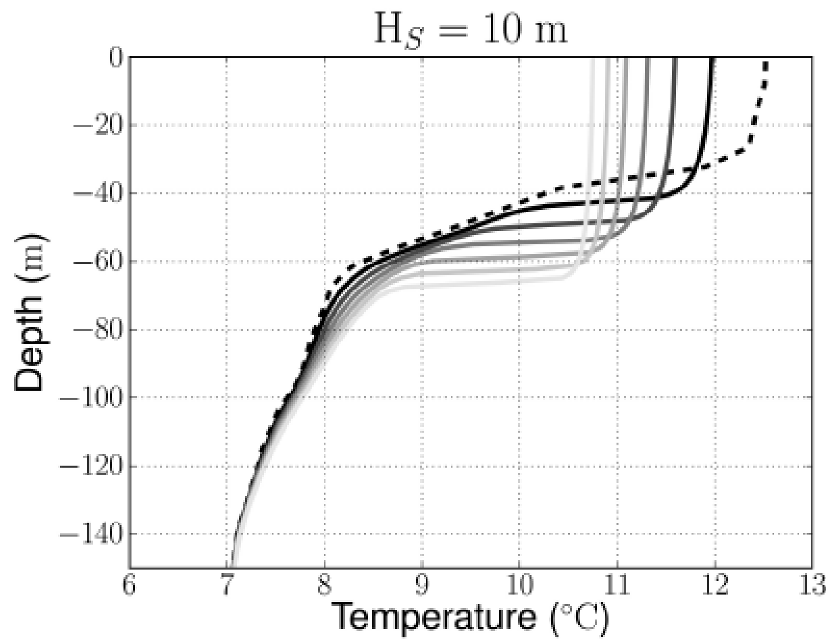
ABSTRACT

Generally, ocean waves are thought to act as a drag on the surface wind so that momentum is transferred downward, from the atmosphere into the waves. Recent observations have suggested that when long wavelength waves—which are characteristic of remotely generated swell—propagate faster than the surface wind, momentum can also be transferred upward. This upward momentum transfer acts to accelerate the near-surface wind, resulting in a low-level wave-driven wind jet. Previous studies have suggested that the sign reversal of the momentum flux is well predicted by the inverse wave age, the ratio of the surface wind speed to the speed of the waves at the peak of the spectrum. Data from the 40-yr ECMWF Re-Analysis (ERA-40) have been used here to calculate the global distribution of the inverse wave age to determine whether there are regions of the ocean that are usually in the wind-driven wave regime and others that are generally in the wave-driven wind regime. The wind-driven wave regime is found to occur most often in the midlatitude storm tracks where wind speeds are generally high. The wave-driven wind regime is found to be prevalent in the tropics where wind speeds are generally light and swell can propagate from storms at higher latitudes. The inverse wave age is also a useful indicator of the degree of coupling between the local wind and wave fields. The climatologies presented emphasize the nonequilibrium that exists between the local wind and wave fields and

Ocean, Weather and Climate

Modelling SST and MLD at the scale of hours

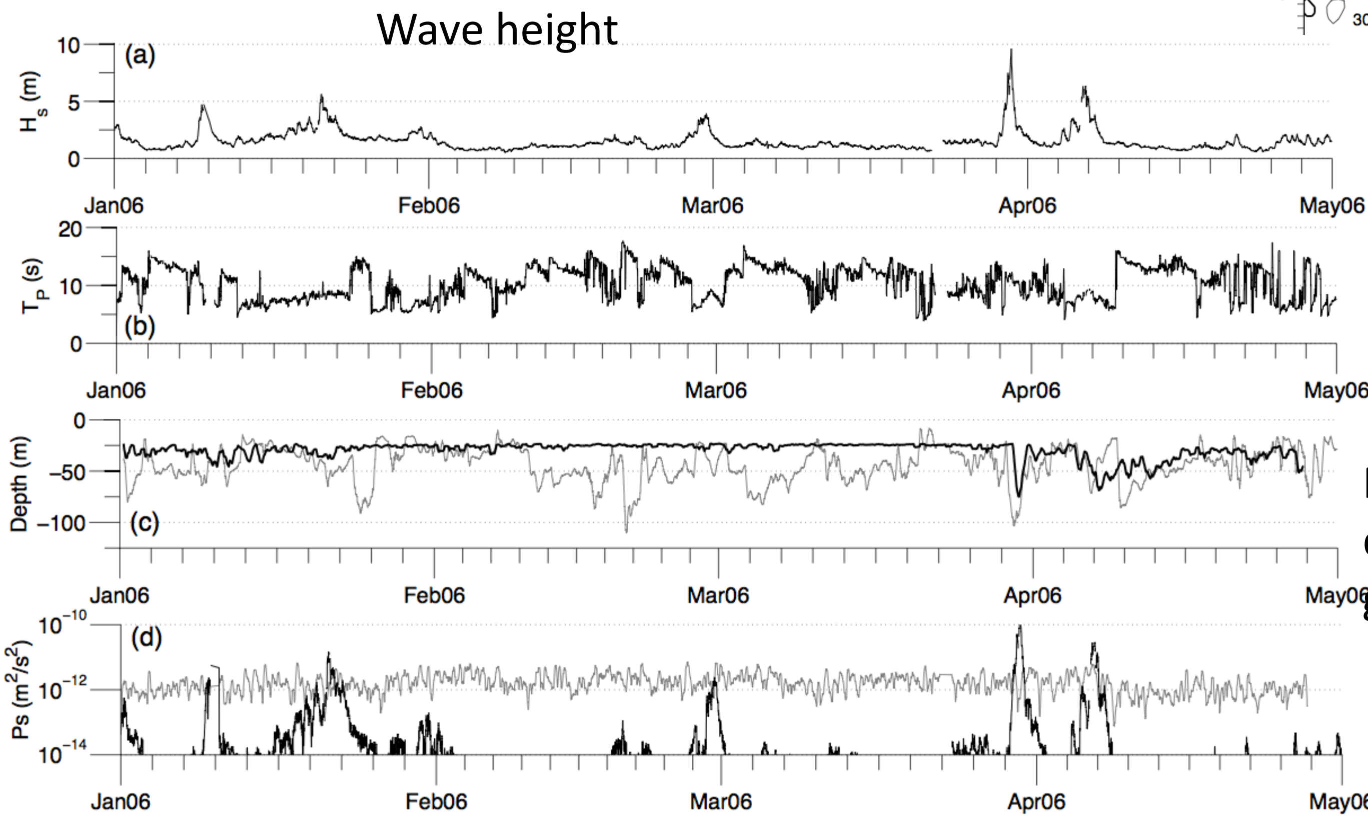


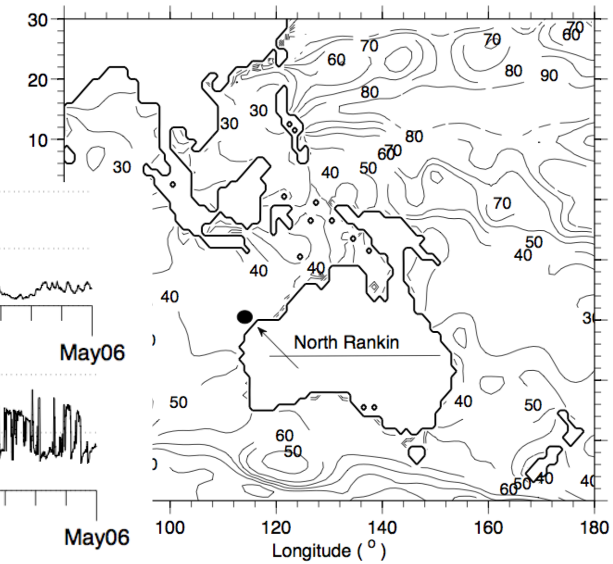


Ghantous & Babanin, 2014, NPG

Toffoli, McConochie, Ghantous, Loffredo, Babanin, 2012, JGR

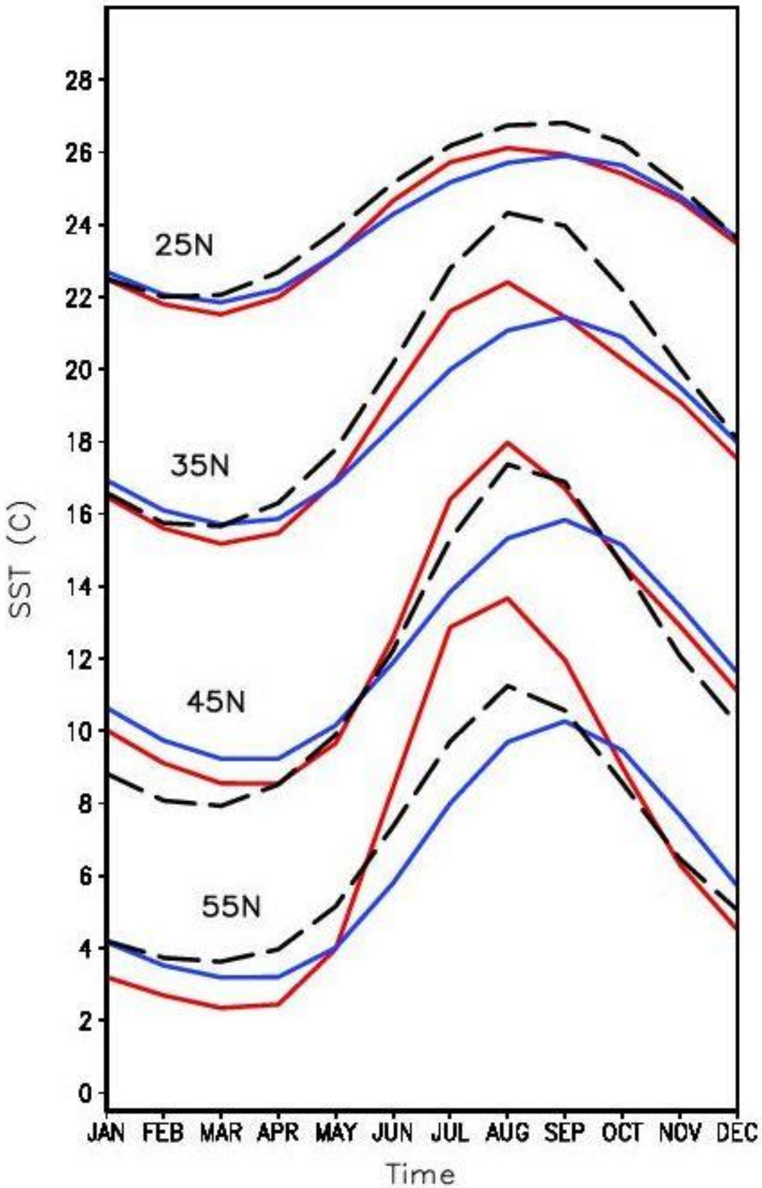
Field observations, North Rankin mixed layer deepening





MLD:
dark – wave induced
grey - measured

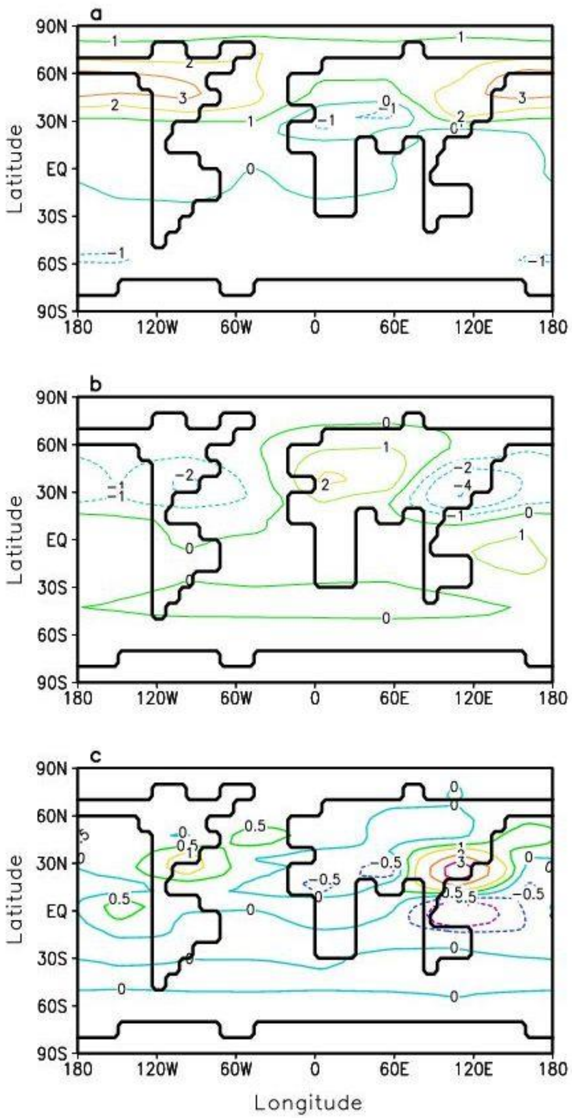
Implementing wave-induced mixing in CLIMBER



- 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

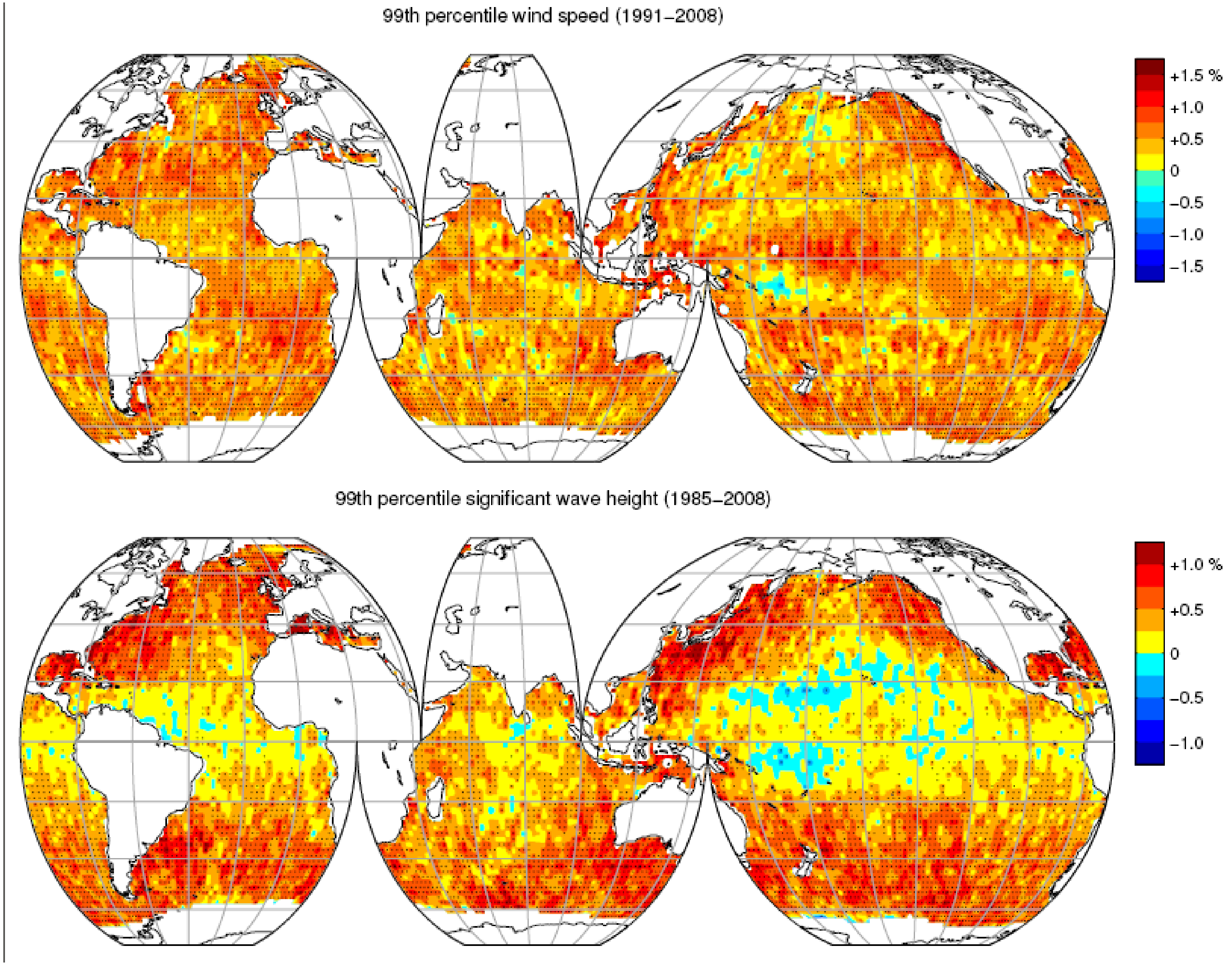


Global distribution (Northern summer)

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

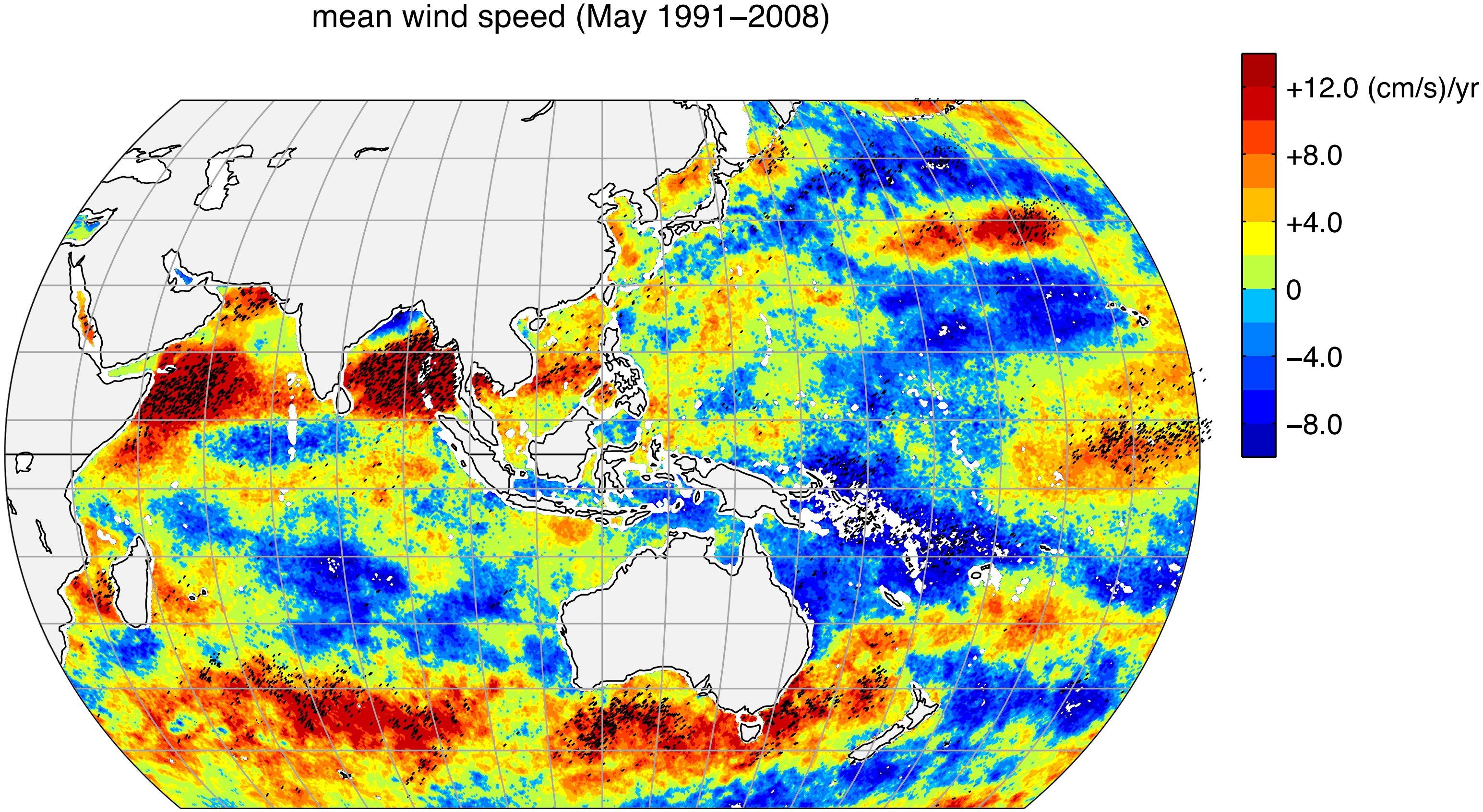
Wave and Wind Climate

Wind and waves as climate indicators



Young, Zieger, Babanin, 2011, Science

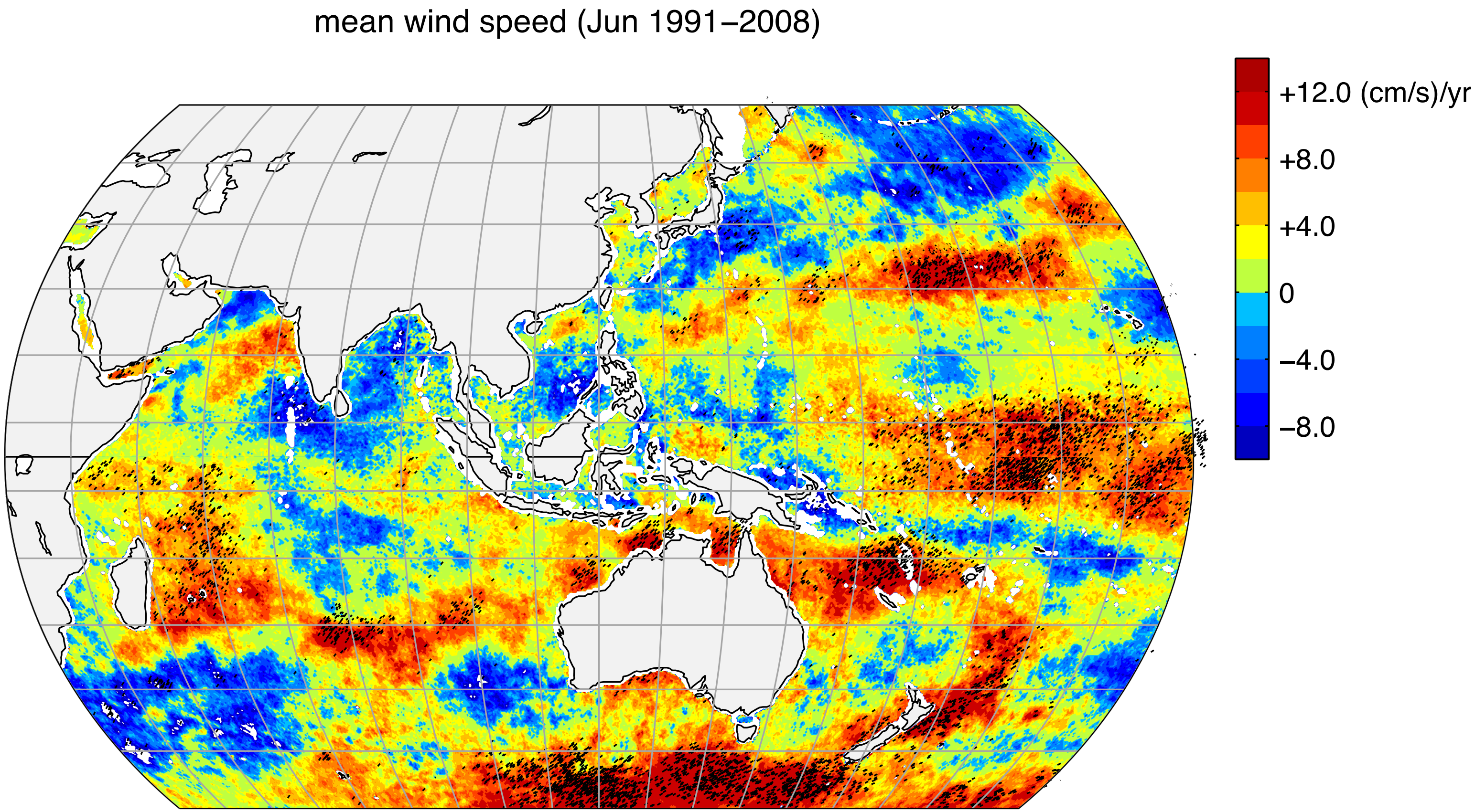
Wind Trends, by SSM/I



Trend analysis (MK test) applied to monthly mean SSM/I (F10,F11,F13) wind and precipitation from 1991 to 2008. Hatching indicates significant changes (normcdf test [95% level]) and contour interval is 2.00 cm s^{-1} per year.

Zieger, Babanin, Young, 2014, DSR P1

Wind Trends, by SSM/I



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Zieger et al., Deep Sea Res., 2014

Conclusions

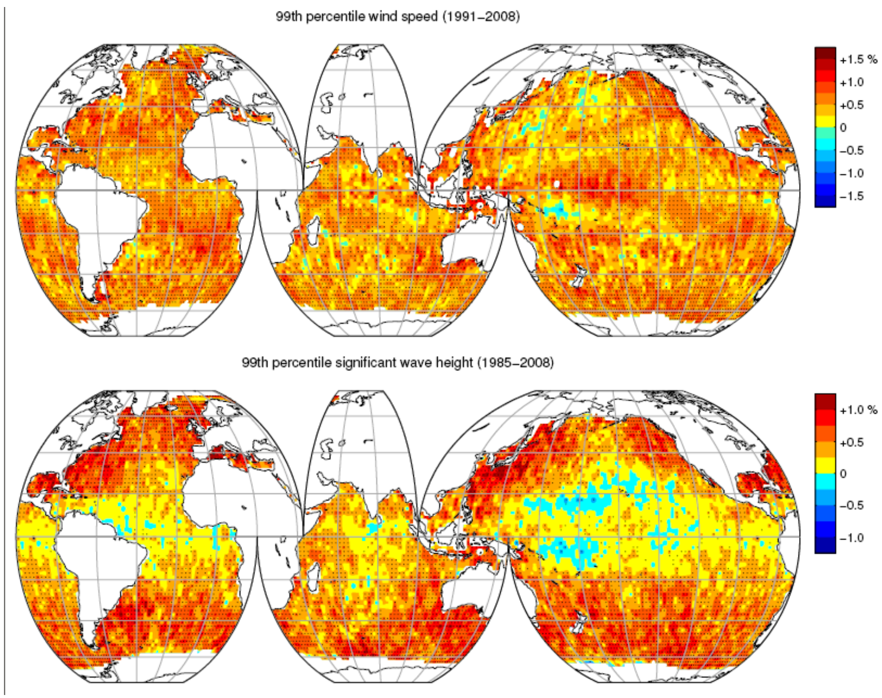
- coupling of small-scale models (waves, turbulence) with large-scale models (weather, climate) is necessary
  - physics is continuous
  - computing capabilities allow the coupling
- waves provide feedback
  - to the atmospheric boundary layer
  - to the upper ocean (essentially overlooked)
  - to the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

Motivation

Waves influences the climate, climate affects the waves

Winds and waves change

Observations



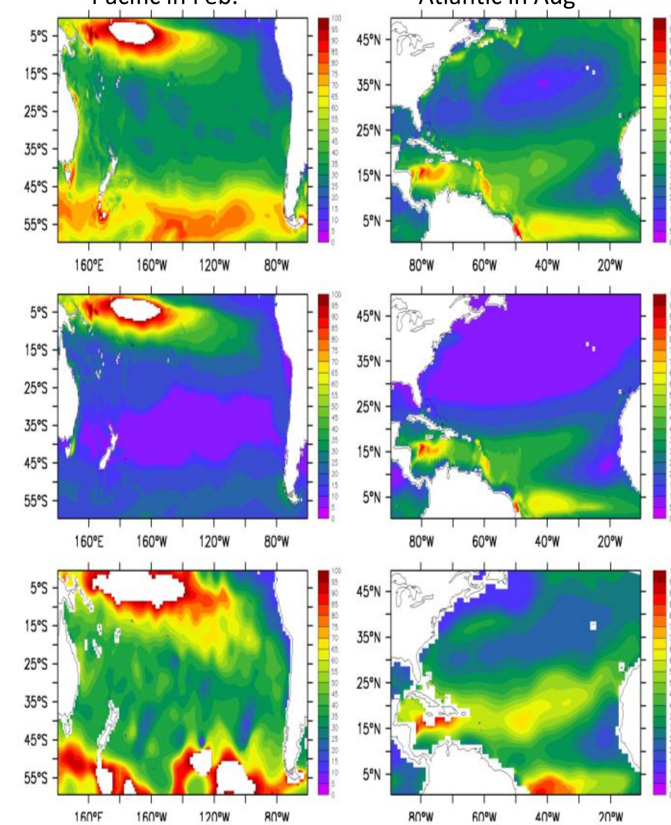
Young et al., Science, 2011

Sea surface temperature changes due to waves

Model

MLD , Southern Pacific in Feb.

MLD, Northern Atlantic in Aug



With waves

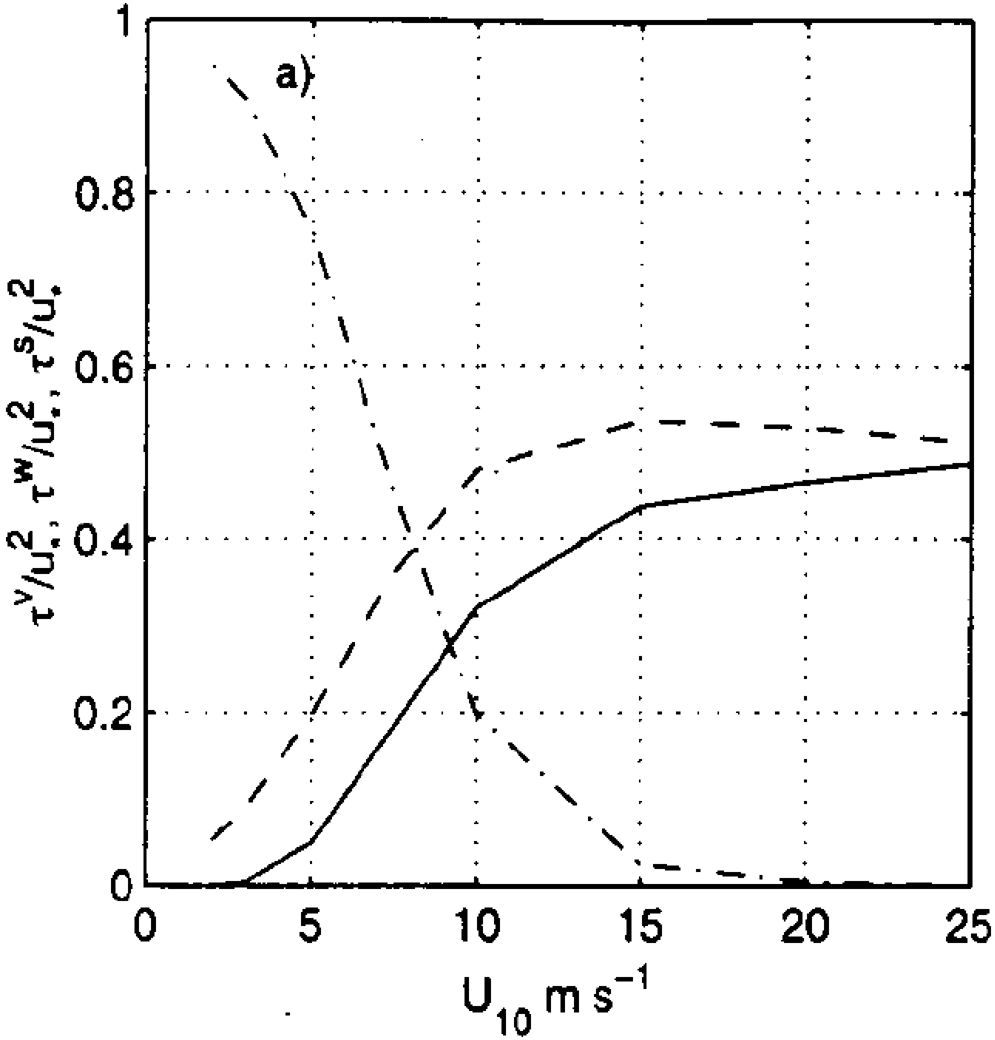
Without waves

World Ocean Atlas

Qiao et al., Ocean Dynamics, 2010

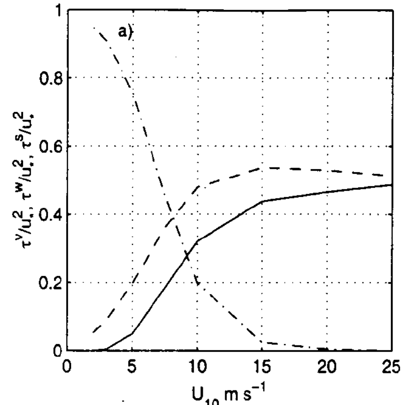
Motivation

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



Kudryavtsev-Makin, 2011, BLM

Waves and ocean turbulence



Kudryavtsev & Makin, 2011, BLM

- 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
- 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 ciruclation
- 2-3m of the ocean water have the same heat capacity as the entire atmosphere

Kinsman, 1965: Wind Waves

based on Phillips (1961)

Navier-Stokes equation

linearised boundary conditions, with surface tension *T*

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

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

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

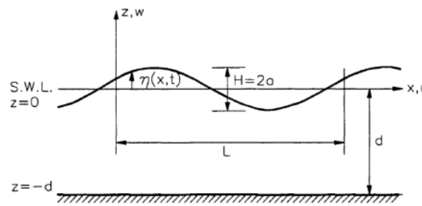
$$\frac{\partial \eta}{\partial t} = w_{z=0}$$

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

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

Solutions

vorticity



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

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

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

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

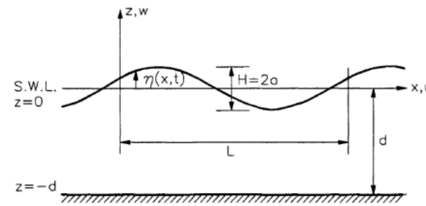
- exponential decay in z and t
- oscillations in x, z and t
- ‘length’ of vertical vorticity oscillation is much smaller than λ

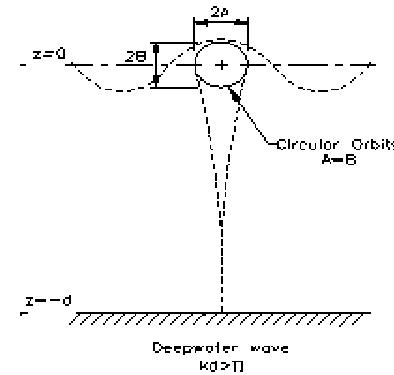
Hypothesis of the Wave Reynolds Number

Babanin, GRL, 2006

η(x,t) = $a_0$cos(ωt+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=ωa 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

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

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

Re_{cr}=3000