Estimated contribution of wind-waves in the coupled climate system

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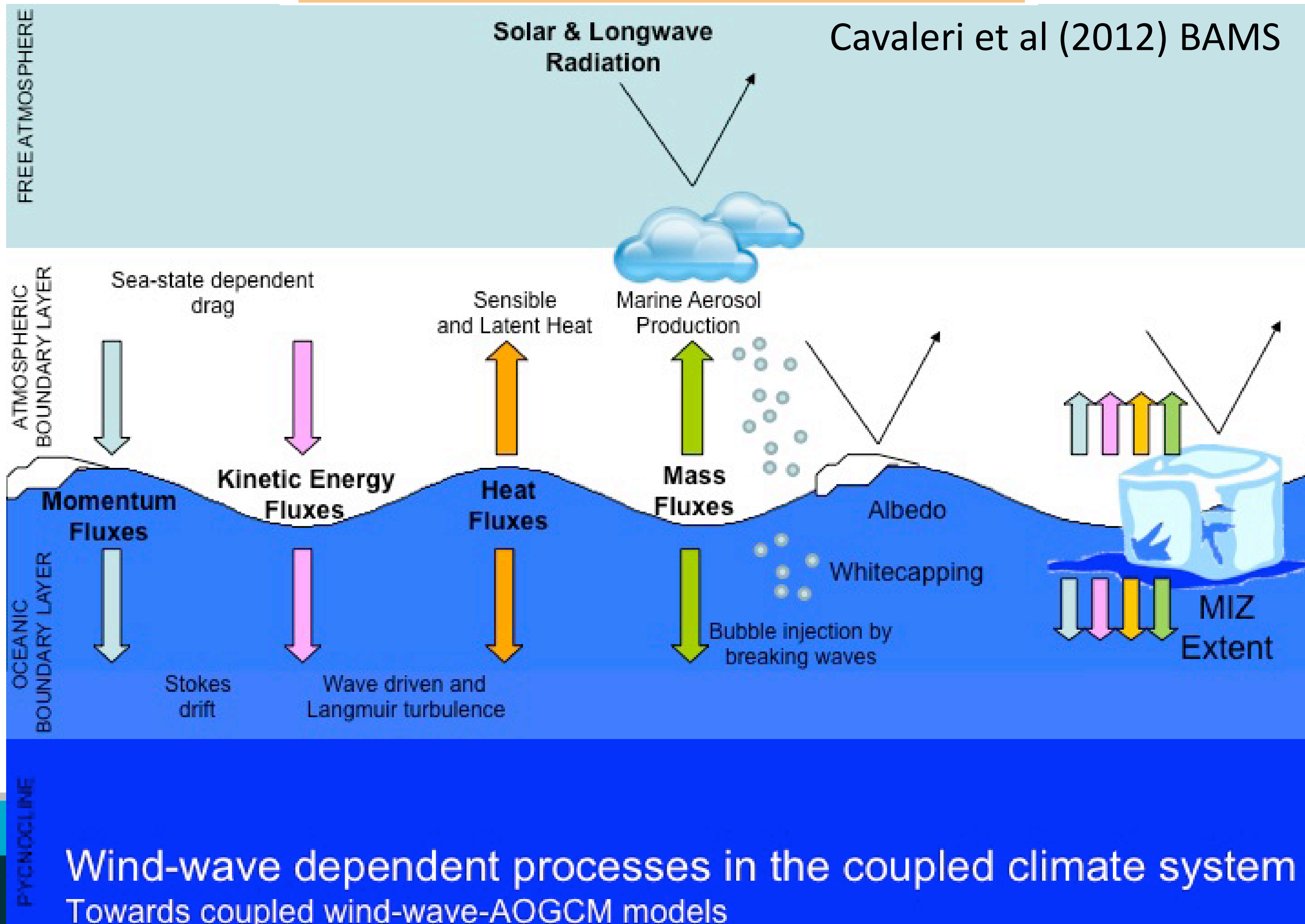
WIND WAVES IN THE COUPLED CLIMATE SYSTEM

BY L. CAVALERI, B. FOX-KEMPER, AND M. HEMER

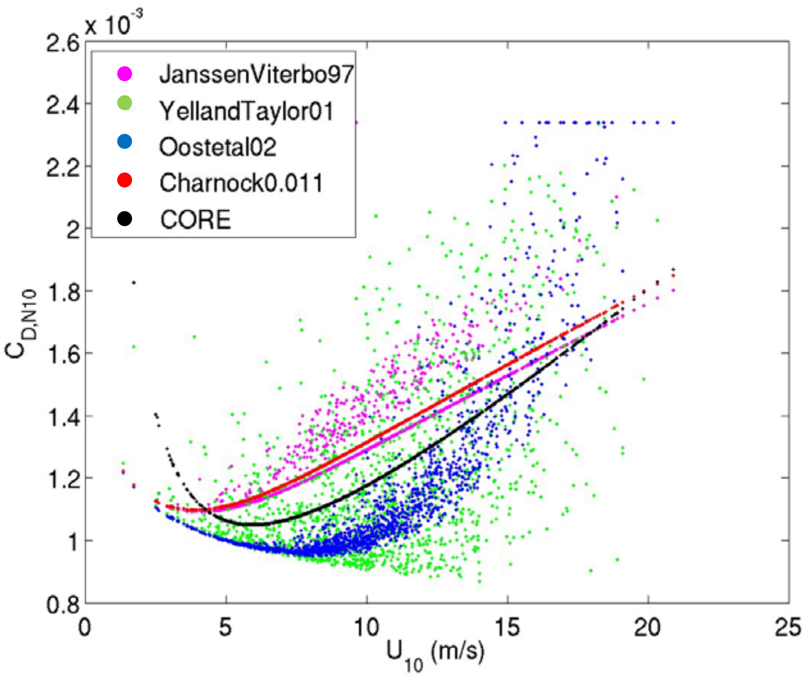
Gravity wind-wave–driven processes at the ocean surface—including radiation fluxes and energy, mass, and momentum exchanges—play an important role in the coupled climate system.



Methodology

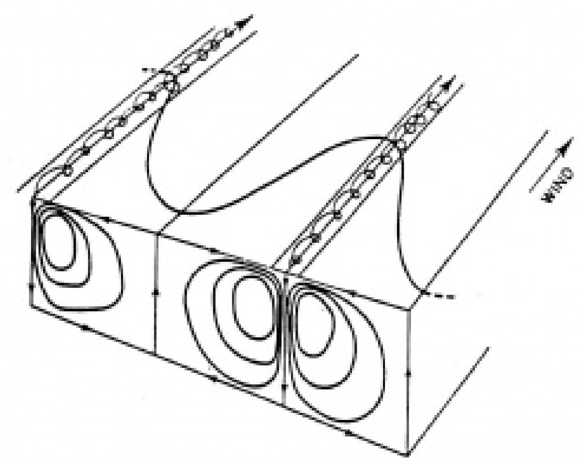
Atmosphere

Assess sensitivity of CORE 1 Large & Yeager ocean-atmosphere surface fluxes to wave dependent parameterisations of sfc roughness.



Ocean

Annual cycle integration of 1-d mixing models with parameterisation of langmuir mixing, applied globally. Assess sensitivity to inclusion of wave driven mixing.



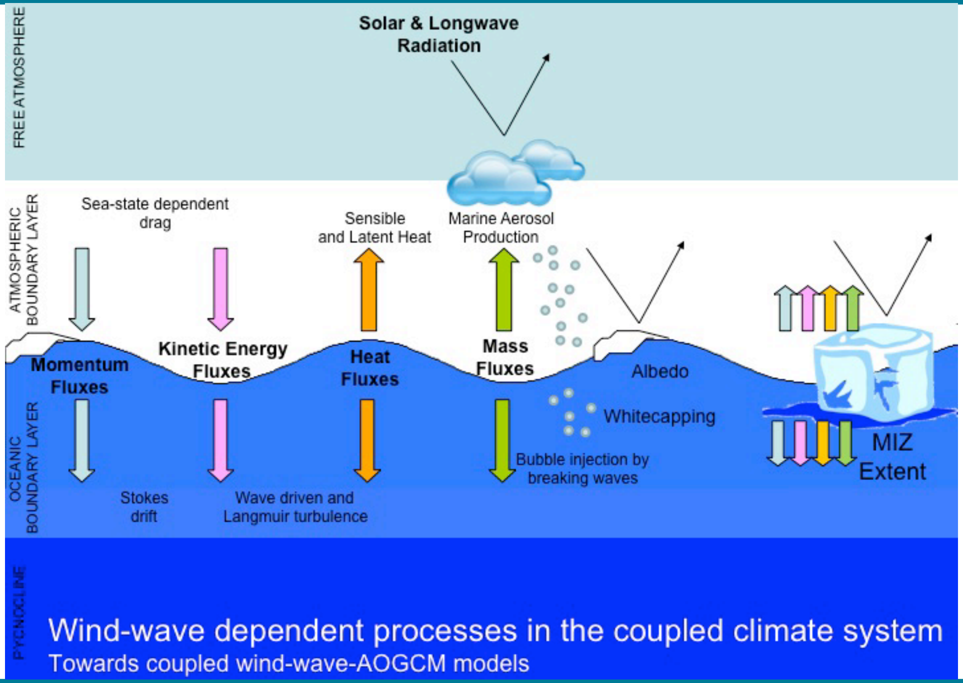
Conclusions

Atmosphere

Wave dependent parameteris^{N} of roughness leads to up to 1PW of additional heat transfer to the ocean (within range of alternate wind-dependent schemes.

Ocean

Langmuir mixing of the surface ocean shows ~25% increase in MLD in extra-tropical storm belts during winter. Equivalent to an additional heat uptake of ~1 PW to the global ocean over one year.



The Future

Many other processes to be considered

This 'back of the envelope' approach used to support decisions as to where to focus effort in a coupled model (Elodie's talk)

c.f. 1.4PW is the estimated heat flux transported by the Gulf Stream

The Large and Yeager CORE forcing

•Annual mean river runoff
•Monthly varying precipitation (12 time steps per year),•Daily varying shortwave and longwave radiative fluxes (365 time steps per year, and so no diurnal cycle and no leap years)
•Six-hourly varying meteorological fields (1948-2006)10m air temperature, humidity, zonal/ meridional winds, SLP

Large and Yeager (2004, 2009) Bulk Formula

Surface boundary condition determines fluxes of heat, freshwater, and momentum
Solved separately for ocean and sea-ice covered areas of each grid cell
Net Heat Flux = Sensible + Latent + Shortwave + Longwave
Net freshwater flux = Precipitation -Evaporation + River Runoff (+ Glacial Calving)
Net momentum exchange is driven by windstress, accounting for ocean-ice stress, and ocean currents

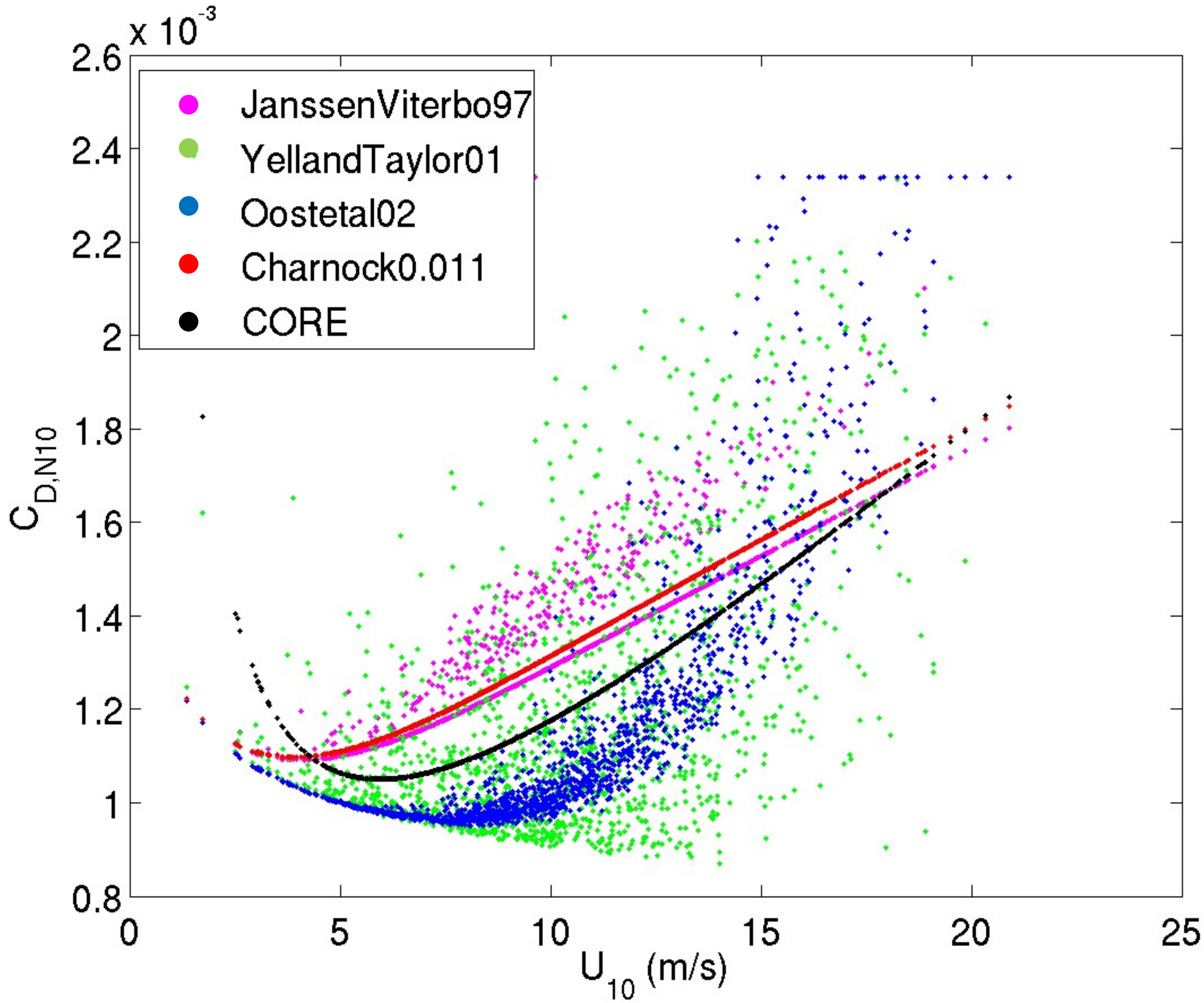
Large and Yeager Solution set (mean monthly 1948-2006)

Solves bulk formula under CORE forcing with Hadley OI-SST
http://dss.ucar.edu/datasets/ds260.2/

Investigate sensitivity of air-sea fluxes to wave dependent parameterisations of roughness.

- Run WaveWatch III forced with CORE normal year winds.
  - 13 month wave model run (WaveWatch III, 1 degree resolution, nf=36, nd=24) CORE Normal Yr forcing.
  - 1 month (dec) spinup + 1 whole CORE normal yr.
  - Full directional spectra archived at 4deg resolution.

Drag Coefficient vs Wind Speed (SOFS, 47S, 142E)



$$C_{DN10} = \left(\frac{2.7}{u_{10}} + 0.142 + 0.0764u_{10}\right) \Big/ 1000$$

$$z_o = \alpha\frac{u_*^2}{g} + \left(\frac{0.11\nu}{u_*}\right)$$

$$z_o = \frac{50}{2\pi}\lambda_p\left(u_*/C_p\right)^{4.5} + \left(\frac{0.11\nu}{u_*}\right)$$

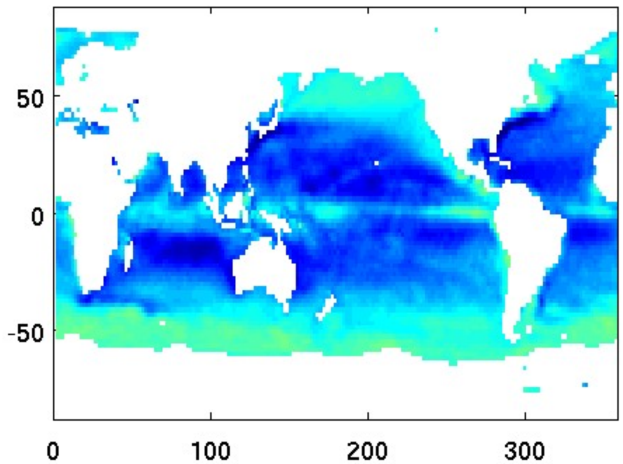
$$z_o = 1200h_s\left(h_s/\lambda_p\right)^{4.5} + \left(\frac{0.11\nu}{u_*}\right)$$

$$z_o = \alpha\frac{u_*^2}{g} + \left(\frac{0.11\nu}{u_*}\right) \quad \alpha = \beta\left(1 - \frac{\tau_w}{\tau}\right)^{-1/2}$$

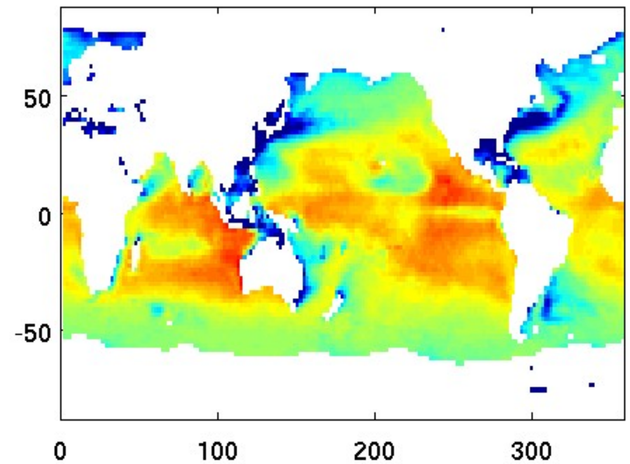
$$C_{DN10} = \frac{\kappa^2}{\log\left(\left(\frac{10}{z_o}\right)^2\right)}$$

Latent Heat Flux Mean Bias (Param - CORE)

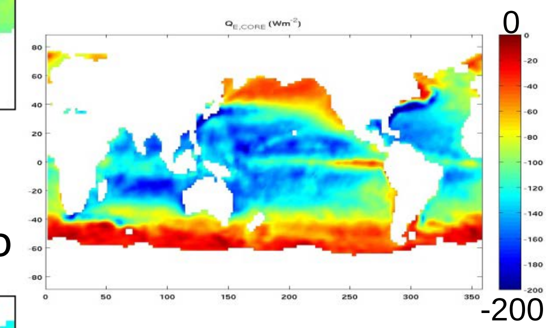
Charnock



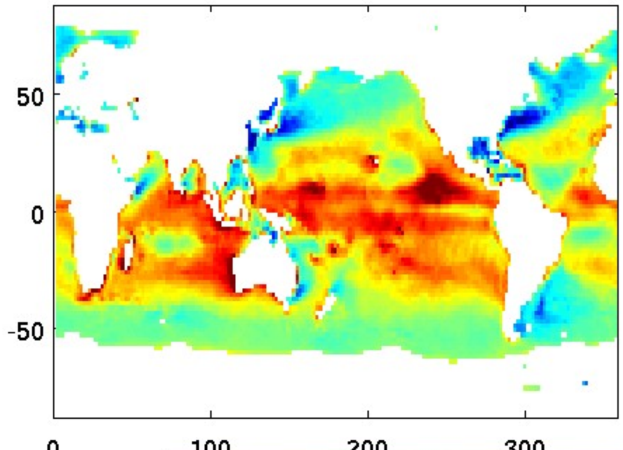
Wave Age: Oost et al.



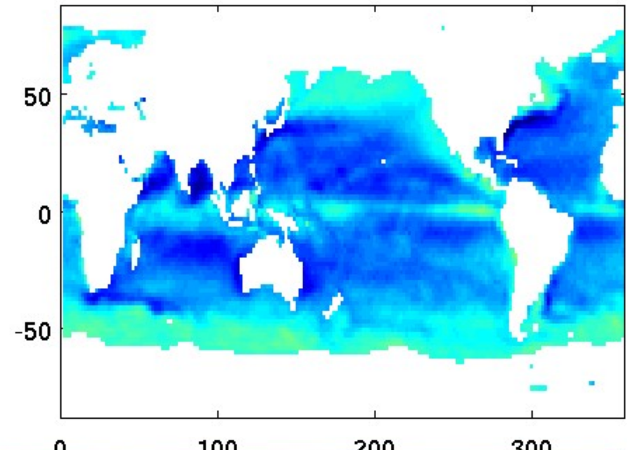
$$\overline{Q_{E,CORE}} = \overline{\Lambda_v \rho C_{E,CORE}\left(q - q_{sat}(SST)\right)|u|},$$



Wave Steepness: YellandTaylor



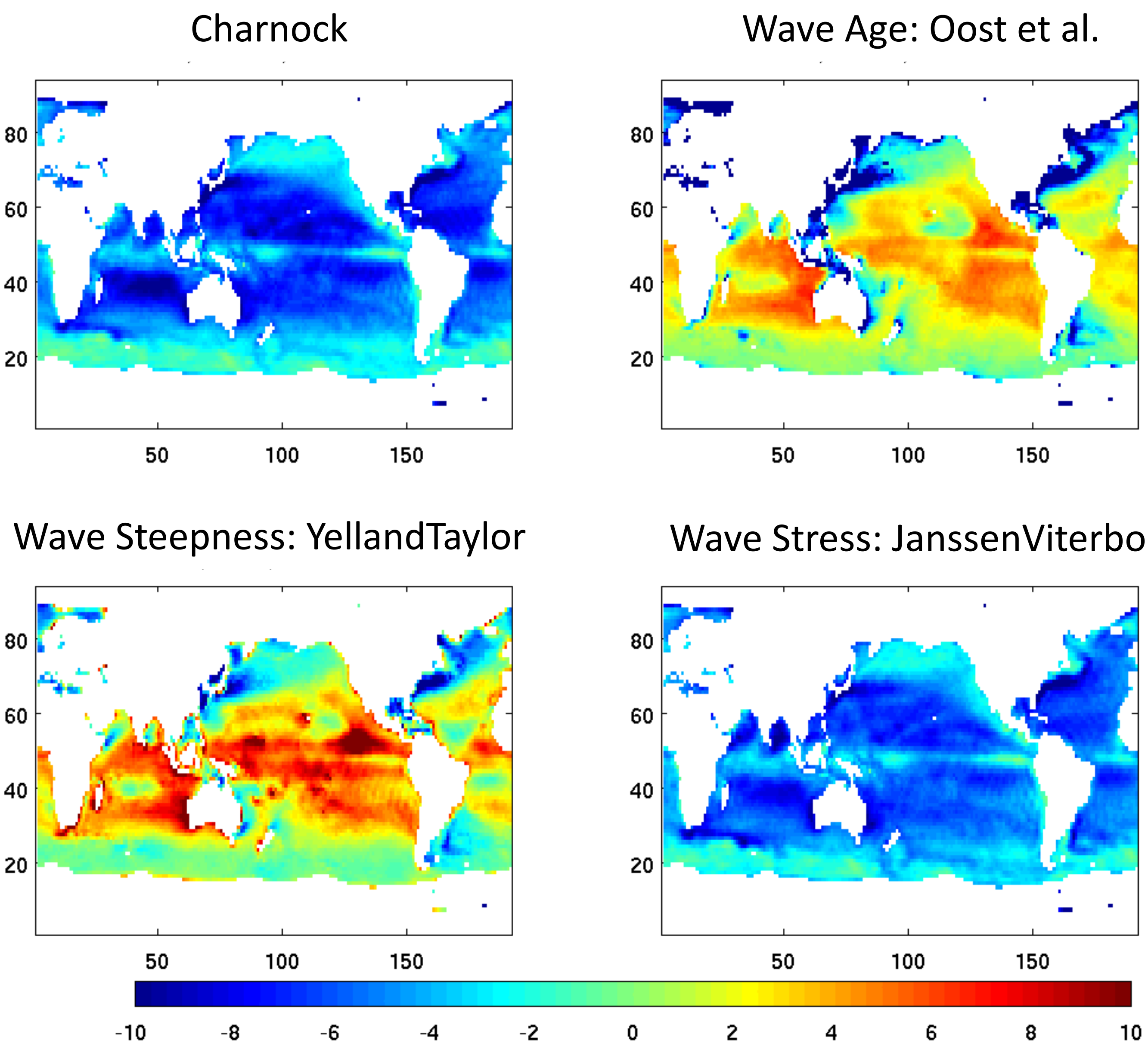
Wave Stress: JanssenViterbo



$$C_{EN} = \frac{\kappa \sqrt{C_{DN}}}{\ln\left(\frac{Z_r}{z_q}\right)}$$

Wm^{-2}

Total Heat Flux $Q_A = Q_S + Q_E + Q_L + Q_H$ (Mean Bias)



$$\overline{\left(Q_{A,param} - Q_{A,core}\right)}$$

Wm-2

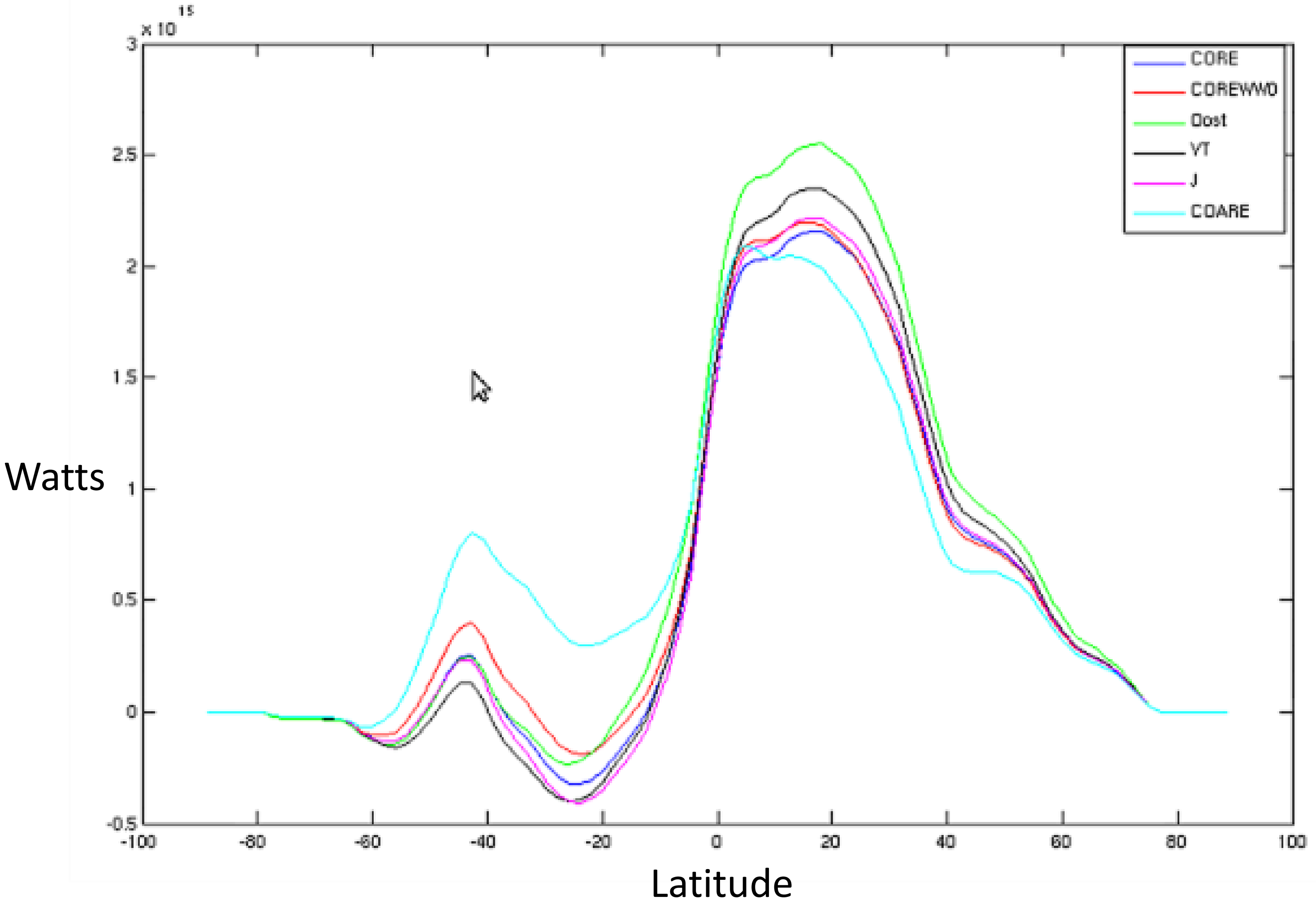
Global mean air-sea fluxes (Wm^{-2})

Corrected 2007 CORE data. C.f., Table 3 Large and Yeager, 2009, but note different masking area defined by wave model.

Flux	CORE	Charn	Oost et al	Taylor-Yelland	Janssen
Qh	-12.8	-13.4	-13.1	-12.7	-13.4
Qs*	178.4[1]	178.3[2]	178.3[3]	178.3[3]	178.3
Ql	-53.9	-53.9	-53.9	-53.9	-53.9
Qe	-107.4	-112.0	-107.0	-105.3	-111.9
Qa	4.3	-1.0	4.3	6.4	-0.8

- 1 no consideration of whitecapping.
- 2 whitecapping parameterised using wind-dependent method of Frouin et al., 2001.
- 3 whitecapping parameterisation is sea-state dependent, following Zhao et al. 2003.
    - This is a function of u*, thus dependent on zo parameterisation.
- No wave dependent long-wave radiation flux is implemented. Note surface emissivity has a sea-state dependent component
- 4 rms (spatial) of annual mean relative to CORE calculation annual mean.

Implied Northward heat transport



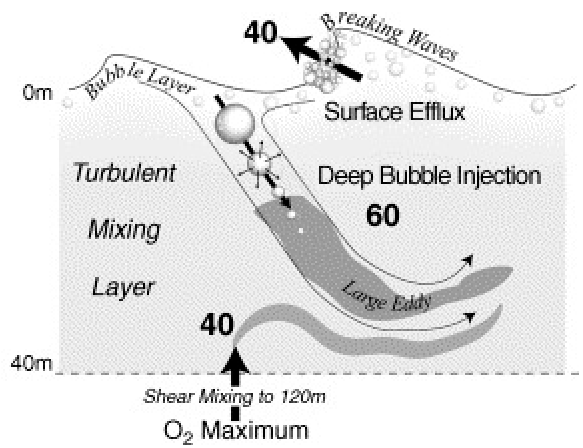
Assume bias/storage (from previous slide) is uniformly distributed across the global ocean.

Integration of this heat budget implies wave-based parameterisation can lead to an increase in heat storage of approx 1PW (Taylor and Yelland, 2001), or decreased capacity of approx 2PW, which are within the limits set by alternative wind dependent parameterisations of roughness.

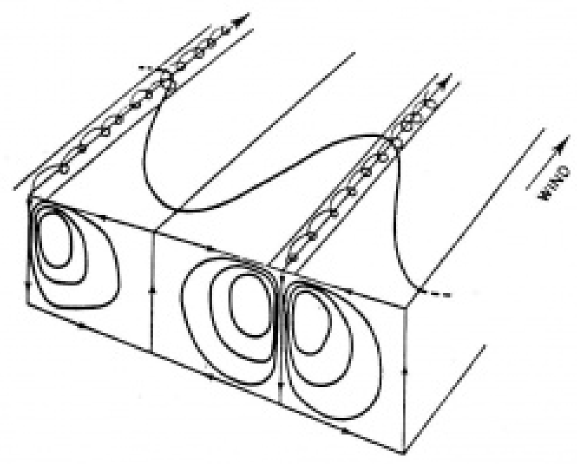
3. Wave Climate Change:
The effect of waves on climate

a. An oceanographic example

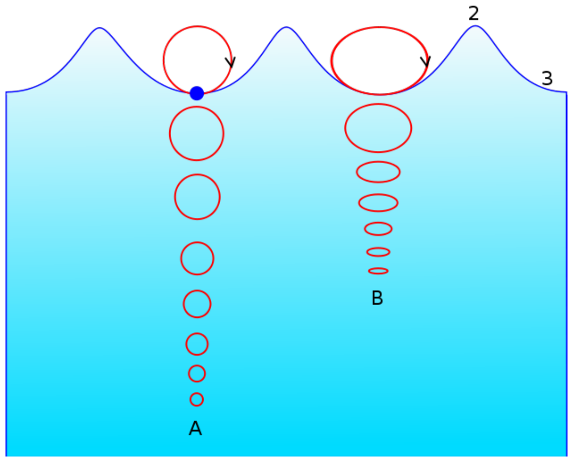
Wave driven surface ocean mixing: 3 possible processes



Injection of turbulence by breaking waves. Mixes to depth approximately equal to the wave height (Craig and Banner, 1994)



Langmuir mixing mixes to a depths of order 100m. (Langmuir, 1938)

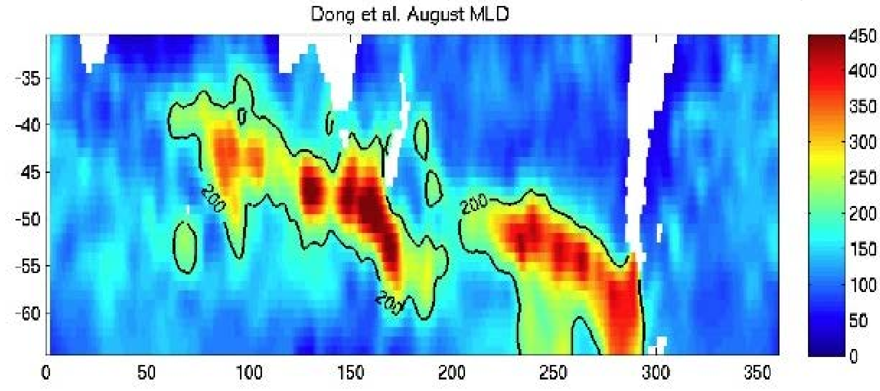


Non-breaking wave mixing. It has been proposed turbulence generated by wave orbital motion can mix to depths of order 100m.
(Babanin, 2006)

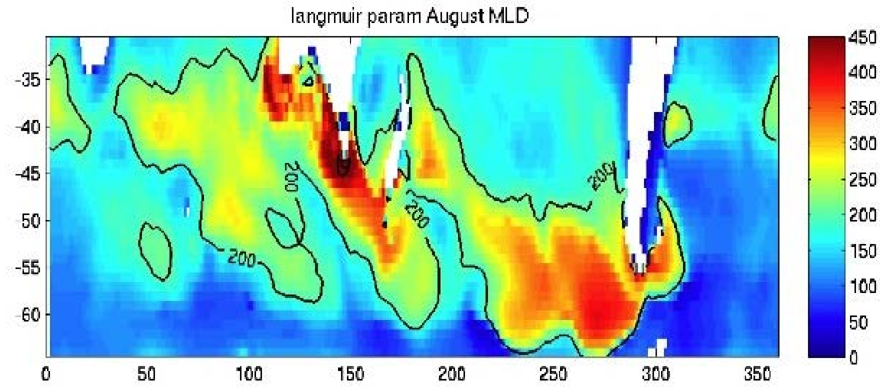
Wave dependent mixing

August Southern Ocean mixed layer bias (Webb et al., 2010)

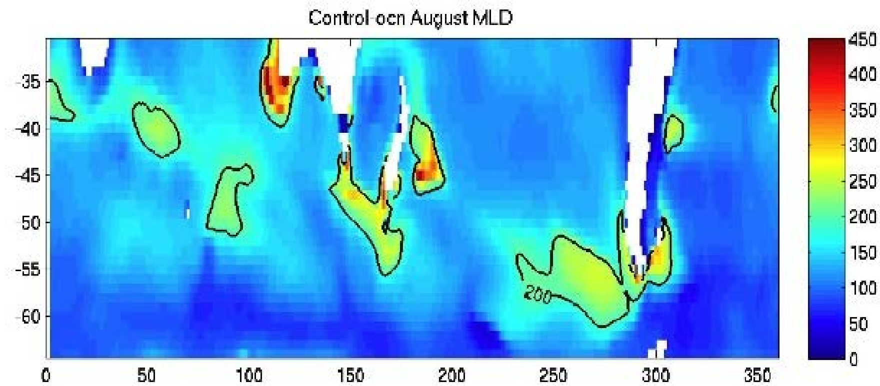
Dong et al. Observations



CCSM3.5 with Langmuir



CCSM3.5 Control without Langmuir



Estimate the global climatological influence of wave driven mixing?

Apply 1-D ocean mixing models with parameterisation of wave driven mixing globally with realistic forcing (surface fluxes, waves) over a full annual cycle.

FORCING

- Surface forcing from CORE Normal Yr, using mixed layer model SST
- Waves (1° resolution CORE Normal Yr forced run, Full Spectra at 4° res.)

INITIALISATION

- Real ARGO profile. Most representative summer solstice (shallow MLD) profile (taken within 1.5 degree radius of wave archive location, +/- 20 days from summer solstice date.

MIXING MODELS (x2)

- Harcourt (2012) Second-moment closure with langmuir parameterisation
- PWP with amended Li and Garrett (1997) langmuir parameterisation

CSIRO

Harcourt SMC model with langmuir turbulence (2013, JPO)

CL vortex force (Craik and Leibovich, 1976) included in momentum equation after McWilliams et al (1997):

$$\frac{Du_j}{Dt} = -\frac{\partial p^*}{\partial x_j} - g_j\alpha\theta - \varepsilon_{jkl} f_k \left(u_l + u_l^S\right) + \varepsilon_{jpl}\varepsilon_{lmn} u_p^S \frac{\partial u_n}{\partial x_m} + \nu\nabla^2 u_j, \quad where \quad p^* = p + u_k^S(u_k + u_k^S/2) \quad \& \quad p = p^* = P/\rho_0$$

u_j is surface-wave phase-averaged Eulerian velocity,
u_j^S is the surface wave Stokes drift,
P is non-hydrostatic pressure,
ρ_0 is reference density,
f_k is Coriolis components,
g_k is gravitational acceleration,
ν is viscosity
θ is a thermodynamic scalar with expansion coefficient α and diffusivity κ_θ.

Kantha and Clayson (2004) included this production term

Deriving Reynolds stress and flux equations for fluctuations u_j', θ□ and a slowly- or non-fluctuating Stokes drift gives:

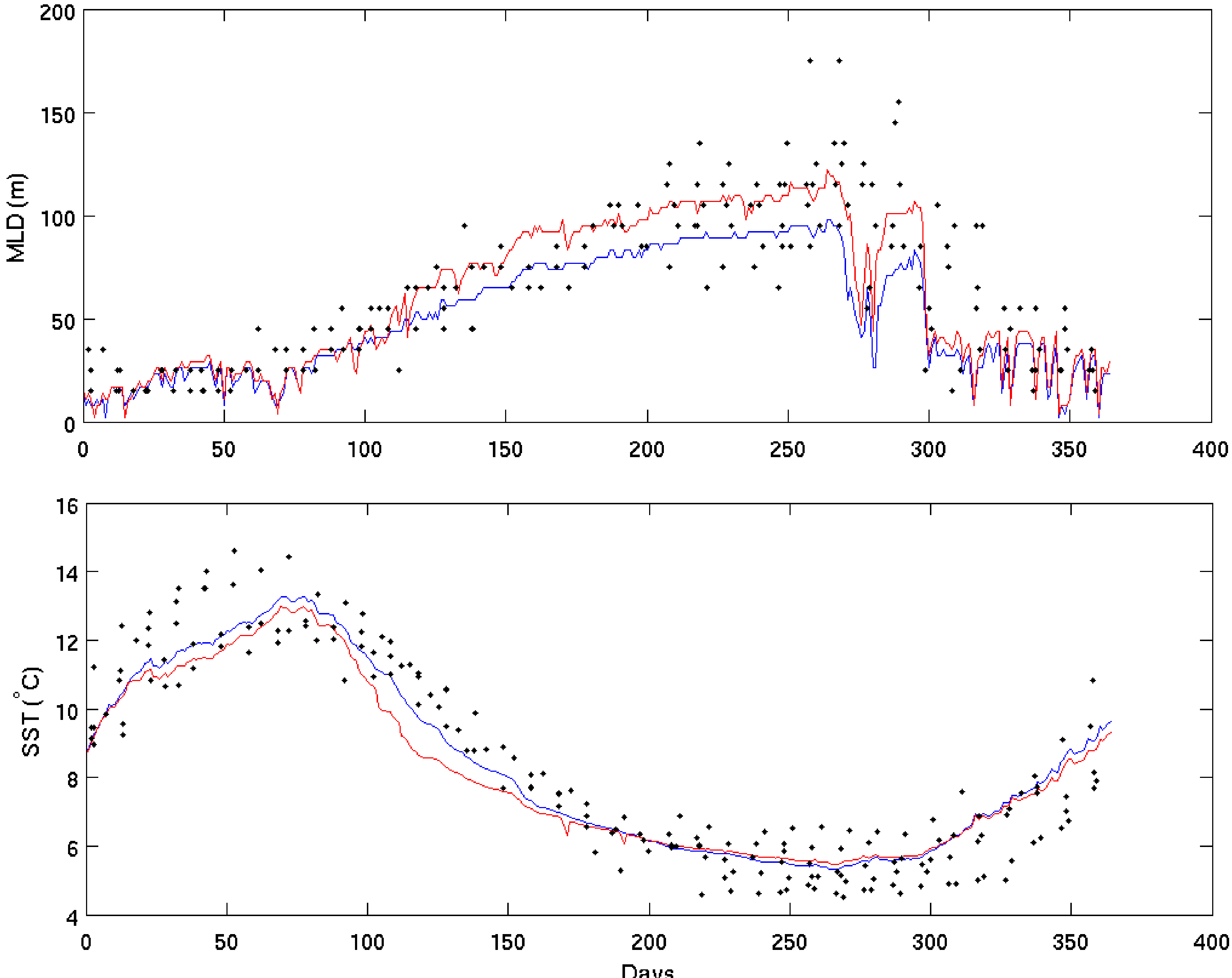
$$\frac{D^L \overline{u_i'u_j'}}{Dt} + \frac{\partial \overline{u_k'u_i'u_j'}}{\partial x_k} + \left(\overline{u_i'\frac{\partial p'}{\partial x_j}} + \overline{u_j'\frac{\partial p'}{\partial x_i}}\right) - \nu\nabla^2\overline{u_i'u_j'} = -2\nu\overline{\frac{\partial u_i'}{\partial x_k}\frac{\partial u_j'}{\partial x_k}} - \left(\overline{u_i'u_k'}\frac{\partial \bar{u}_j}{\partial x_k} + \overline{u_j'u_k'}\frac{\partial \bar{u}_i}{\partial x_k}\right) - \left(\overline{u_i'u_k'}\frac{\partial u_k^S}{\partial x_j} + \overline{u_j'u_k'}\frac{\partial u_k^S}{\partial x_i}\right)$$

$$-\alpha\left(g_j\overline{u_i'\theta'} + g_i\overline{u_j'\theta'}\right) - f_k\left(\varepsilon_{jkl}\overline{u_i'u_l'} + \varepsilon_{jkl}\overline{u_j'u_l'}\right) \quad , \quad where \quad \frac{D^L u_j}{Dt} = \frac{\partial u_j}{\partial t} + (u_k + u_k^S)\frac{\partial u_j}{\partial x_k}, \quad and$$

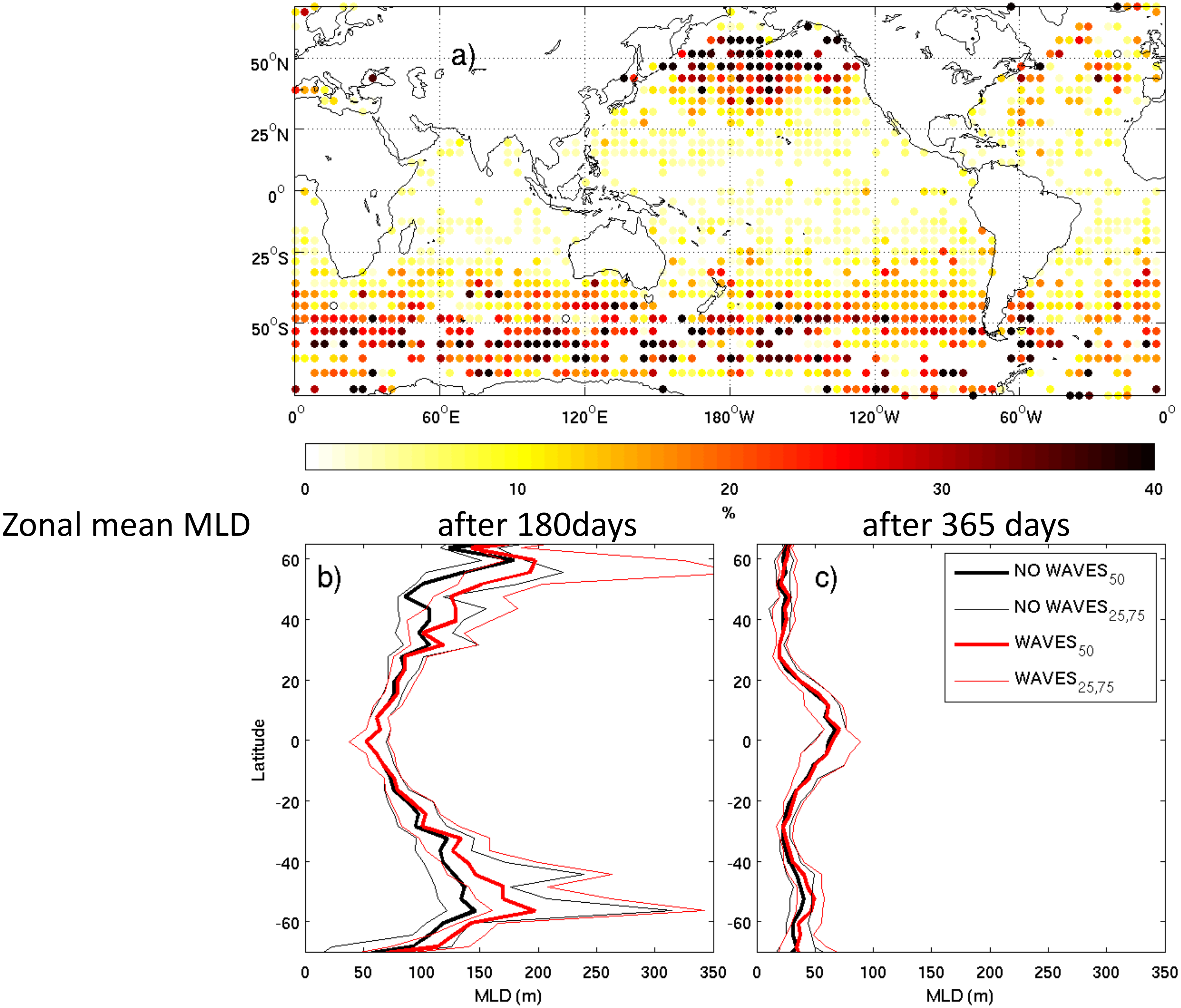
$$\frac{D^L\overline{u_j'\theta'}}{Dt} + \frac{\partial\overline{u_k'u_j'\theta'}}{\partial x_k} + \overline{\theta'\frac{\partial p'}{\partial x_j}} - \frac{\partial}{\partial x_k}\left(\kappa_\theta\overline{u_i'\frac{\partial\theta'}{\partial x_j}} + \nu\overline{\theta'\frac{\partial u_i'}{\partial x_j}}\right) = -\left(\kappa_\theta+\nu\right)\overline{\frac{\partial u_j'}{\partial x_j}\frac{\partial\theta_i'}{\partial x_j}} - \overline{u_j'u_k'}\frac{\partial\bar{\theta}}{\partial x_k} - \overline{u_k'\theta'}\frac{\partial \bar{u}_j}{\partial x_k} - f_k\varepsilon_{jkl}\overline{u_l'\theta'} - \alpha g_j\overline{\theta'\theta'} - \overline{\theta' u_k'}\frac{\partial u_k^S}{\partial x_j}$$

Harcourt SMC

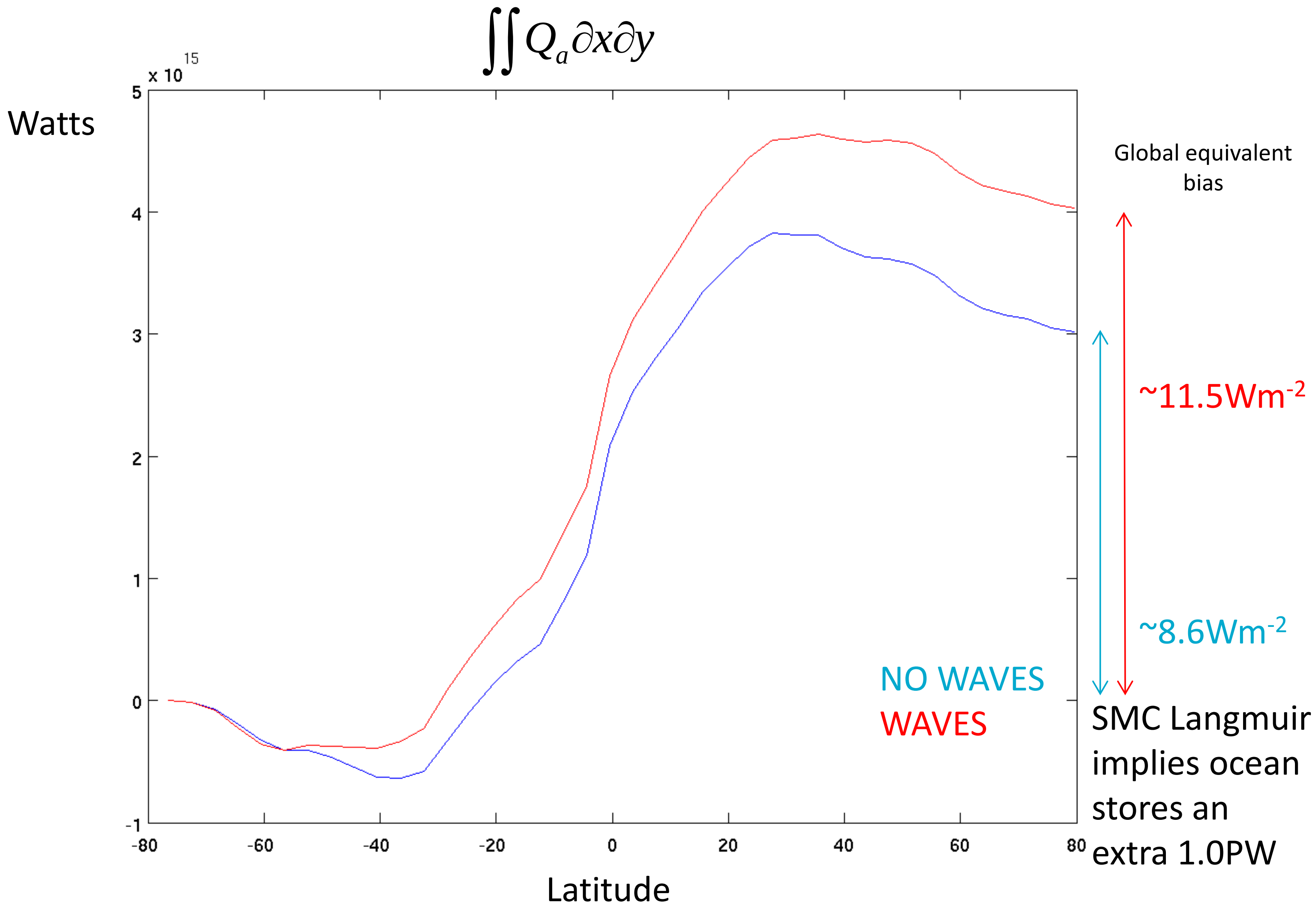
50N, 145W PAPA NO WAVES WAVES ARGO PROFILES



Percentage increase in MLD with introduction of SMC (E6=7) langmuir mixing ~180 days after Summer Solstice

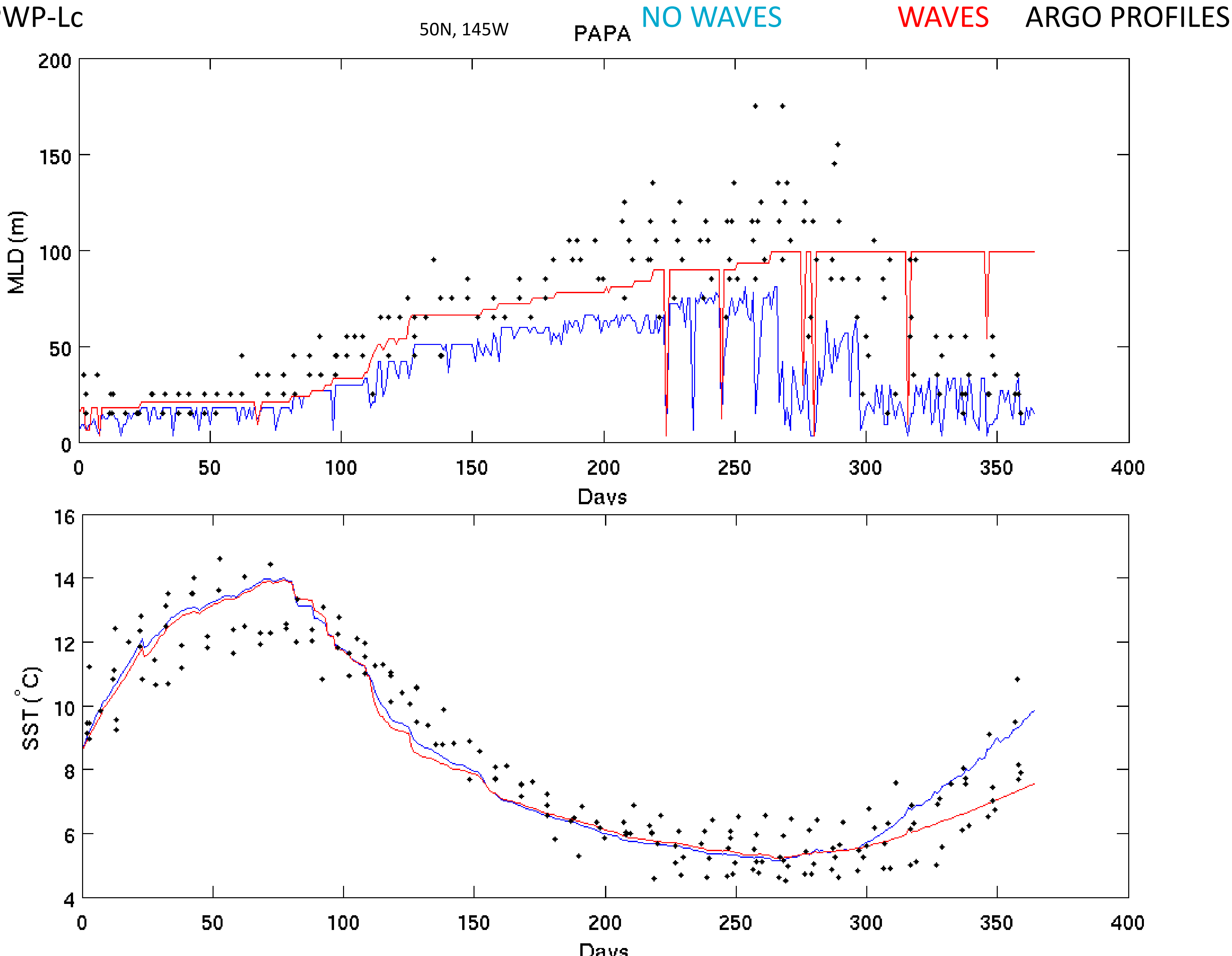


Implied Annual Mean Northward Heat Transport (Wm^{-2})



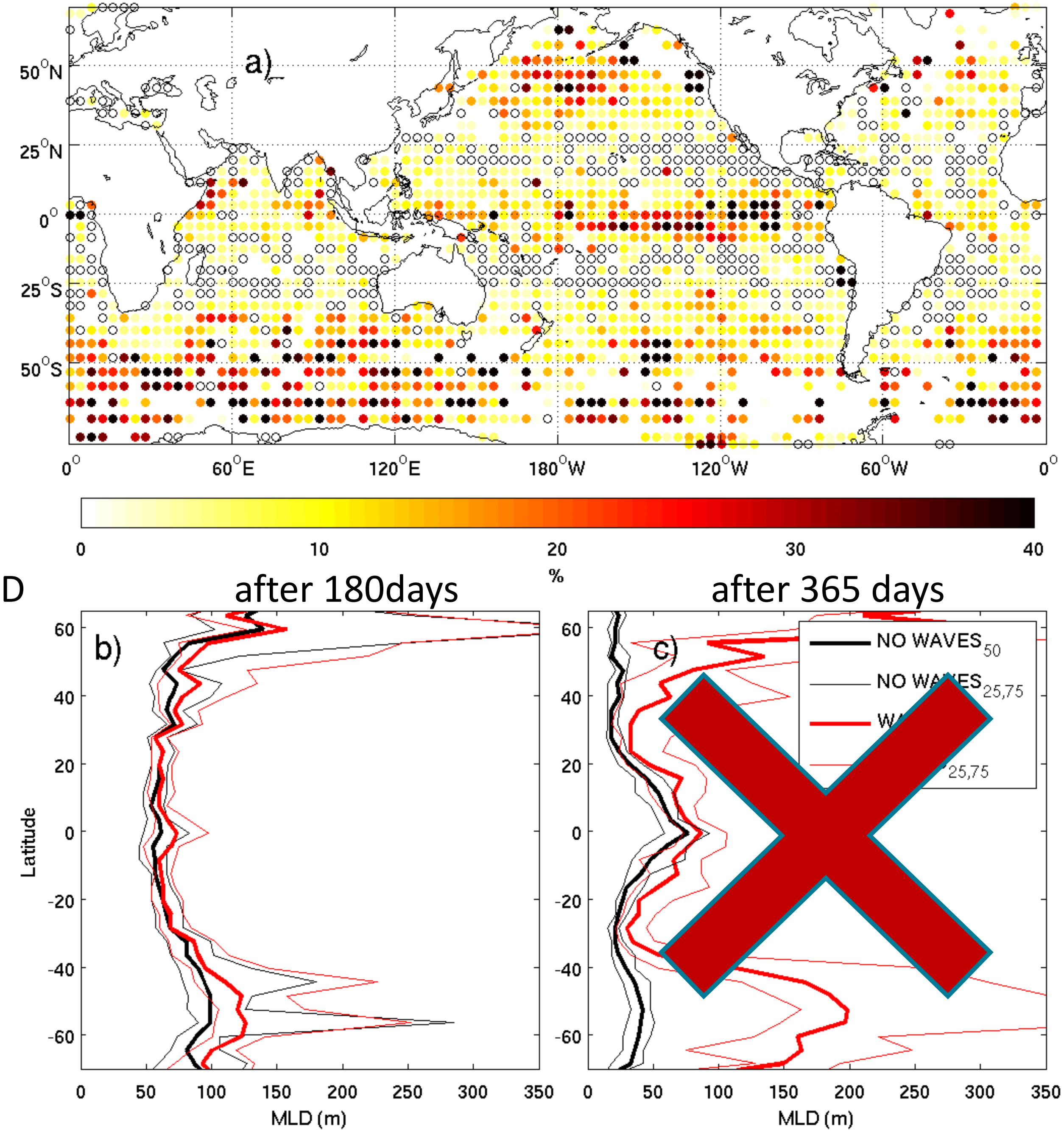
Note: Plot not directly comparable to Atm flux calcs. Can not assume that storage is uniformally distributed, as 1d model with/without waves is calculating spatial distribution of storage. If we remove this influence (by the above assumption applied previously to ensure convergence to zero at Northern boundary), the lines overlay one another.

PWP-Lc


50N, 145W
PAPA
NO WAVES
WAVES
ARGO PROFILES
MLD (m)
Days
SST (°C)
Days

Percentage increase in MLD with introduction of PWP langmuir mixing ~180 days after Summer Solstice

Zonal mean MLD



Summary of Mixing Model contributions

- Harcourt SMC (E6=7) => 1.0 PW additional storage
- Harcourt SMC (E6=5) => 0.77 PW additional storage
- Harcourt SMC (KanthaClaysonApprox) => 0.35 PW additional storage
- PWP + Langmuir => 1.0 PW additional storage

Conclusions of contribution of waves to climate system.

- Quantitative estimates of the contribution of waves in the coupled climate system have been determined.
- Global heat and momentum budgets display considerable sensitivity to available parameterisations of drag, with seastate dependent parameterisations resulting in a range of up to 1PW of additional heat transfer to the ocean. This supposed contribution however remains within the bounds set by alternative wind-dependent parameterisations of drag.
- Wave driven forcing of 1-d mixing models applied globally show an approximate 25% increase in mixed layer depth in extra-tropical storm belts, which is greater during the winter mixing season. Expressed as a surface heat flux, this is equivalent to up to $10 Wm^{-2}$, or an additional heat uptake of ~1PW to the global ocean over one year.
- Estimates of the contribution of other wave processes to follow

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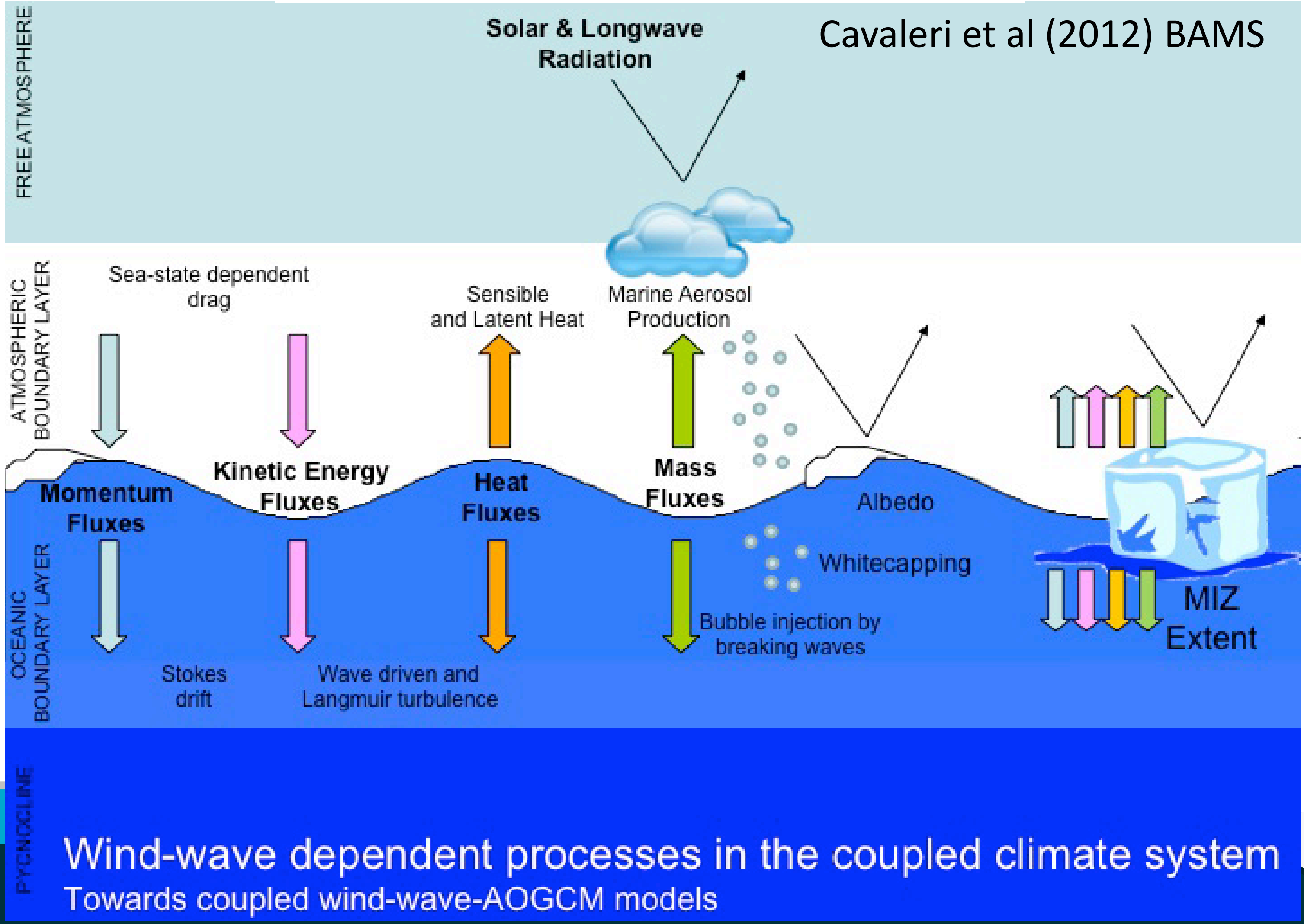
Gravity wind-wave–driven processes at the ocean surface—including radiation fluxes and energy, mass, and momentum exchanges—play an important role in the coupled climate system.



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Thank you

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A Partnership between CSIRO and the Bureau of Meteorology

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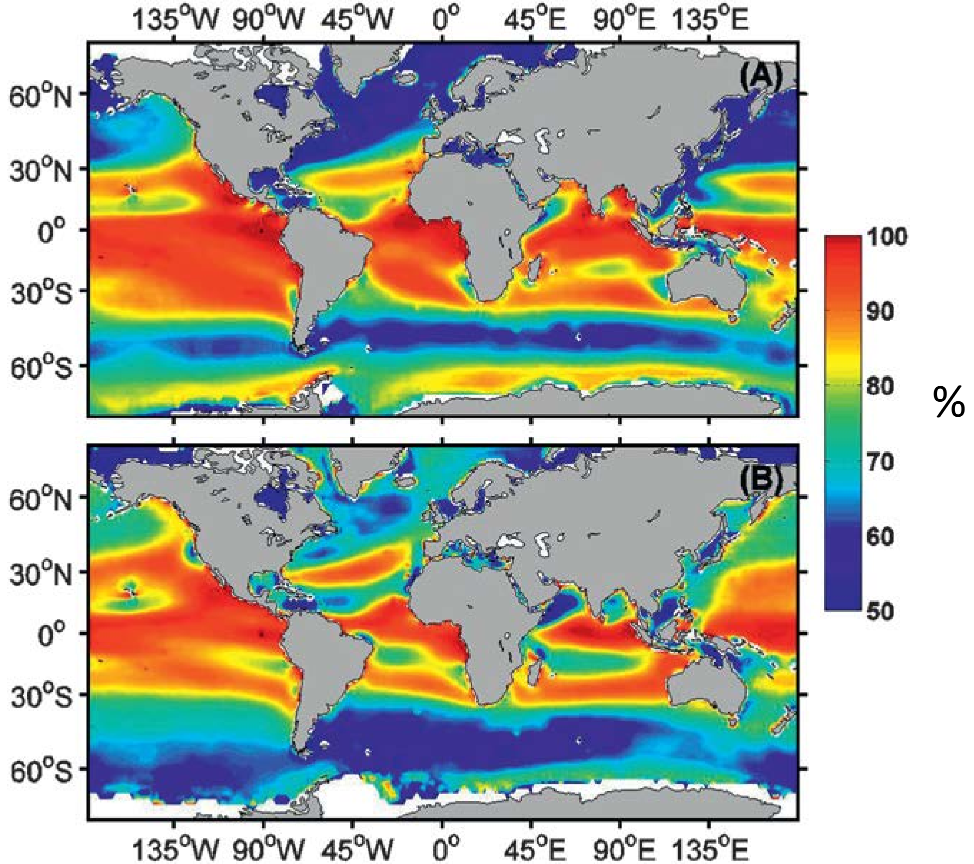
Wind and waves not in equilibrium

- Swell dominates global wave field. (Semedo et al., 2011)
- Global distribution of fraction of wave energy which is swell for DJF and JJA.



Southern Ocean GCM wind bias

Swart and Fyfe, 2011

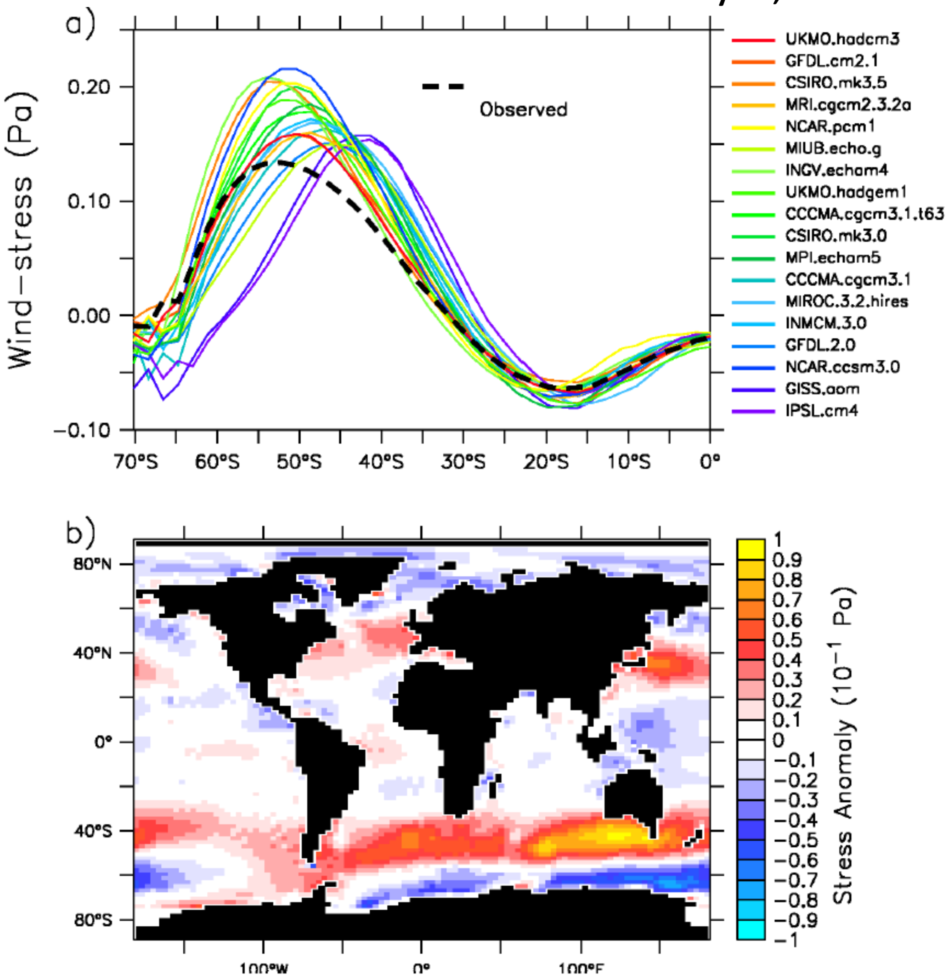


Figure S1: **Wind-stress comparisons. a,** The zonal mean pre-industrial wind-stress from the 18 CMIP3 models used in this study. The heavy black dashed line shows the observationally derived pre-industrial wind-stress; **b,** Zonal wind-stress anomaly map, computed as the ensemble mean of the 18 CMIP3 wind-stresses minus the observationally derived pre-industrial wind-stress.

SO winds drive large components of present and future ocean uptake of heat and CO2.
=>This bias has broad implications for GCM's

Sea-state dependent drag influence

The Southern Ocean Wind Bias

Janssen and Viterbo (1996)
Sea-state dependent drag in
Seasonal prediction model

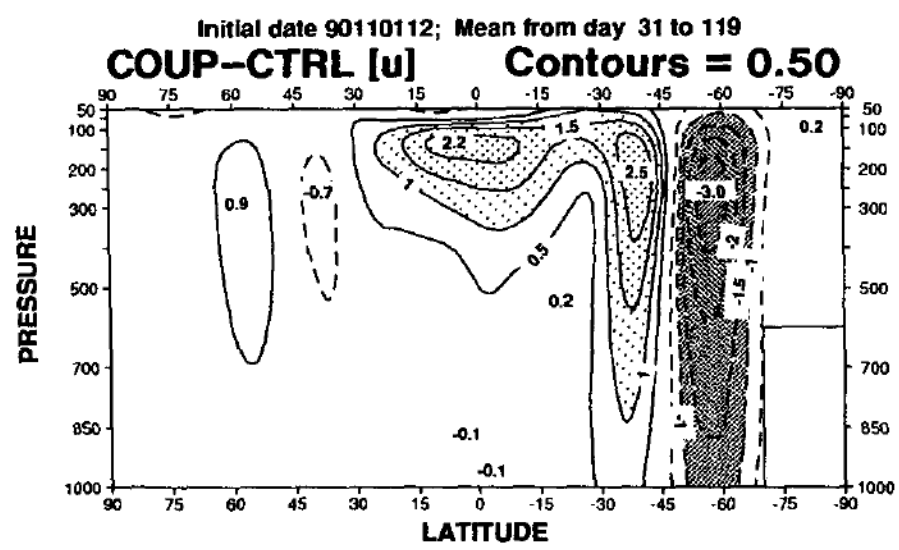
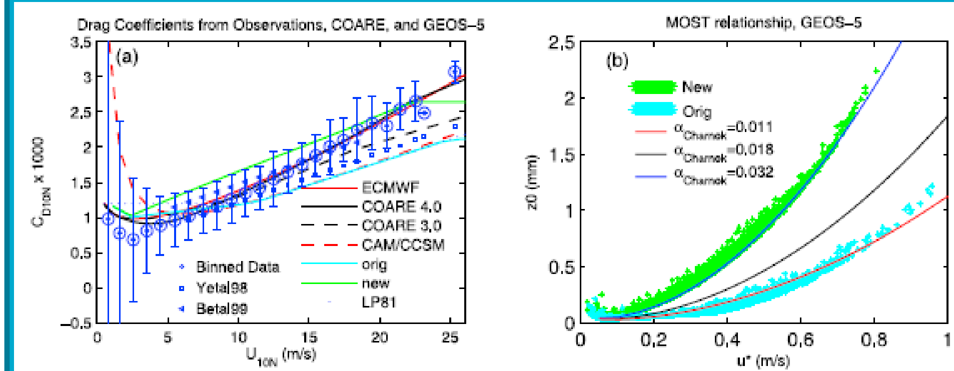
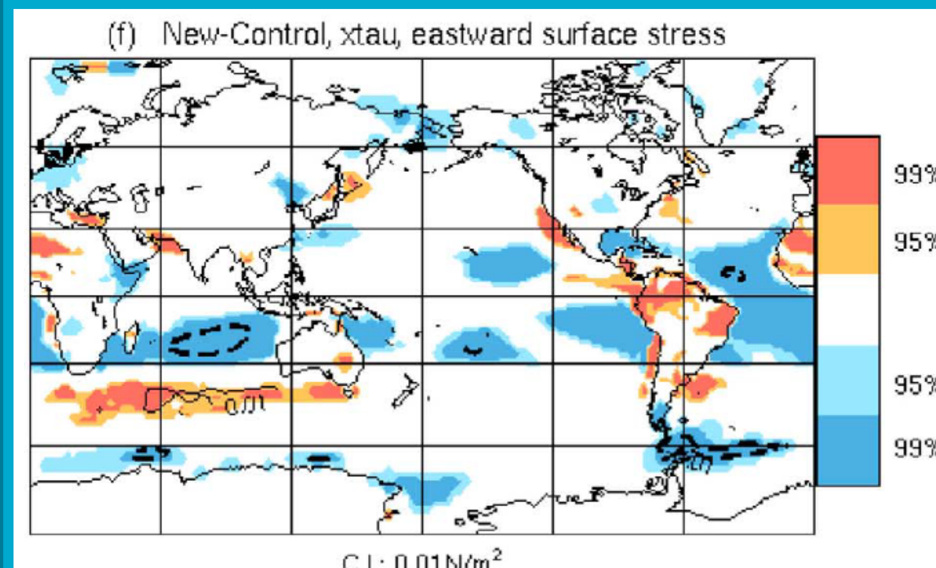


FIG. 8. Latitude–height cross sections of the differences in the zonal mean wind.

Garfinkel et al. (2011)
Increased ocean roughness in GEOS-5
GCM improved SO wind bias.





CORE (Large and Yeager, 2004, 2009)

Standard air-sea flux dataset of WGOMD

Atmospheric Fields

- NCEP/NCAR
  - Near surface winds, U
  - Near surface atmospheric temperature, θ
  - Near surface specific humidity, q

Radiation

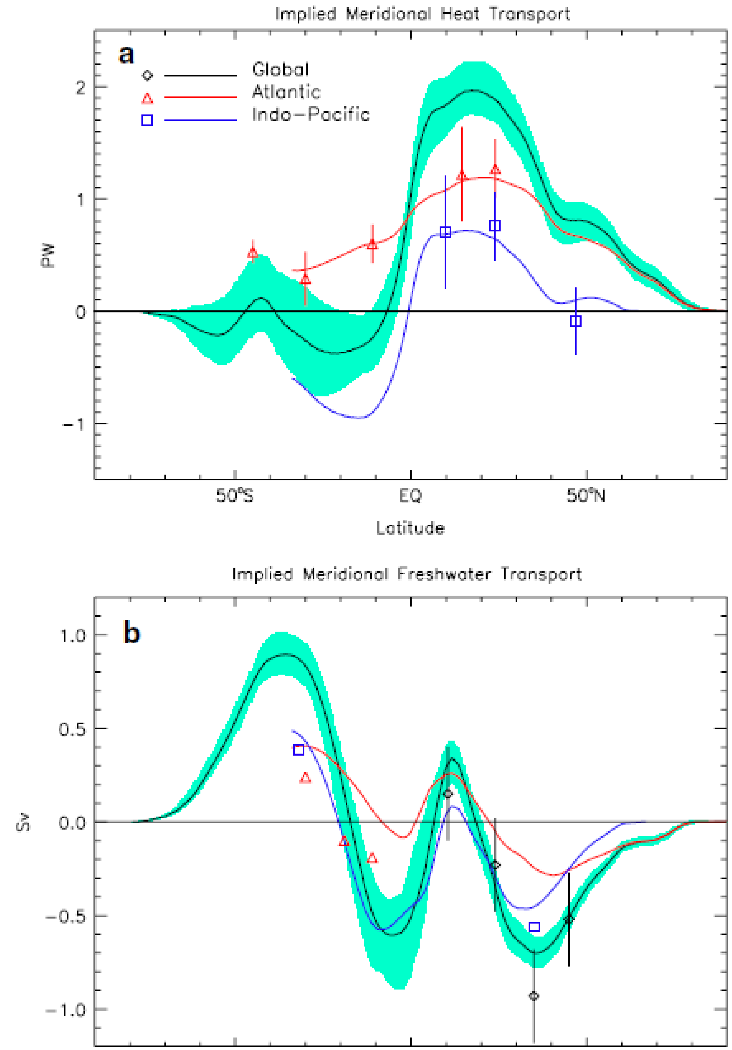
- International Satellite Cloud Climatology Experiment
  - Short wave insolation, Q_I
  - Downwelling Long wave Radiation, QA

Precipitation

- GCGCS (Merged GPCP, CMAP, S-H-Y data)

SST

- Hadley Centre sea Ice and SST dataset version 1 (HadISST1)



Bulk air-sea fluxes

- Bulk flux formulae

$$\overline{w'x'} = c_x^{1/2} c_d^{1/2} S \Delta X = C_x S \Delta X$$
$$\Delta X = X_{sea} - X(z); S = windspeed$$

- Neutral Conditions

$$c_{xn}^{1/2} = \frac{\kappa}{\ln(z / z_{ox})}$$

Z_{ox} – roughness length

- Correction dependent on surface stability

$$c_x^{1/2}(\zeta) = \frac{c_{xn}^{1/2}}{\left[1 - \frac{c_{xn}^{1/2}}{\kappa}\psi_x(\zeta)\right]}$$

Comparisons with CORE.v2

Heat budget closure is dependent on parameterisation of transfer coefficient.

$$C_{DN10} = \left(\frac{2.7}{u_{10}} + 0.142 + 0.0764 u_{10} \right) \Big/ 1000$$

Roughness length

- Charnock relation (constant coefficient, plus smooth flow limit)

$$z_o = \alpha \frac{u_*^2}{g} + \left(\frac{0.11\nu}{u_*} \right) \quad \text{where} \quad C_{DN10} = \frac{\kappa^2}{\log\left(\left(\frac{10}{z_o} \right)^2 \right)}$$

- Other parameterisations suggest zo is a function of wave age, steepness or stress.

E.g., Oost et al. (2002),

$$z_o = \frac{50}{2\pi} \lambda_p \left(u_* / C_p \right)^{4.5} + \left(\frac{0.11\nu}{u_*} \right)$$

and Taylor and Yelland (2001)

$$z_o = 1200 h_s \left(h_s / \lambda_p \right)^{4.5} + \left(\frac{0.11\nu}{u_*} \right)$$

And Janssen and Viterbo (1996)

$$z_o = \alpha \frac{u_*^2}{g} + \left(\frac{0.11\nu}{u_*} \right) \qquad \alpha = \beta \left(1 - \frac{\tau_w}{\tau} \right)^{-1/2}$$

Stokes' drift

$$\boldsymbol{u}^S = \frac{16\pi^3}{g} \int_0^\infty \int_{-\pi}^{\pi} (\cos\theta, \sin\theta, 0) f^3 S_{f\theta}(f,\theta) e^{\frac{8\pi^2 f^2}{g} z} d\theta df.$$

PWP

Static stability $$\frac{\partial \rho}{\partial z} \geq 0$$

Mixed layer stability $$R_b = \frac{g \Delta \rho h}{\rho_0 (\Delta \mathbf{V})^2} \geq 0.65$$

Shear flow stability $$R_g = \frac{g \partial \rho / \partial z}{\rho_0 (\partial \mathbf{V} / \partial z)^2} \geq 0.25$$

Δ h	mixed layer depth difference between mixed layer and the level just beneath.
Rb	bulk richardson number
Rg	gradient richardson number

Incorporation of LC into the PWP model (Li et al.,1995)

Langmuir cells penetration depth depends on competition between vertical motion and stratification, represented by the Froude number.

$$Fr = \frac{w_{dn}}{Nh}$$

Vertical penetration is inhibited when Fr reaches a critical value Frc = 0.6 (LG97). $Fr \leq Fr_c$

This was parameterised by Li and Garret as:

$$w_{dn} = 0.72\left(\frac{u_s}{u_*}\right)^{1/3} La^{-1/3} u_* \qquad \text{(after LG93)}$$

giving

$$h = 1.2\left(\frac{u_*}{N}\right)\left(\frac{u_s}{u_*}\right)^{1/3} La^{-1/3}$$

$$\Delta b = \frac{1}{2} h N^2$$

So, stability occurs if

$$\frac{\Delta b}{\left(h u_*^2\right)} \geq 50$$

50 is taken as fully developed sea case of Fr_c = 0.72(us/u*)^(2/3). La^(-2/3)
La being the langmuir number (not the turbulent La).

Amended Fr scaling of Lc in PWP

Flor et al. (2010, JGR) suggest $w_{dn} = 5.2 w_{rms.}$

Van Roekel et al (2012) give scaling of :

$$w_{rms}^2 = 0.6\, u_*^2 (1.0 + (c1\, La_t)^{-2} + (c2\, La_t)^{-4}),$$ where c1 = 1.5 and c2=5.4.

For the case where wind and waves are non-aligned,

$$La_t^2 = La_{SLproj}^2 = u^*\cos(\alpha)/(u_{s,0.2ML} \cos(\theta_{ww} - \alpha))$$

α = angle b/w wind and langmuir cell direction, θ_{ww} is angle b/w stokes drift and wind.

$$\alpha \approx \text{atan}(\sin(\theta_{ww})/(u_*/(u_{s0}\,\kappa)).(\log(H_{ml}/z_1) + \cos(\theta_{ww}))$$

So that Stability occurs if :

$$Fr = 5.2 * \sqrt{(w_{rms}^2/(g\Delta\rho h))} <= 0.6$$