New results on atmosphere-ocean interaction

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INTRODUCTION

In this talk I would like to illustrate that ocean waves play an important role in the interaction between atmosphere and ocean.

Ocean waves play a role in air-sea **momentum transfer** and in **upper ocean mixing**. The associated Stokes drift combined with the earth' rotation results in a additional force on the mean ocean circulation: the **Stokes-Coriolis force**. Also, momentum transfer and the sea state are affected by surface currents, hence it makes sense to introduce a **three-way coupling** between atmosphere, ocean circulation and surface waves. The end result is **one model for the geosphere**. At ECMWF, a first version of such a model will be introduced shortly in the ensemble prediction system and in the monthly forecasting system.

The programme of my talk is therefore as follows:

- **Air-sea interaction**

  Describe the scheme to model effects of the waves on the air-sea momentum transfer. Basically, when the sea state is young the waves are steep and are potentially extracting more momentum from the air flow then when the sea state is old and the waves are gentle. The enhanced momentum transfer usually leads to a more rapid filling up of the pressure lows.

  This effect improves the atmospheric climate on a seasonal time scale and also improves forecast skill in the medium range (both atmosphere and ocean waves).

- **Upper-Ocean Mixing**

  Upper ocean mixing is to a large extent caused by breaking, dissipating ocean waves. As a consequence there is an energy flux Φ_{oc} from atmosphere to ocean. It is given by $\Phi_{oc} = m\rho_a u_*^3$ where m depends on the sea state. Wave breaking and its associated mixing penetrates into the ocean at a scale of the order of the significant wave height H_S. In addition, the shear in the Stokes drift gives an extra production of turbulent kinetic energy which penetrates into the ocean at a scale of the order the typical wavelength of the surface waves.

  Developed a simple scheme to model these effects and applied it to the diurnal cycle in SST.

- **New Results**

  At ECMWF we have build a new version of our ensemble prediction system. This system consists of a coupled atmosphere, ocean-wave, ocean circulation model where each component interacts with the two others. The model includes a sea state dependent momentum and heat exchange, the wave-induced upper ocean mixing (where Stokes drift and breaking energy flux are supplied by the wave model) and Stokes-Coriolis forcing while the currents affect both the momentum exchange between air and water and the ocean wave propagation.

  This coupled system shows certain improvements in forecast skill of the ensemble prediction system, in particular in the Tropics.

Air-Sea INTERACTION

Discuss the basic air-sea interaction model. Ocean waves, described by the action spectrum $N(\mathbf{k};\mathbf{x},t)$, are governed by the **action balance equation**

$$\frac{D}{Dt}N = S = S_{in} + S_{nl} + S_{ds},$$

and the source functions S represent the physics of wind input, dissipation by wave breaking and nonlinear four-wave interactions. Air-sea interaction is governed by momentum conservation. In the steady state:

$$\tau = \tau_w(z) + \tau_{turb}(z),$$

with $\tau_w(z)$ the wave induced stress profile with surface value

$$\tau_w = \left.\frac{\partial \mathbf{P}}{\partial t}\right|_{wind} = \rho_w \int \mathrm{d}\omega \mathrm{d}\theta \, \mathbf{k} S_{in}.$$

This then results in a dimensionless roughness length, or Charnock parameter, as given by

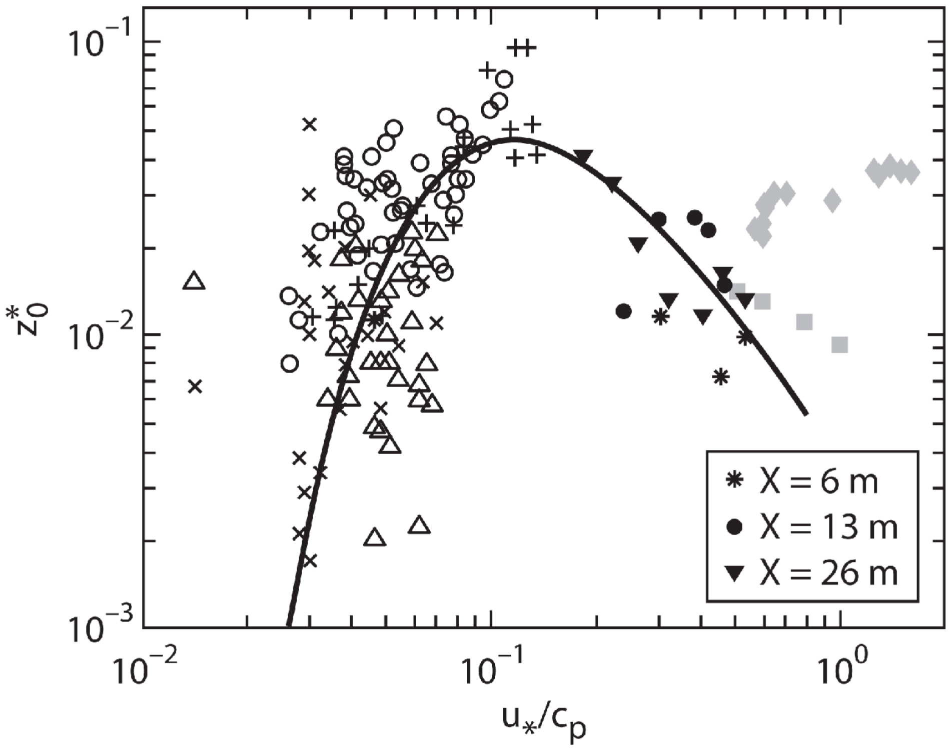
$$z_0^* = \frac{g z_0}{u_*^2} = \frac{\alpha}{\sqrt{1 - \frac{\tau_w}{\tau}}}, \quad \alpha \simeq 0.01$$

and depends on the ratio of wave-induced stress τ_w to total stress τ.

Using the Charnock parameter the neutral drag coefficient is given by

$$C_D(10) = \frac{u_*^2}{U_N(10)^2} = \left(\frac{\kappa}{\log(10/z_0)} \right)^2$$

As the coupled system results in a sea state dependent Charnock parameter, the drag over the ocean is sea state dependent as well. This is illustrated below where observed Charnock parameter is plotted against the inverse of the wave age parameter c_p/u_*. The wave age parameter measures the stage of development of windsea.

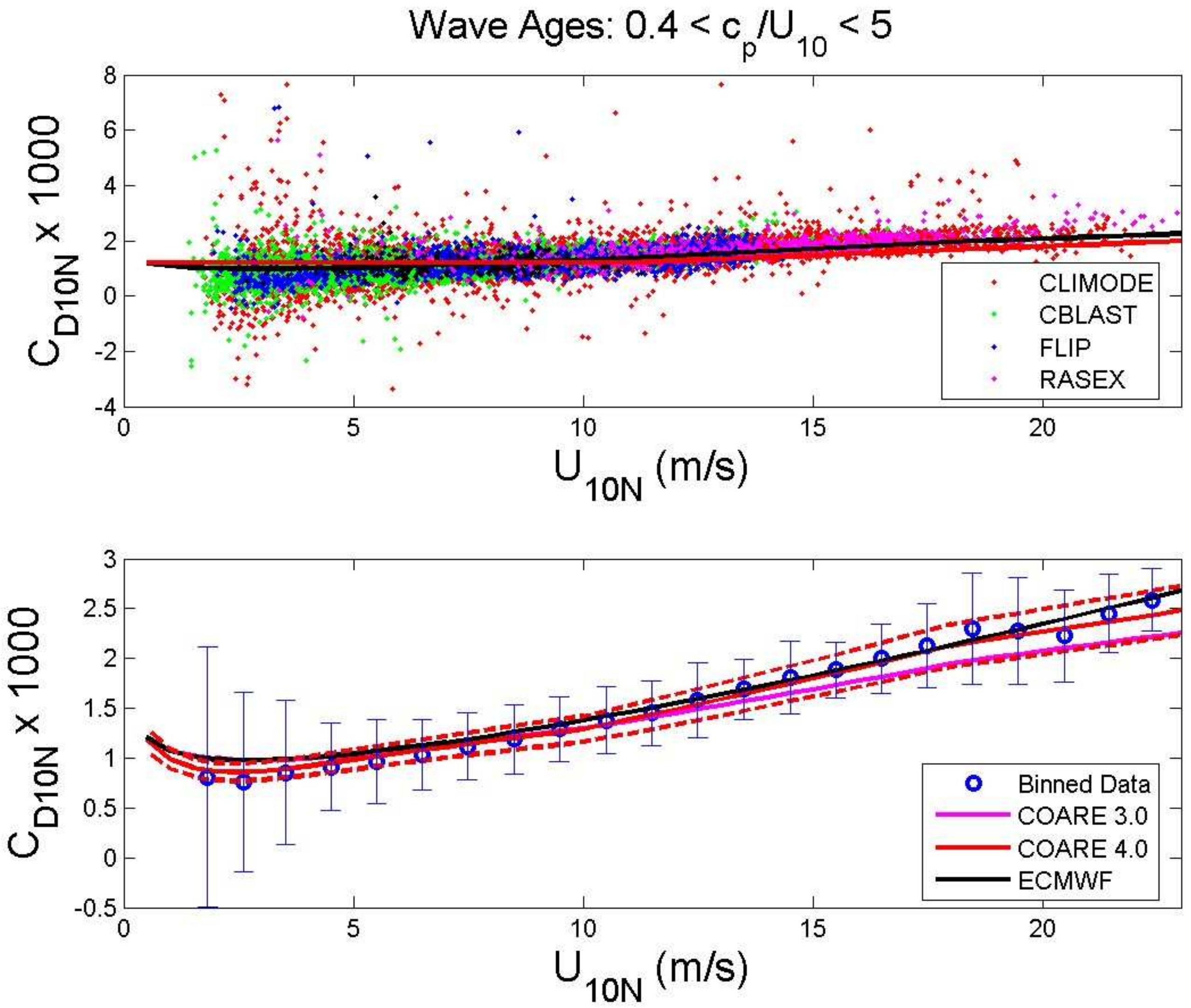


10^{-1}
10^{-2}
10^{-3}
z_0^*
10^{-2}
10^{-1}
10^{0}
u_*/c_p
X = 6 m
X = 13 m
X = 26 m

The graph of Charnock parameter versus inverse wave age shows two regimes: for **extreme young windseas** roughness increases with wave age (occurs in Hurricane conditions), while for larger wave ages but still **young** windseas the roughness decreases with wave age.

The first regime hardly ever occurs, so let us give some results for the ’normal’ regime of young windseas.

- Check on statistical properties of the ECMWF coupled system: compare average drag as function of windspeed with most recent observations.
- Impact on Model Climate.
- Impact on forecast skill.


Wave Ages: 0.4 < c_p/U_{10} < 5
C_{D10N} x 1000
U_{10N} (m/s)
CLIMODE
CBLAST
FLIP
RASEX
C_{D10N} x 1000
U_{10N} (m/s)
Binned Data
COARE 3.0
COARE 4.0
ECMWF

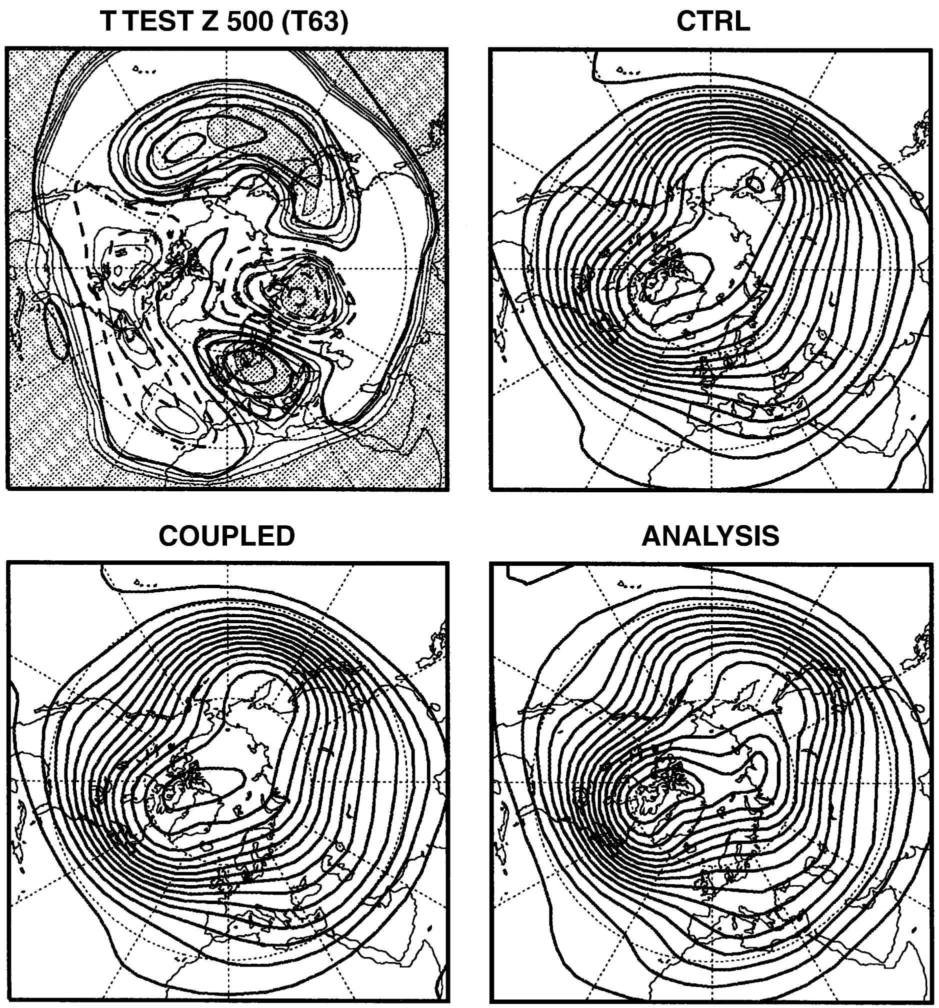
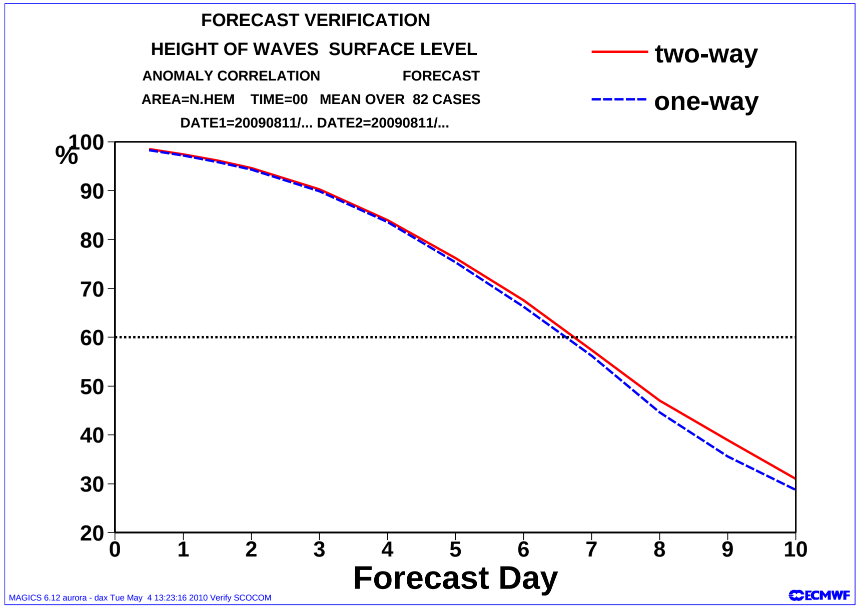


Figure 1: Ensemble mean of 500 mb geopotential height of coupled and control run and their differences.Period is winter 1990 and area is Northern Hemisphere. Heavy shading means that there is a probability of 95% that the difference is significant.


FORECAST VERIFICATION
HEIGHT OF WAVES SURFACE LEVEL
ANOMALY CORRELATION FORECAST
AREA=N.HEM TIME=00 MEAN OVER 82 CASES
DATE1=20090811/... DATE2=20090811/...
two-way
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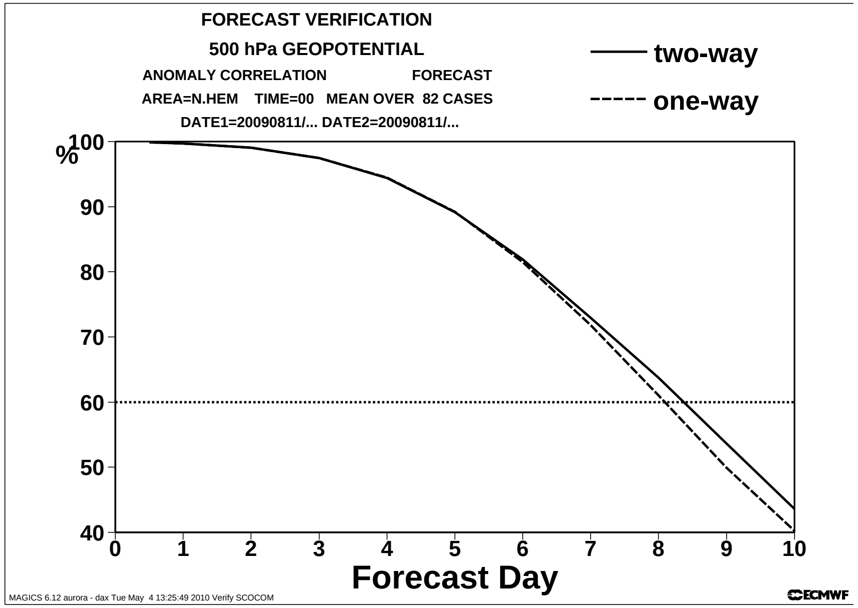
Forecast Day
MAGICS 6.12 aurora - dax Tue May 4 13:23:16 2010 Verify SCOCOM
ECMWF


FORECAST VERIFICATION
500 hPa GEOPOTENTIAL
ANOMALY CORRELATION FORECAST
AREA=N.HEM TIME=00 MEAN OVER 82 CASES
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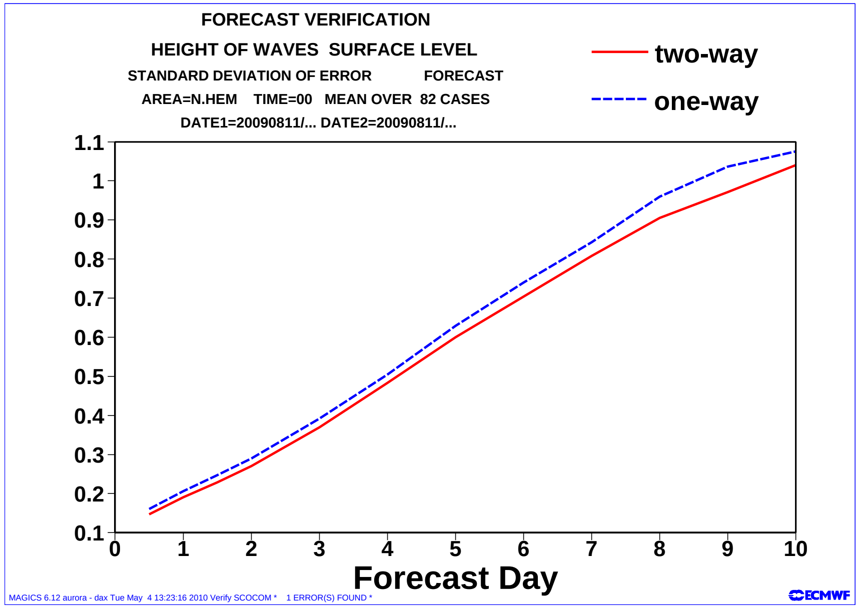
FORECAST VERIFICATION
HEIGHT OF WAVES SURFACE LEVEL
STANDARD DEVIATION OF ERROR FORECAST
AREA=N.HEM TIME=00 MEAN OVER 82 CASES
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two-way
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Forecast Day
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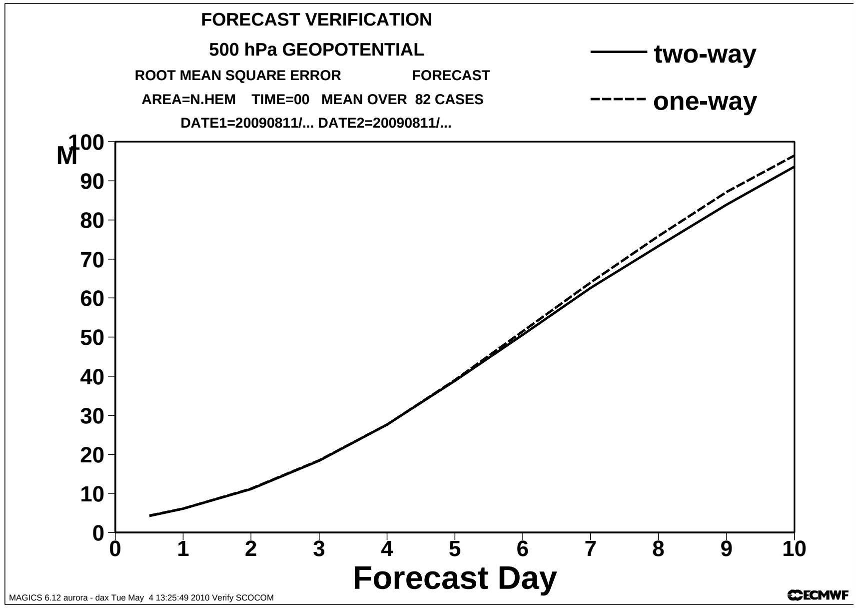
ECMWF


FORECAST VERIFICATION
500 hPa GEOPOTENTIAL
ROOT MEAN SQUARE ERROR FORECAST
AREA=N.HEM TIME=00 MEAN OVER 82 CASES
DATE1=20090811/... DATE2=20090811/...
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MAGICS 6.12 aurora - dax Tue May 4 13:25:49 2010 Verify SCOCOM
ECMWF

DIURNAL CYCLE IN SST

In the past 15 years observational evidence has been presented about the role of wave breaking and Langmuir turbulence in the upper ocean mixing.

For diurnal cycle simulation only wave breaking is relevant. It can be shown that near the surface, in a layer of the order of the wave height H_S, the turbulent velocity is enhanced by a factor of 2-3, while, in agreement with observations there is an enhanced turbulent dissipation. This deviates from the 'law-of-the-wall'.

Combined with a proper modelling of buoyancy effects a realistic simulation of the diurnal cycle may be obtained. Here, the energy flux from waves to ocean column follows from the dissipation term in the action balance equation:

$$\Phi_{oc} = -\rho_w \int \mathrm{d}\omega \mathrm{d}\theta \; \omega S_{ds}.$$

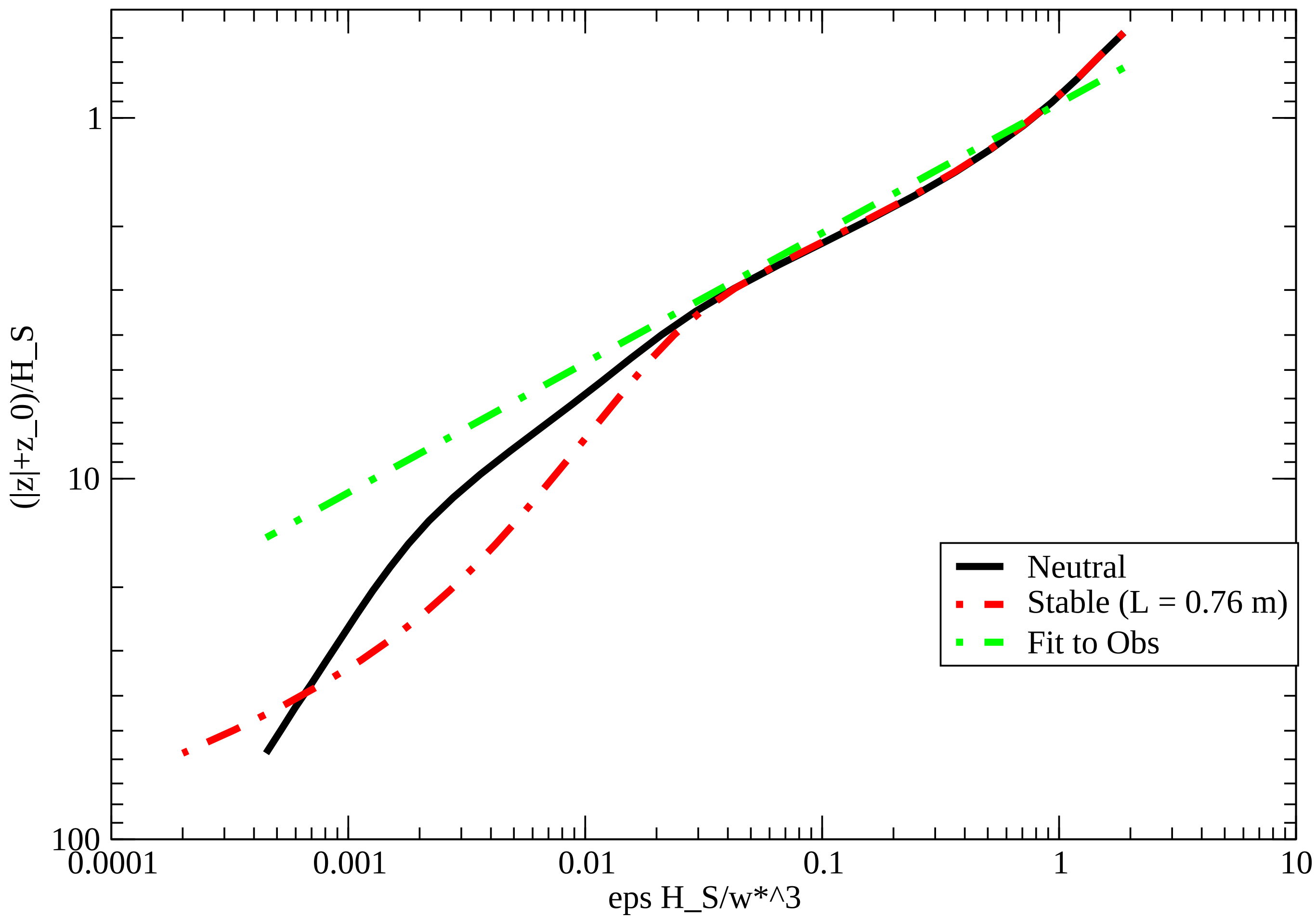


Figure 2: Dimensionless dissipation $\varepsilon_* = \varepsilon H_S/\Phi_{oc}$ versus $(z+z_0)/H_S$

TKE EQUATION

The enhanced turbulent dissipation can be described in the context of the turbulent kinetic energy (TKE) equation. If effects of advection are ignored, the TKE equation describes the rate of change of turbulent kinetic energy e due to processes such as shear production (including the shear in the Stokes drift), damping by buoyancy, vertical transport of pressure and TKE, and turbulent dissipation ε. It reads

$$\frac{\partial e}{\partial t} = \nu_m S^2 + \nu_m S \frac{\partial U_S}{\partial z} - \nu_h N^2 + \frac{1}{\rho_w} \frac{\partial}{\partial z} (\overline{\delta p \delta w}) + \frac{\partial}{\partial z} (\overline{e \delta w}) - \varepsilon,$$

where $e = q^2/2$, with q the turbulent velocity, $S = \partial U/\partial z$ and $N^2 = g\rho_0^{-1}\partial\rho/\partial z$, with N the Brunt-Väisälä frequency, ρ_w is the water density, δp and δw are the pressure and vertical velocity fluctuations and the over-bar denotes an average taken over a time scale that removes linear turbulent fluctuations.

The turbulent production of Langmuir circulation is modelled by the second term on the right-hand side of the TKE equation which represents works against the shear in the Stokes drift. Here U_S is the magnitude of the Stokes drift for a general wave spectrum $F(\omega)$,

$$U_S = \frac{2}{g}\int_0^\infty d\omega\, \omega^3 F(\omega) e^{-2k|z|},\ k = \omega^2/g.$$

Although in principle the depth dependence of the Stokes drift is known it still is a fairly elaborate expression through the above integral. In the final result we will use the approximate expression

$$U_S = U_S(0) e^{-2k_S|z|},$$

where $U_S(0)$ is the value of the Stokes drift at the surface and k_S is an appropiately chosen wavenumber scale.

The dissipation term is taken to be proportional to the cube of the turbulent velocity divided by the mixing length $l = \kappa|z|$,

$$\varepsilon = \frac{q^3}{Bl},$$

Here, B is another dimensionless constant.

The pressure transport term can be determined by explicitely modelling the energy transport caused by wave dissipation. The correlation between pressure fluctuation and vertical velocity fluctuation at the surface is

$$I_w(0) = +\frac{1}{\rho_w}\overline{\delta p \delta w}(z=0) = -\int_0^\infty \omega S_{diss}(\mathbf{k})\,d\mathbf{k} = m\frac{\rho_a}{\rho_w}u_*^3 = m\frac{\rho_w^{1/2}}{\rho_a^{1/2}}w_*^3 = \alpha w_*^3$$

and the main problem is how to model the depth dependence of $\overline{\delta p \delta w}$. Assume depth scale is controlled by significant wave height H_S:

$$I_w(z) = +\frac{1}{\rho_w}\overline{\delta p \delta w} = I_w(0) \times \hat{I}_w,\ \hat{I}_w = e^{-|z|/z_0}$$

where the depth scale $z_0 \sim H_S$ will play the role of a roughness length. Thus, the TKE equation becomes

$$\frac{\partial e}{\partial t} = \frac{\partial}{\partial z}\left(lqS_q\frac{\partial e}{\partial z}\right) + \frac{\partial I_w(z)}{\partial z} + \nu_m S^2 + w_*^2\frac{\partial U_S}{\partial z} - \nu_h N^2 - \frac{q^3}{Bl(z)}.$$

At the surface there is no direct conversion of mechanical energy to turbulent energy and therefore the turbulent energy flux is assumed to vanish. Hence the boundary conditions become

$$lqS_q\frac{\partial e}{\partial z} = 0 \qquad \text{for} \qquad z = 0,$$

$$\frac{\partial e}{\partial z} = 0 \qquad \text{for} \qquad z \to \infty.$$

STEADY STATE PROPERTIES

NEUTRALLY STABLE

The properties of the steady state version of the TKE equation were studied extensively. Without presenting any of the details, for neutral stratification the following **’1/3’-rule** is found. Introducing the dimensionless turbulent velocity $Q = \left(S_M/B\right)^{1/4} \times q/w_*$ the approximate solution of the TKE equation becomes

$$w(z) = Q^3 \approx 1 + \alpha\kappa|z|\frac{d\hat{I}_w}{dz} + La^{-2}\kappa|z|\frac{d\hat{U}_S}{dz},$$

where $La = (w_*/U_S(0))^{1/2}$ is the turbulent Langmuir number. So in terms of Q^3 there is a **superposition principle**, i.e. contributions due to wave dissipation and Langmuir turbulence may be added to the shear production term.

The next graph shows the contributions of wave dissipation and Langmuir turbulence to the turbulent velocity

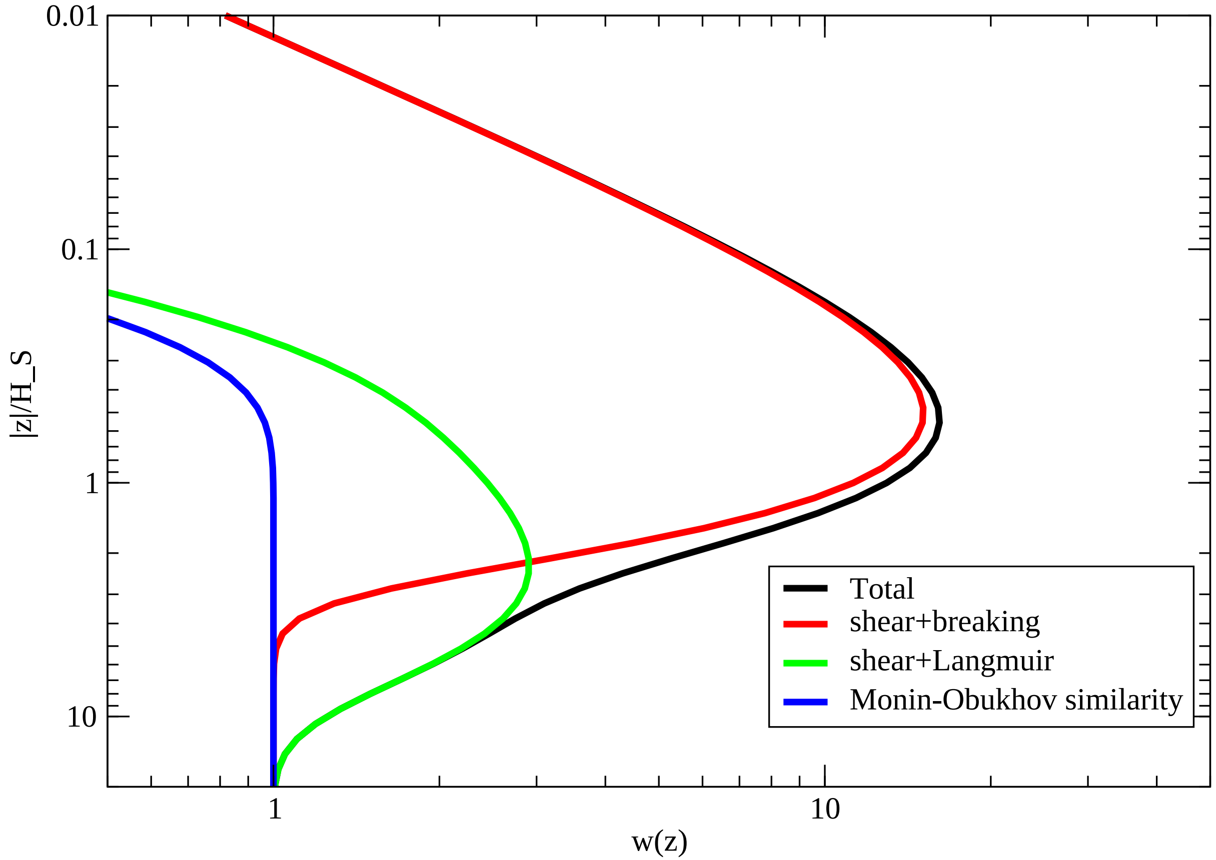


Figure 3: Profile of $w = Q^3$ according to the local approximation in the ocean column near the surface. The contributions by wave dissipation (red line) and Langmuir turbulence (green line) are shown as well. Finally, the w-profile according to Monin-Obukhov similarity, which is basically the balance between shear production and dissipation, is shown as the blue line.

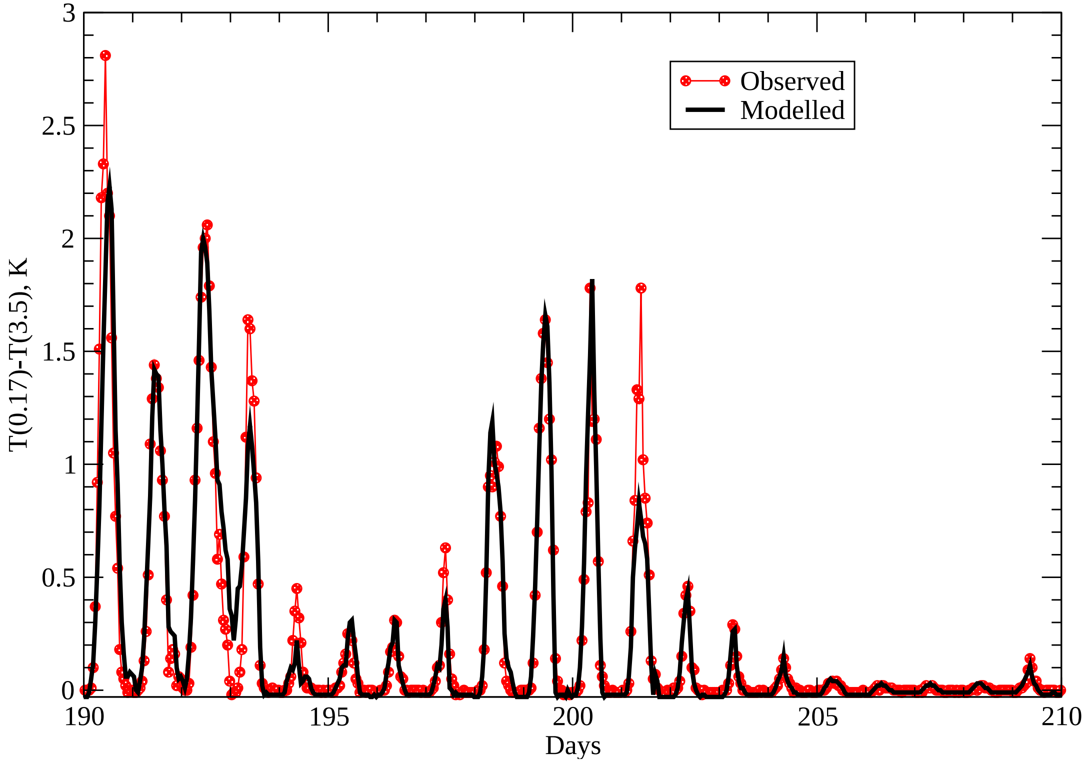


Figure 4: Observed and simulated ocean temperature $\Delta T = T(0.17) - T(3.5)$ at 15^{o}30' N, 61^{o}30' E in the Arabian Sea for 20 days from the 23rd of April.

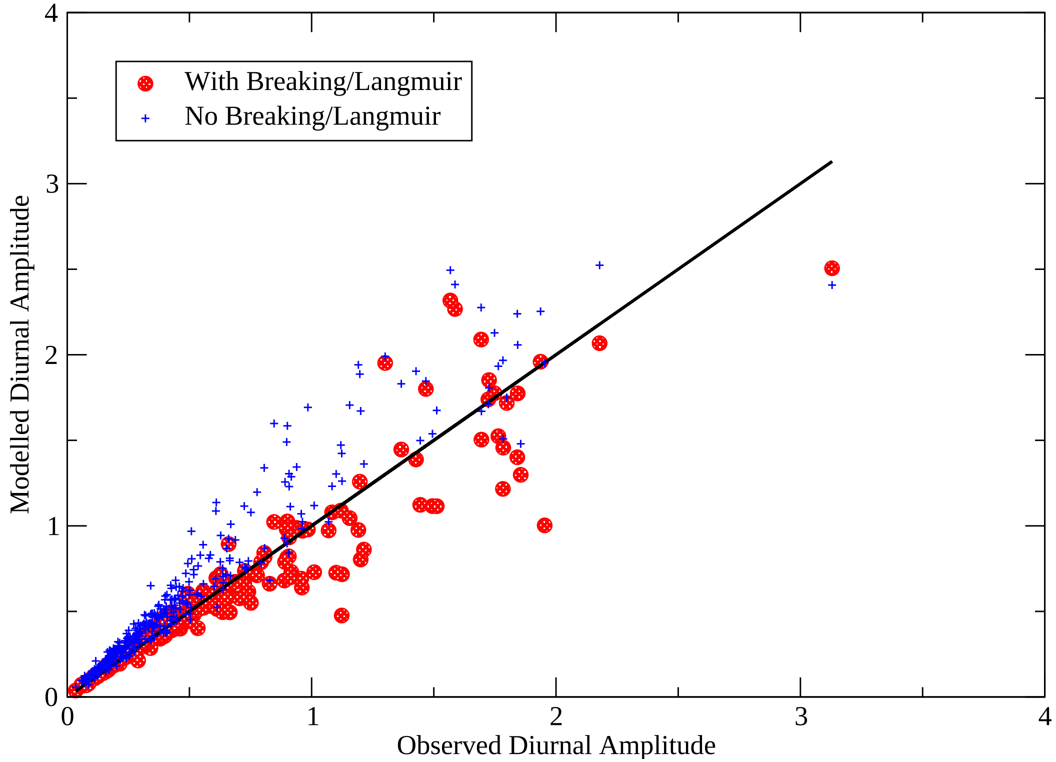


Figure 5: Comparison of simulated and observed diurnal amplitude at 15^o30' N, 61^o30' E in the Arabian Sea for the one year period starting from 16th of October 1994.

NEW RESULTS

The insights gained during the diurnal cycle work have been used in further developing the Nemo ocean model. The ocean model is forced by the momentum flux to the ocean column, while the mixing due to wave breaking now explicitely depends on the dissipation produced by the WAM model. The Stokes drift is explicitely determined as well. This allows for explicitely taking into account the effects of Langmuir turbulence and the Stokes-Coriolis force.

Specifically the momentum flux to the ocean is given by

$$\tau_{oc} = \tau_a - \rho_w \int_0^{2\pi} \int_0^{\infty} d\omega d\theta \, \mathbf{k} \left(S_{in} + S_{diss} \right), \tag{1}$$

while the energy flux to the ocean reads

$$\Phi_{oc} = -\rho_w \int_0^{2\pi} \int_0^{\infty} d\omega d\theta \, \omega S_{diss}, \tag{2}$$

Monthly means of these quantities are shown in the next two figures.

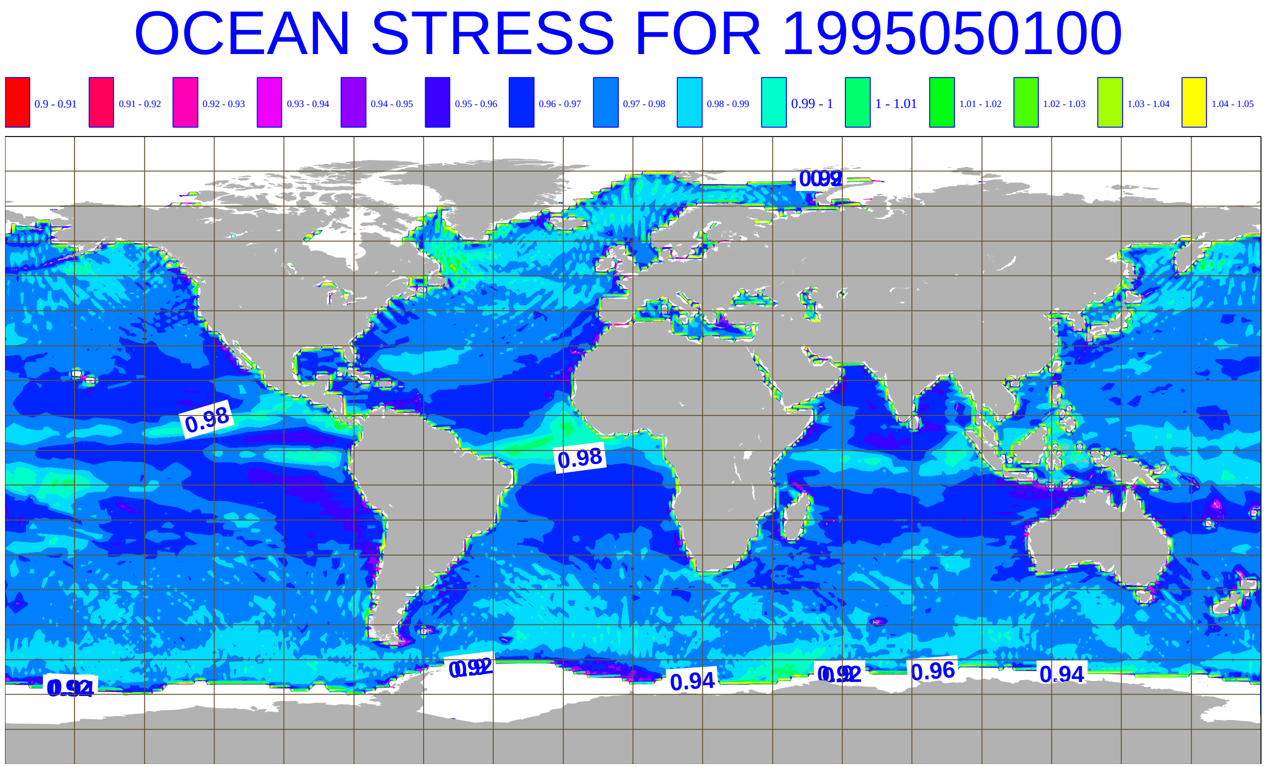


Figure 6: Monthly mean of momentum flux into the ocean, normalized with the monthly mean of the atmospheric stress.

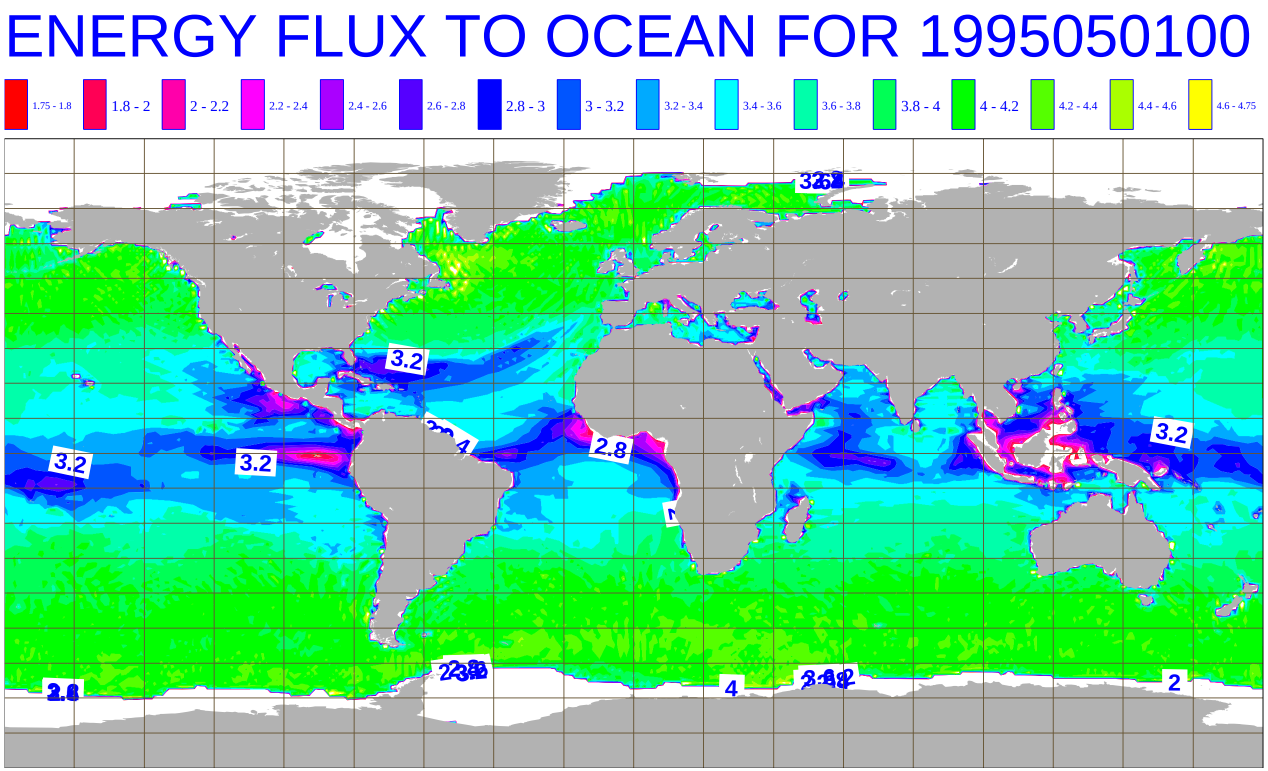


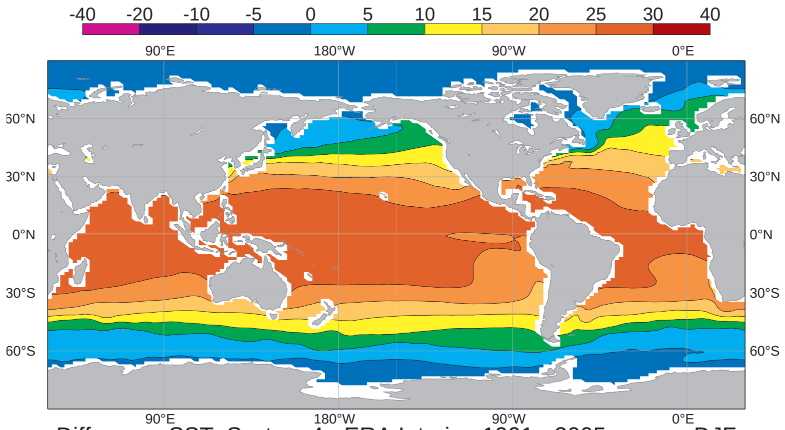
Figure 7: Monthly mean of energy flux into the ocean, normalized with the monthly mean of $\rho_a u_*^3$.

Systematic errors

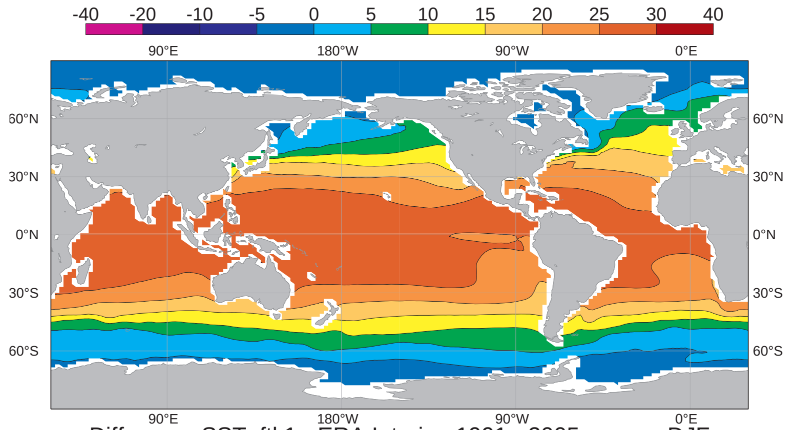
At the moment we are in the process of extensively testing the various new options we have introduced in the Nemo model. First results are now shown for the choice that the upper ocean mixing is provided explicitely by the wave model, by comparing coupled seasonal forecasts with and without the wave effect on mixing.

The period is 1991 until 2005 and starting dates are November generating 5 member ensembles. Shown is the average difference between modelled and observed SST over DJF.

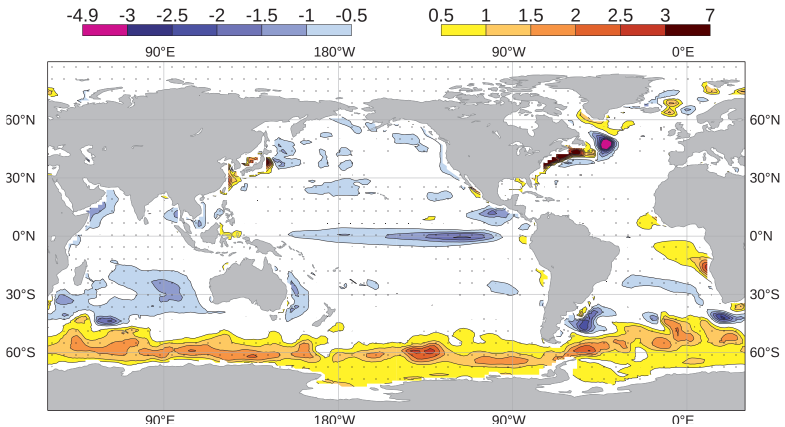
SST ERA Interim 1991 - 2005 season DJF



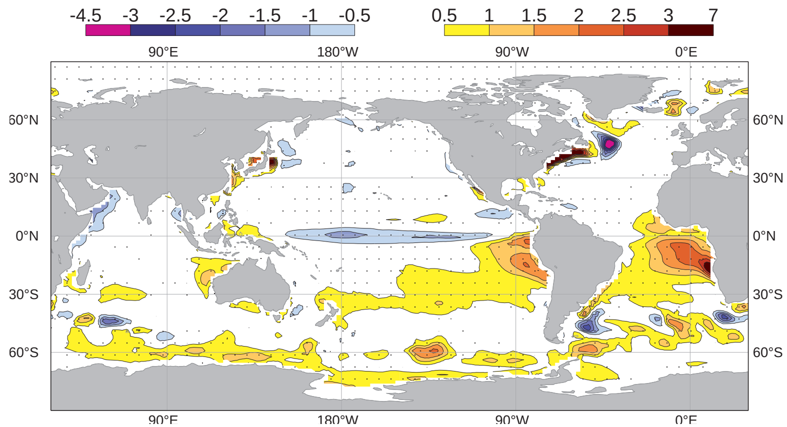
SST ERA Interim 1991 - 2005 season DJF



Difference SST, System 4 - ERA Interim, 1991 - 2005 season DJF
MAE: 0.47, MeanBias: -0.022, Dotted: 5 % significance



Difference SST, ftk1 - ERA Interim, 1991 - 2005 season DJF
MAE: 0.46, MeanBias: 0.2, Dotted: 5 % significance

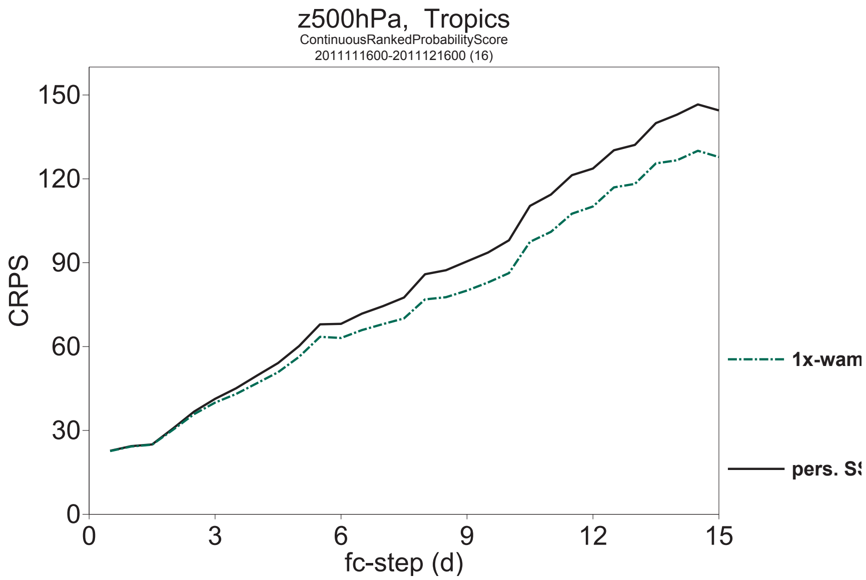


Ensemble Prediction

At the moment our ensemble prediction system is not based on a coupled model, in stead persisted SST anomalies are being used. The next plot shows what happens when a coupled system is used to produce the probability distribution of T850 and T200 in the Tropics. Results are for 16 cases with 50 members in the ensemble running a $T_l 639/1^\circ$model.

The quantity called CRPS measures the rms error in the modelled cumulative pdf against observed occurrence.

The impact is big and shows that coupling with the ocean is an advantage. We also tested the impact of the sea state dependent mixing. This gives about half of the size of the impact.


z500hPa, Tropics
ContinuousRankedProbabilityScore
2011111600-2011121600 (16)
CRPS
150
120
90
60
30
0
0
3
6
9
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15
fc-step (d)
1x-wam
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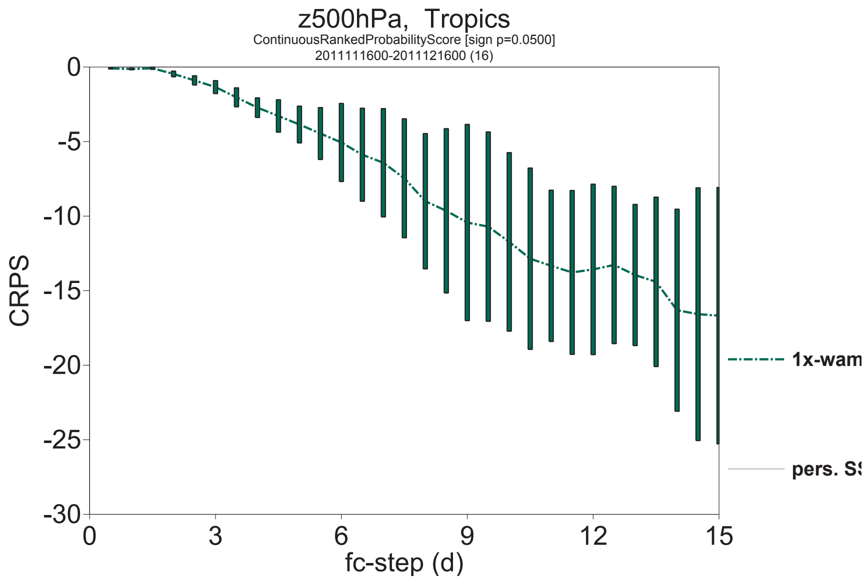
z500hPa, Tropics
ContinuousRankedProbabilityScore [sign p=0.0500]
2011111600-2011121600 (16)
CRPS
0
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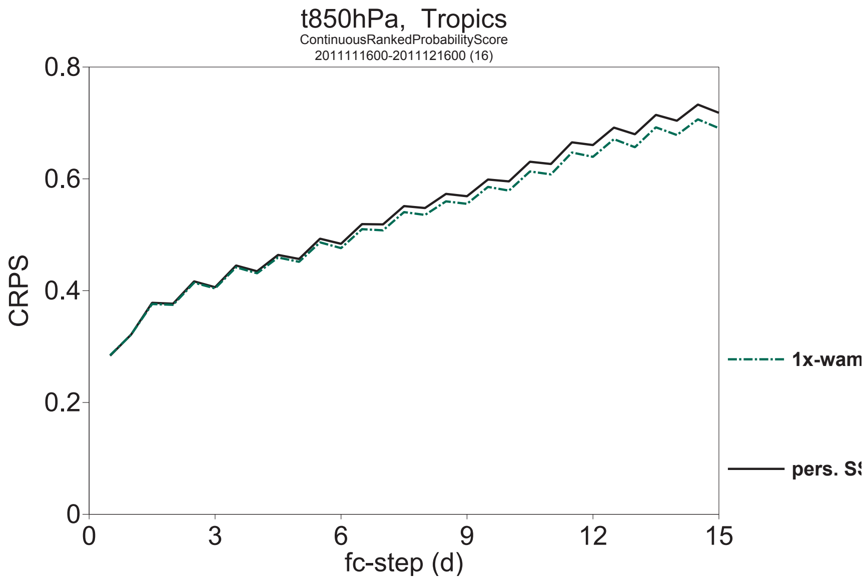
t850hPa, Tropics
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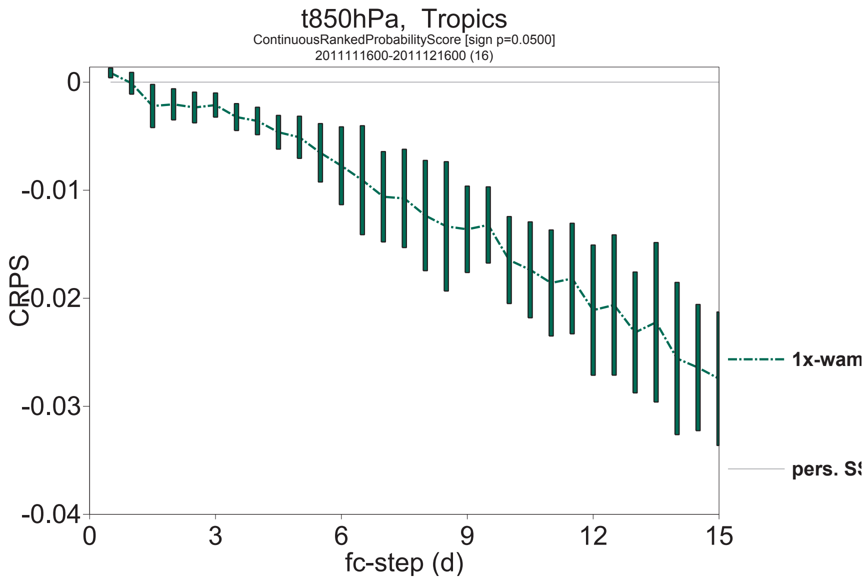
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ContinuousRankedProbabilityScore [sign p=0.0500]
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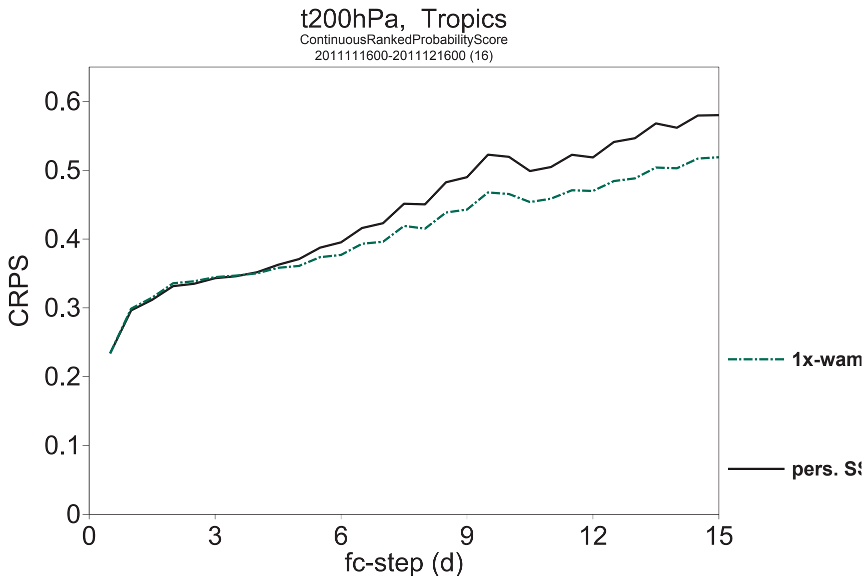
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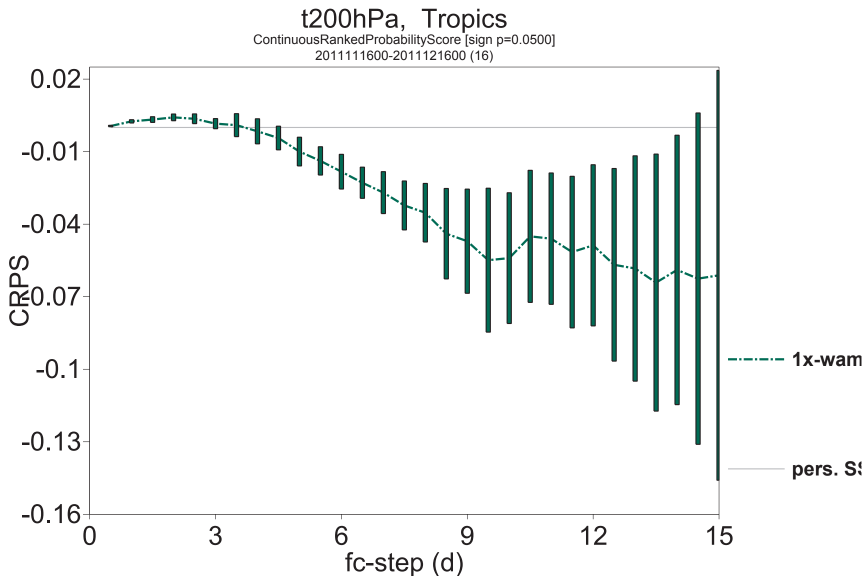
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ContinuousRankedProbabilityScore [sign p=0.0500]
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CRPS
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CONCLUSIONS

- For some time now there has been the expectation, based on physical considerations, that sea state effects should be relevant for upper ocean mixing.
- At ECMWF, we have developed an efficient tool to study these effects and first results are promising so that by the end of this year a coupled ensemble prediction system will be introduced operationally.
- Clearly, at the moment we are just at the beginning of a exciting new development.