Air-sea interaction and earth-system modelling

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INTRODUCTION

In this lecture I would like to discuss air-sea interaction in the context of an **earth-system model** (ESM). We are developing an ESM which at the moment consists of three components:

1. atmosphere (IFS)
2. ocean waves (WAM)
3. and the ocean/sea-ice (NEMO)

These three components have been brought together in a single executable. It will be argued, and hopefully become clear at the end, that at the interface of ocean and atmosphere, ocean waves play an important mediating role in the exchange of momentum and heat.

These lectures address the following topics:

- **Brief Introduction in Ocean Waves.**

  Since most people are not familiar with the subject of ocean waves I will give a brief introduction into this topic.

  Important quantity to predict is the wave spectrum. It gives the distribution of wave energy over frequency and direction. Waves, and in particular Ocean Waves, are governed by the energy balance equation. I will briefly discuss the physics of wind-wave generation, nonlinear interactions and dissipation.

  Wind-input and wave dissipation play, together with the wave-induced current (called the Stokes Drift), a key role in the air-sea interaction and hence in the coupling between ocean and atmosphere.

- **Numerics of the coupling**

  As occasionally there is a strong coupling between the three components of the EMS there is a need to study the numerical scheme involved in such a coupling.

  So far I am not aware of a systematic study of this problem. I will give an example of a serious problem we had in the early part of this Century, namely the generation of **mini vortices** by the coupling between wind and ocean waves, and how this was fixed.

- **Ocean waves and Upper-Ocean Mixing**

  Upper ocean mixing is to a large extent caused by breaking, ocean waves. As a consequence there is an energy flux Φ_{oc} from waves to ocean. It is given by

$$\Phi_{oc} = m\rho_a u_*^3,$$

  where m depends on the sea state and u_* is the friction velocity. Wave breaking and its associated mixing penetrates into the ocean at a scale of the significant wave height H_S. In addition, Langmuir turbulence penetrates deeper into the ocean with a scale of the typical wavelength of the surface waves.

  In the NEMO model there is a simple scheme to model these effects using the **turbulent kinetic energy** (TKE) equation. However, in the present version of NEMO there are only averaged sea state effects included, hence m is constant. Here, it is shown that when actual sea state effects are included this may have an impact on the mean SST field and even on the temperature field to 400 m depth.

- **Coupling from Day 0**

  For some time now, in the operational medium-range/monthly ensemble forecasting system (ENS) the interaction between atmosphere and ocean was only switched on at Day 10 in the forecast. In the autumn of 2013 a new version of ENS/monthly has been introduced in operations where the coupling starts from Day 0. Also, sea state effects on upper ocean mixing and dynamics have been switched on.

  Coupling from Day 0 has beneficial impacts on hurricane forecasting, the MJO and the statistical properties of ENS.

- **Coupled data assimilation**

  If one develops a coupled system, then it is also prudent to do coupled data assimilation. Presently, we are developing a prototype weakly coupled data assimilation system and some first results will be presented.

Reading material

- ECMWF lecture notes on Wave Modelling
- ECMWF Technical Memorandum 712, P. Janssen *et al.*, Air-sea interaction and surface waves, November 2013.
- Dynamics and Modelling of Ocean Waves, G.J. Komen *et al.*, Cambridge University Press, 1996.
- The interaction of ocean waves and wind, P. Janssen, Cambridge University Press, 2004.

OCEAN WAVES

Wave forecasting took off as an important subject when knowledge of the sea state was required for numerous landing operations during the second World War, for D-Day for example.

From then onwards a rapid development because of improved weather forecasting (better surface winds), enormous increase in the number of observations (from buoys and satellites) and faster and bigger computers.

From the beginning it was clear, however, that the sea state was so complicated that only a prediction of the average sea state at a location of interest was possible. Example of parameters are the average wave height and the average period of the waves.

Nomenclature

Sea state is represented by the wave spectrum $F = F(\mathbf{k}, \mathbf{x}, t)$. Here $\mathbf{k}$ is the wavenumber vector, and $\mathbf{x}$ the position vector.

Spectrum is normalized in such a way that integral over the spectrum gives, apart from a factor $\rho_w g$, the wave energy:

$$\int d\mathbf{k} F(\mathbf{k}) = \frac{E}{\rho_w g}, \; E = \rho_w g \langle \eta^2 \rangle$$

where η denotes the surface elevation, which has the dimension of length. Thus, spectrum is normalized so that it equals the wave variance $\langle \eta^2 \rangle$.

In order establish the connection with the common practice to think in terms of wave heights (for a single wave this the distance between crest and trough), we use the concept of significant wave height. This is a statistical measure, defined as

$$H_S = 4\sqrt{\langle \eta^2 \rangle}$$

In a similar vein, lots of other variables may be defined using the spectrum, e.g. mean frequency, mean wave direction, steepness of the waves also called the wave slope.

For given wind (and bathymetry etc.) a wave model calculates at a location of interest the evolution in time of the two-dimensional wave spectrum $F = F(\mathbf{k}, \mathbf{x}, t)$. Ocean waves satisfy a **dispersion relation** given by

$$\Omega = \sigma(k, D) + \mathbf{k} \cdot \mathbf{U}, \; \sigma = \sqrt{gk \tanh(kD)}$$

with D the local depth, $\mathbf{U}$ the surface current and g acceleration of gravity.

Then, the evolution equation for F, called the energy balance equation, is given by

$$\frac{\partial}{\partial t} F + \frac{\partial}{\partial \mathbf{x}} \cdot (\dot{\mathbf{x}} F) + \frac{\partial}{\partial \mathbf{k}} \cdot (\dot{\mathbf{k}} F) = S = S_{in} + S_{nl} + S_{diss}$$

where $\dot{\mathbf{x}} = \partial \Omega / \partial \mathbf{k}$ is the group velocity, and $\dot{\mathbf{k}} = -\partial \Omega / \partial \mathbf{x}$ gives the refraction of ocean waves, caused e.g. by gradients in the current and variable depth.

The source functions describe the generation of waves by wind (S_{in}), the dissipation of ocean waves by e.g. wave breaking (S_{diss}) and the energy/momentum conserving resonant four-wave interactions (S_{nl}).

The source functions

A lot of work in the past 50 years has been devoted to the formulation of the source functions:

- **Nonlinear transfer**

  The nonlinear interactions describe the energy transfer between the different wave components in the spectrum. Energy transfer takes place between 4 different waves that satisfy the resonance conditions

$$\Omega_1 + \Omega_2 = \Omega_3 + \Omega_4, \ \mathbf{k}_1 + \mathbf{k}_2 = \mathbf{k}_3 + \mathbf{k}_4.$$

  Nonlinear interactions play an important role in shaping the spectrum. Its representation is known exactly, but is operationally very expensive, therefore an approximation, called the Direct Interaction Approximation (DIA) is used.

- **Wind Input**

The wind input source function S_{in}, which represents the **interaction between wind and waves**, depends on the surface stress $\tau = u_*^2$ and is proportional to the wave spectrum! Hence, $S_{in} = S_{in}(F, u_*/c)$.The **strength of the interaction** is given by the wave-induced stress

$$\tau_w = \int \mathrm{d}\mathbf{k}\, S_{in}/c(k).$$

with $c = \Omega/k$ the phase speed of the waves. Note that there is only wind input when the wave speed is smaller then the wind velocity.

The wave growth by wind implies that there is momentum transfer from the air to the water waves. As a consequence, the airflow slows down. In the ECMWF system this slowing down is expressed as a sea state dependent roughness length $z_0 = \alpha\tau/g$, where the Charnock parameter α depends on the sea state, i.e.

$$\alpha = \frac{const}{\sqrt{1 - \tau_w/\tau}}.$$

Strong interaction when τ_w is close to τ!

- **Dissipation**

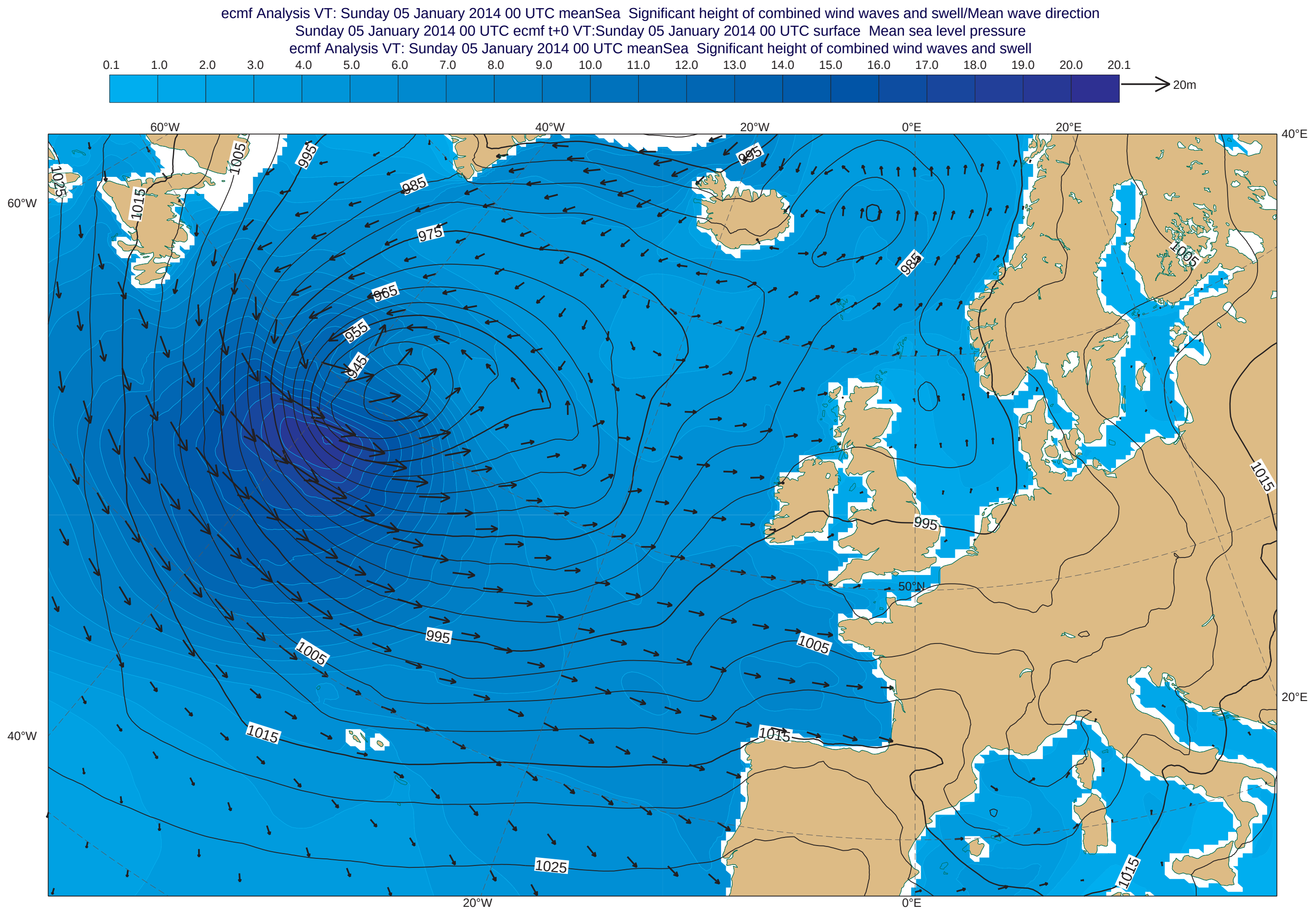
The dissipation source function describes the reduction of wave momentum and energy caused by processes such as white capping/ wave breaking, the damping by water turbulence, etc.

It is modelled in such a way that steep waves are more damped than gentle waves, as steep waves are more likely to break and or have whitecaps. The breaking waves will give rise to turbulence in the water, resulting in additional mixing of momentum and heat in the upper part of the ocean. The upper ocean mixing is controlled by the energy flux due to wave breaking, i.e.

$$\Phi_{oc} = -\rho_w g \int \mathrm{d}\mathbf{k}\, S_{diss} = m \rho_a u_*^3. \tag{1}$$

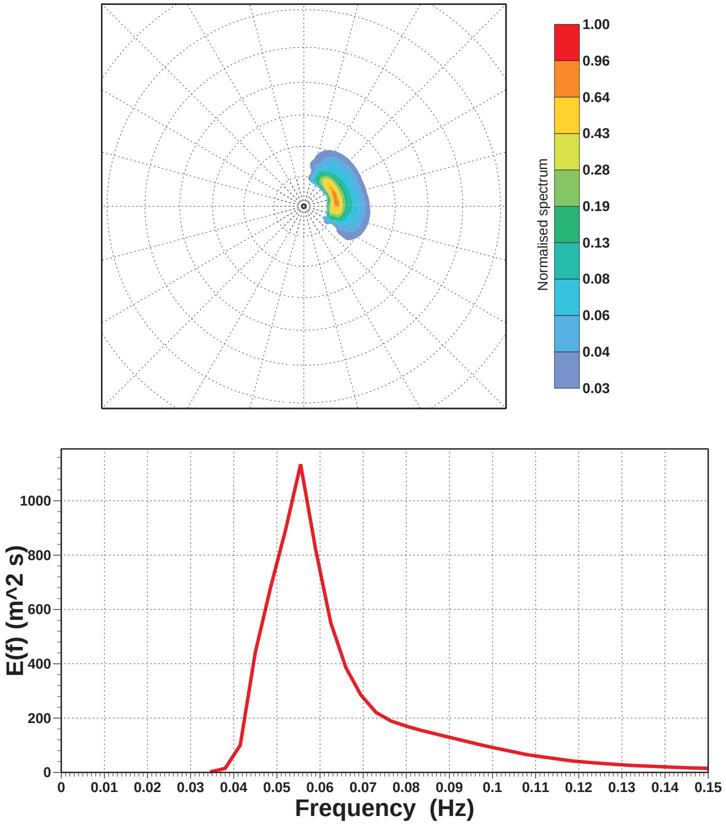
This will be discussed later when treating the coupling between WAM and NEMO.

An example of an extreme sea state

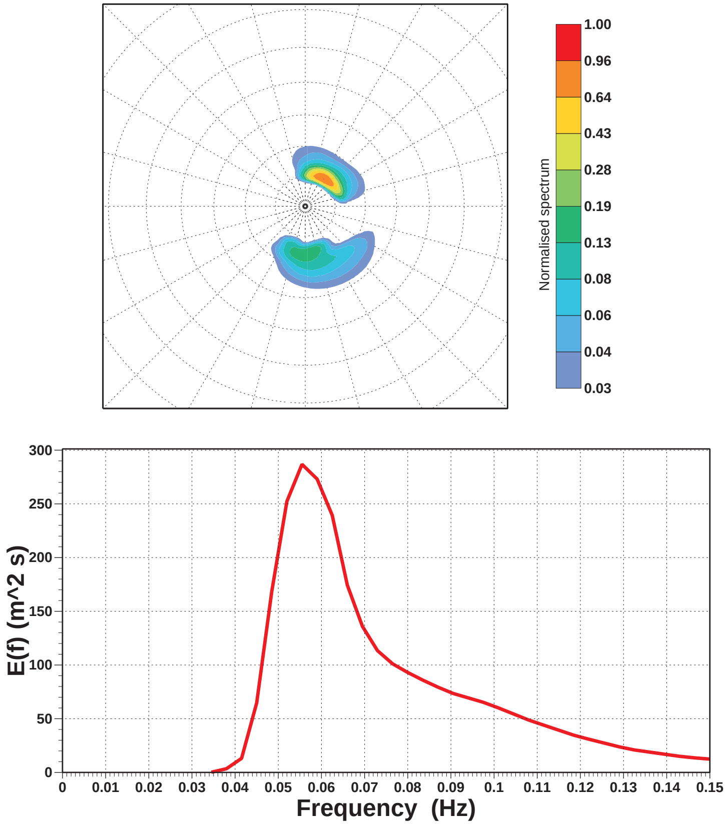


Spectra

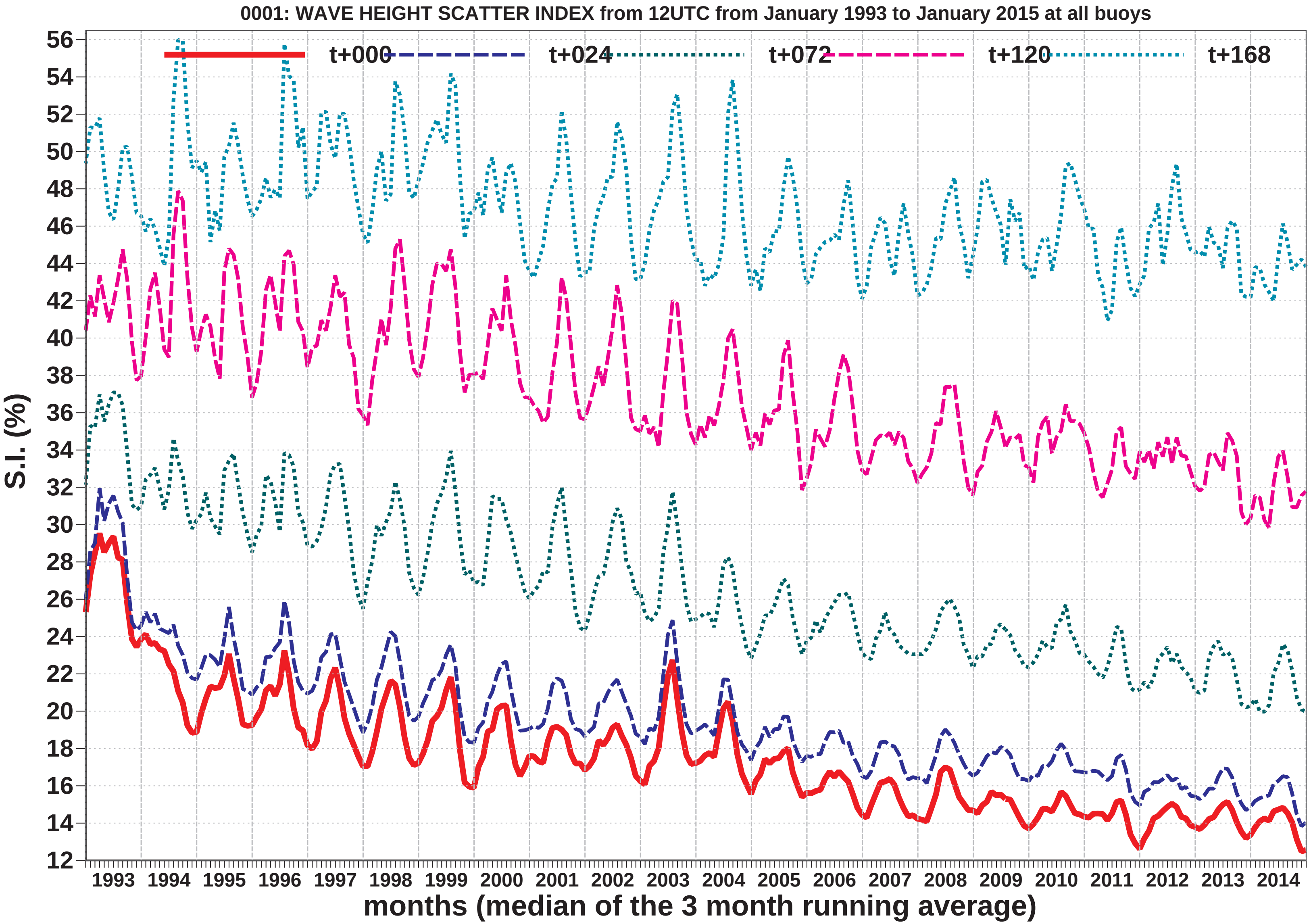
NORMALISED 2-D SPECTRUM for 0001 wave od
00:00Z on 05.01.2014
at xxxxx (47.75 , -38.50)
Hs= 20.14 m, Tm= 16.21 s, Tp= 17.99 s
Peakedness Qp = 1.79, Directional Spread = 0.50
MWD = 67 degrees PWD = 80 degrees
Propagation direction is with respect to North
North is pointing upwards
Concentric circles are every 0.05 Hz



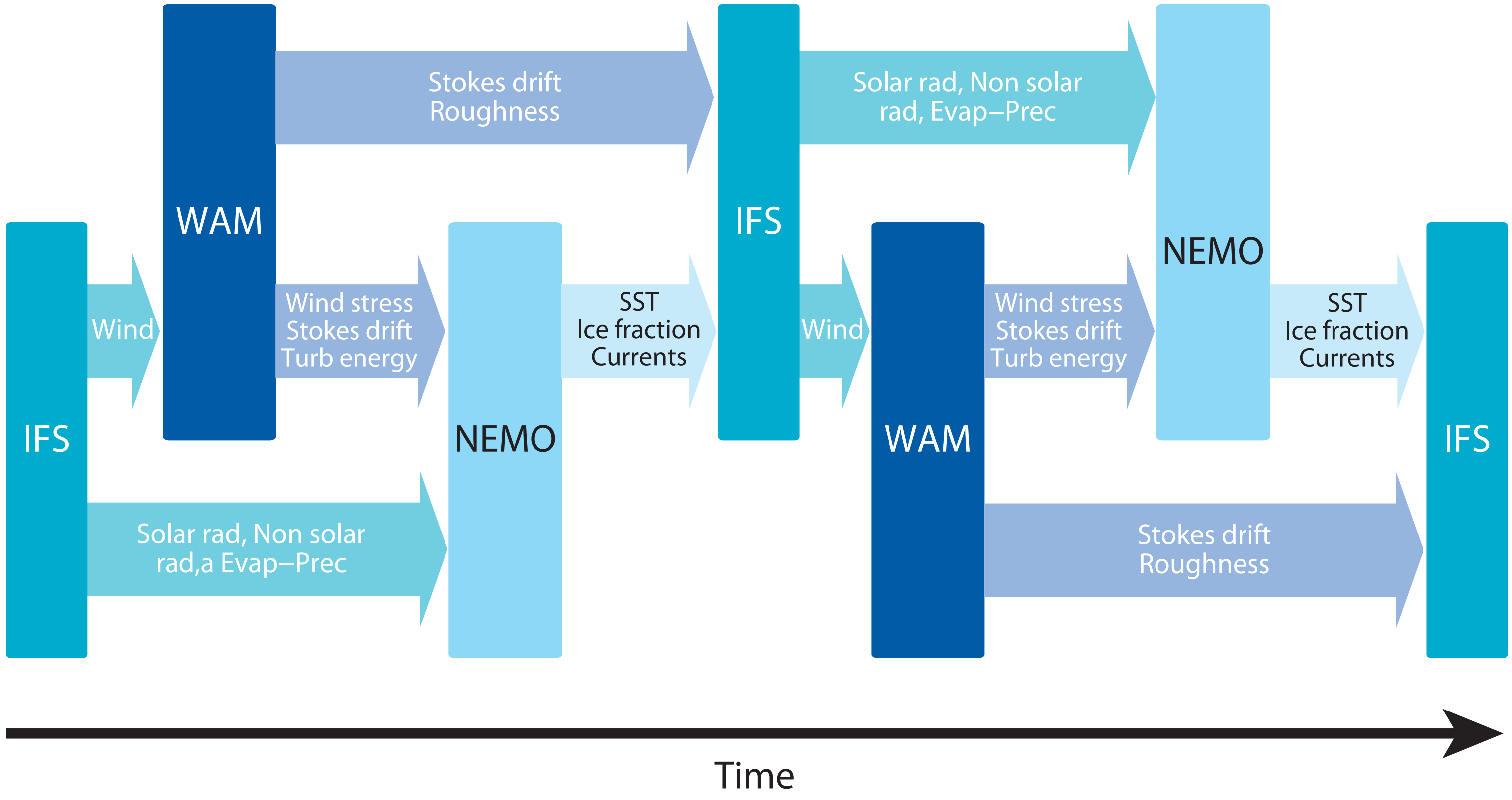
NORMALISED 2-D SPECTRUM for 0001 wave od
00:00Z on 05.01.2014
at aaaaa (50.00 , -42.00)
Hs= 12.76 m, Tm= 14.26 s, Tp= 17.99 s
Peakedness Qp = 1.26, Directional Spread = 0.81
MWD = 59 degrees PWD = 30 degrees
Propagation direction is with respect to North
North is pointing upwards
Concentric circles are every 0.05 Hz



Verification against buoy data: Some progress



THE COUPLED MODELLING SYSTEM



Flow Chart of the coupled model, here two time steps are shown.

Numerics of the coupling

I will give a few examples showing that occasionally there is a strong coupling between the three components of the ESM. One example involves the deepening of a low during IOP17 of FASTEX (atm=ocean-wave).

Another example is from our ocean-atmosphere ensemble system and involves hurricane Nadine and the cooling of SST by the strong wind circulation.

The final example is from our coupled deterministic system, which is presently under development. It involves typhoon Neoguri and a verification of the cooling effect.

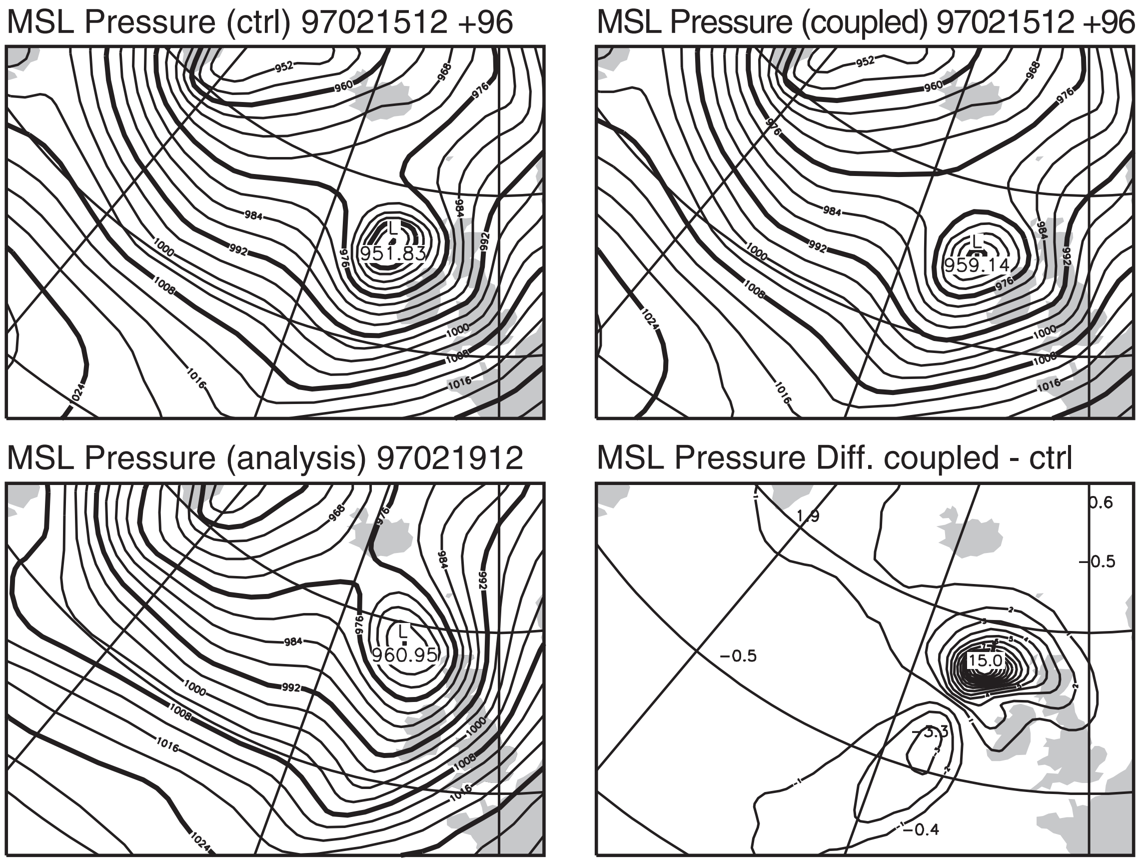


Figure 1: Comparison of 4-day forecast of surface pressure over the North Atlantic, valid for 19 February 1997. Version of coupled model is $T213/L31-0.5\,\mathrm{deg}$.

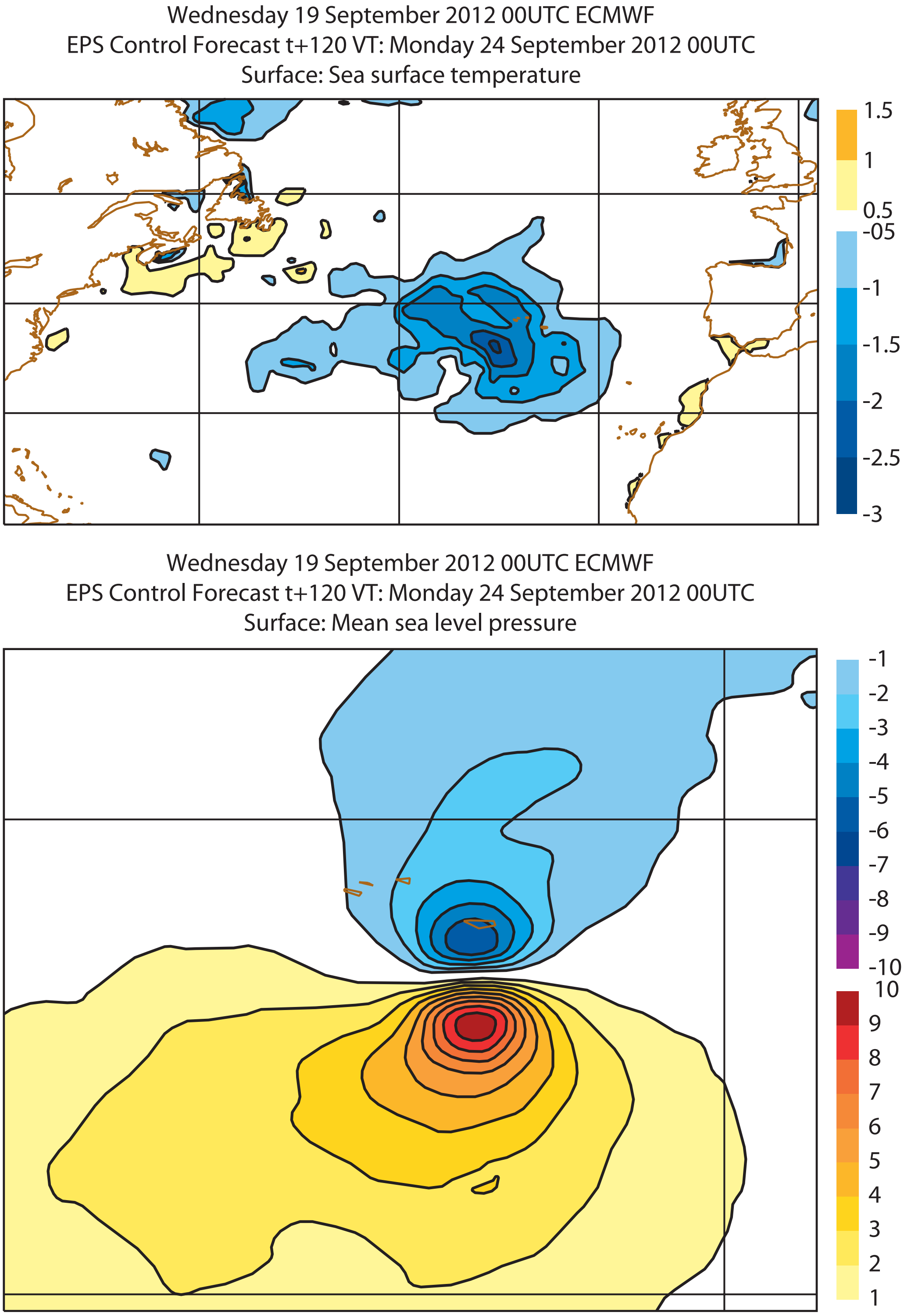


Figure 2: NADINE. Top: ensemble mean sst difference day5-day0. Bottom: ensemble mean pressure difference between coupled and control for day 5 forecast.

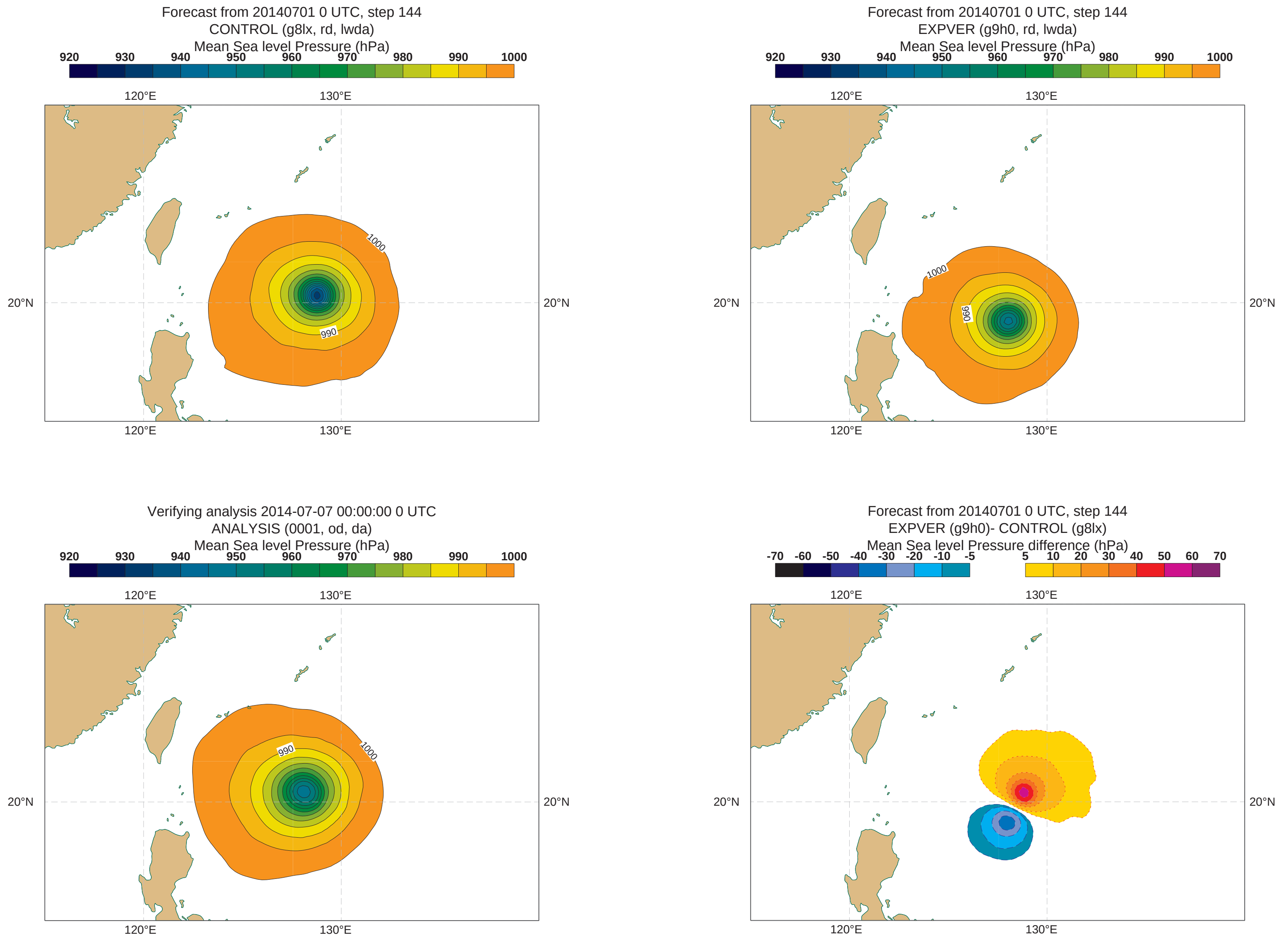
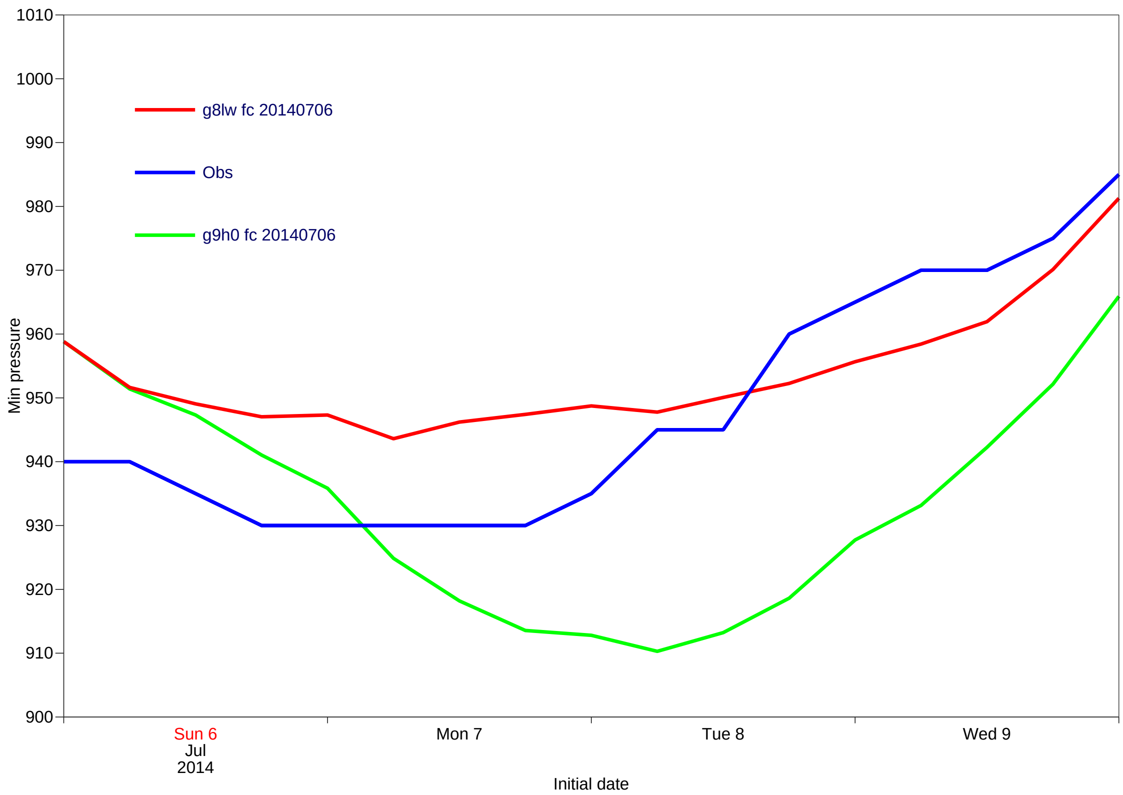
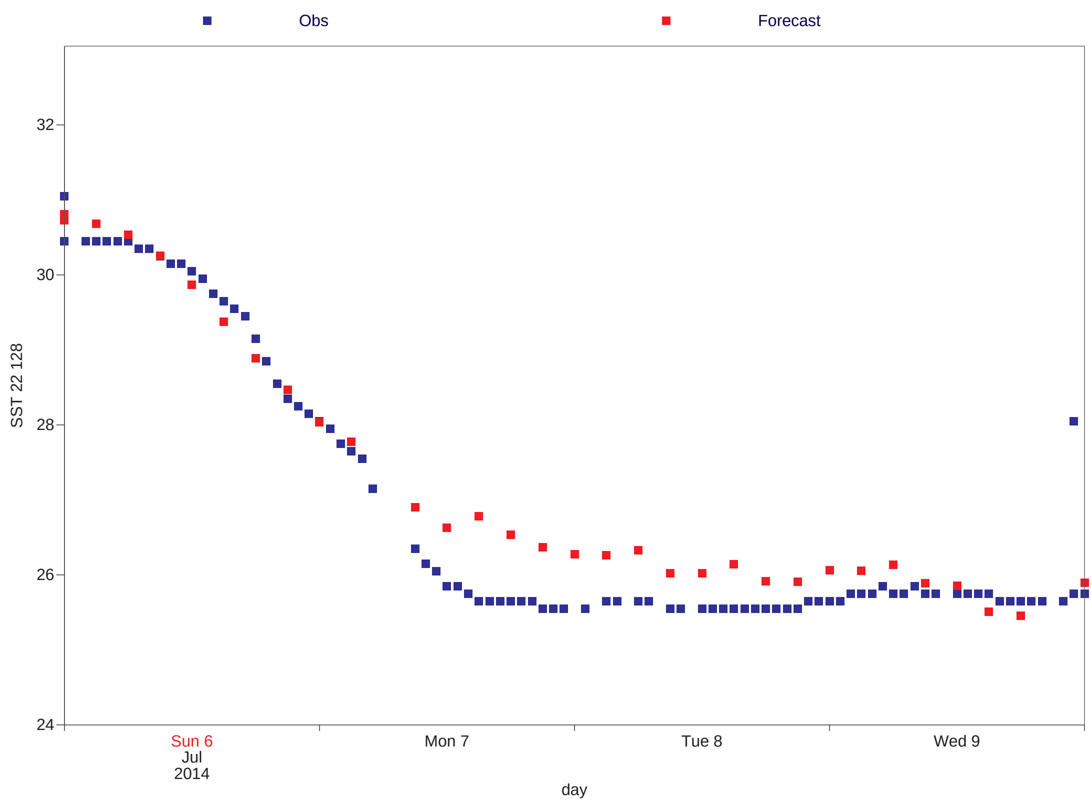


Figure 3: Neoguri prediction at day 6.


1010
1000
990
980
970
960
950
940
930
920
910
900
Min pressure
g8lw fc 20140706
Obs
g9h0 fc 20140706
Sun 6
Jul
2014
Mon 7
Tue 8
Wed 9
Initial date


Obs
Forecast
32
30
28
26
24
SST 22 128
Sun 6
Jul
2014
Mon 7
Tue 8
Wed 9
day

Because of this occasional strong interaction between atmosphere and ocean waves there is a need to study the numerical scheme involved in such a coupling. For example, if the coupling is strong are there possibilities of numerical instability, is there therefore a need to couple in an implicit manner, etc.

This might require a systematic study. I will give one example, namely the generation of spurious mini-vortices caused by the coupling between wind and ocean waves.

Generation of spurious mini-vortices

1. Two-way interaction of wind and waves was introduced on June 29 1998. The coupling time step was 4 wave model time steps, hence ample time for the wave model to respond to rapidly varying winds, resulting in realistic values of the roughness length.

2. With the introduction of the T_L511 version of the IFS the coupling time step was reduced to one wave model time step. From the start of the operational introduction occasional small scale, compact features occurred in the surface pressure field that propagated rapidly over the oceans. Called mini-vortices, or even **cannon balls**.

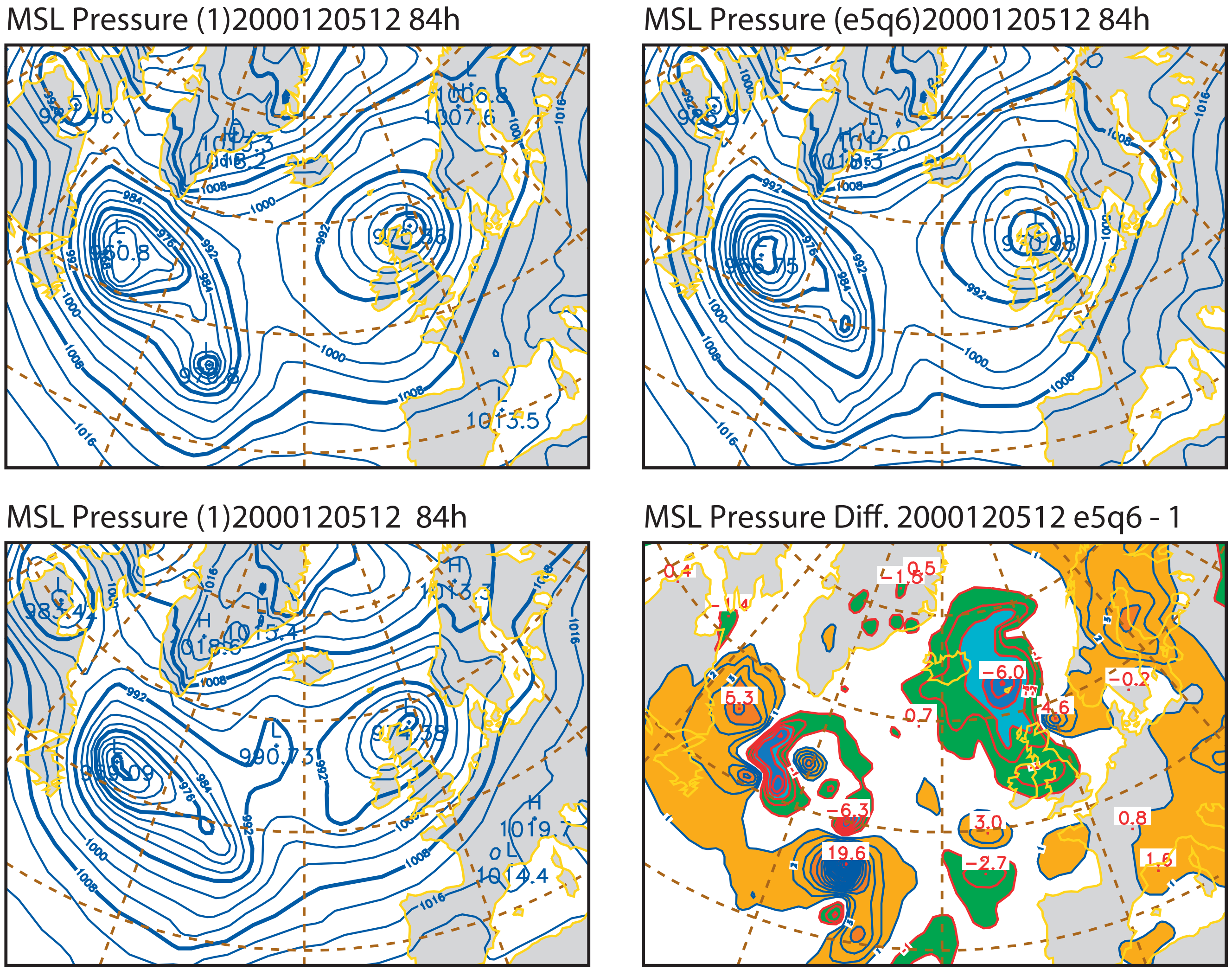


Figure 4: Generation of mini vortices by wind-wave interaction. Top left OPER, Top right EXP, Bottom left ANALYSIS, Bottom right diff between EXP and OPER.

Ocean waves

The integration in time was done with a (semi-) implicit scheme as follows.

1. Calculate dimensionless roughness or the Charnock parameter gz_0/u_*^2 from wave-induced stress at $t = t_n$ and wind speed at new time level t_{n+1}. Calculate friction velocity u_*^{n+1}

2. Spectral increments ΔF are obtained from an implicit scheme:

$$\Delta F = \Delta t S_n(u_*^{n+1}) \left[1 - \Delta t \frac{\delta S_n}{\delta F}(u_*^{n+1}) \right]^{-1}$$

Problem is that under rapidly varying winds (e.g. sudden drop in wind) the waves are still steep given a far too large roughness. This results in considerably enhanced heat fluxes that may generate a mini vortex.

Fix: Do the roughness calculation also after the spectral update $F_{n+1} = F_n + \Delta F$.

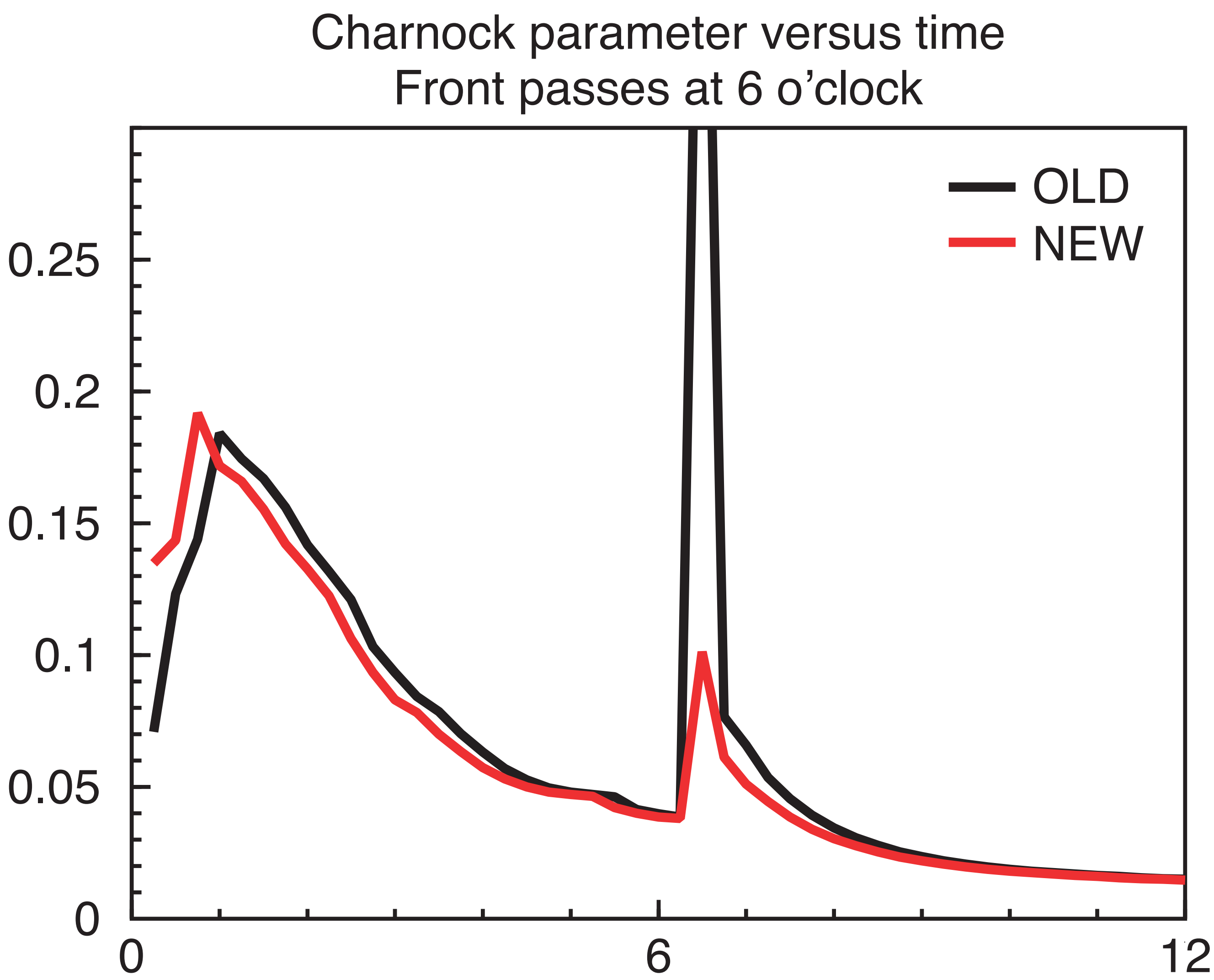


Figure 5: Evolution in time of the Charnock parameter during the passage of a frontal system at $t = 6$ hrs.

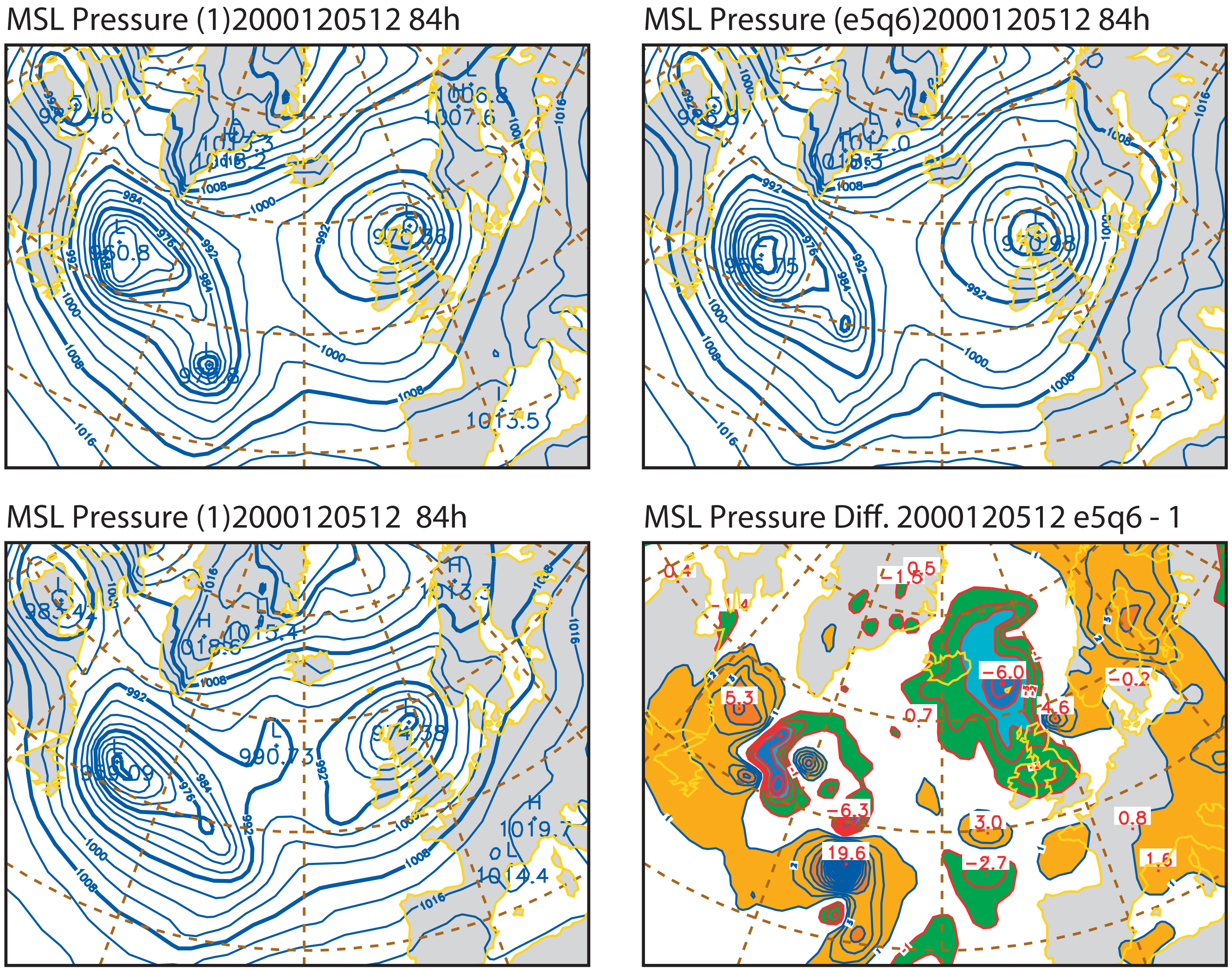
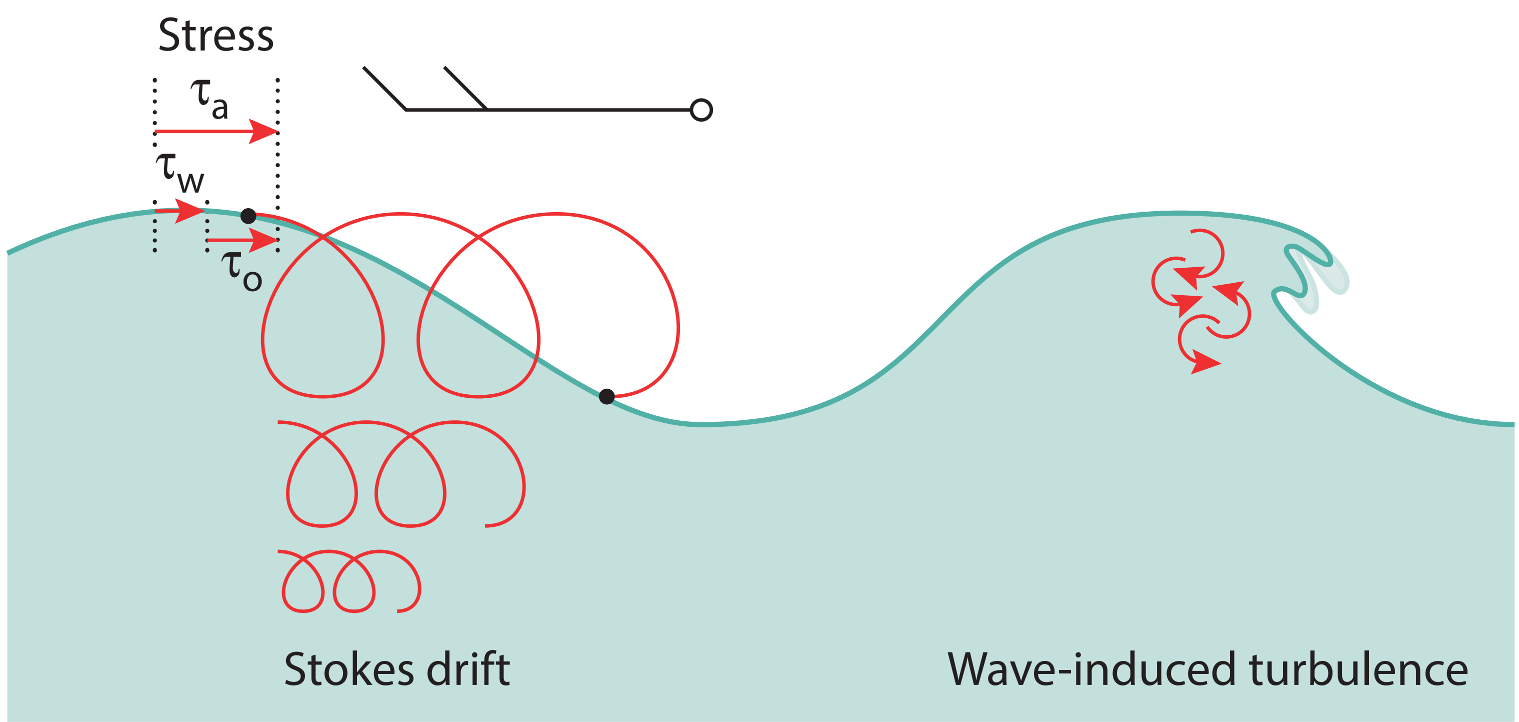


Figure 6: Generation of mini vortices by wind-wave interaction. Top left OPER, Top right EXP, Bottom left ANALYSIS, Bottom right diff between EXP and OPER.

COUPLING OF WAM AND NEMO



WAVE BREAKING and UPPER OCEAN MIXING

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

Wave breaking generates turbulence near the surface, in a layer of the order of the wave height H_S, which enhances the turbulent velocity 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'.

The turbulence modelling is based on an extension of the **Mellor-Yamada scheme** with sea state effects. Here, the turbulence is enhanced by means of the energy flux from waves to ocean column which follows from the dissipation term in the energy balance equation:

$$\Phi_{oc} = -\rho_w g \int \mathrm{d}\mathbf{k}\, S_{diss} = m\rho_a u_*^3.$$

and in general m is not a constant, as shown next.

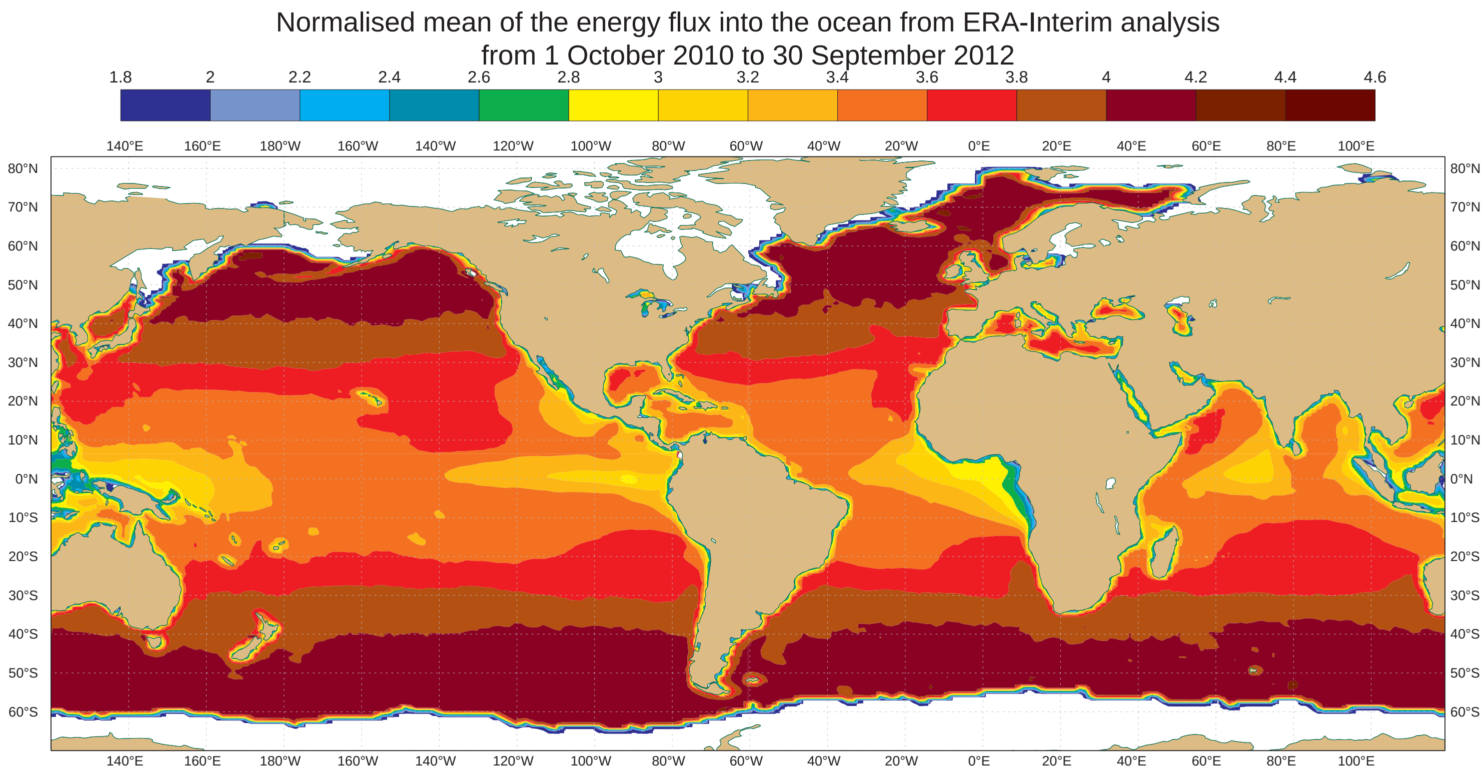


Figure 7: Mean of energy flux into the ocean, normalized with the mean of $\rho_a u_*^3$. Averaging period is two years.

TKE EQUATION

If effects of advection are ignored, the turbulent kinetic energy (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 TKE, and turbulent dissipation ε. It reads

$$\frac{\partial e}{\partial t} = \frac{\partial}{\partial z}\left(\nu_q \frac{\partial e}{\partial z}\right) + \nu_m S^2 + \nu_m S \frac{\partial U_S}{\partial z} - \nu_h N^2 - \varepsilon,$$

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. The eddy viscosities for momentum, heat, and TKE are denoted by ν_m, ν_h and ν_q. E.g $\nu_m = l(z)q(z)S_M$ where $l(z)$ is the mixing length and S_M depends on stratification.

Wave-induced turbulence is introduced by the boundary condition:

$$\rho_w \nu_q \frac{\partial e}{\partial z} = \Phi_{oc}, \quad z = 0.$$

while effects of Langmuir turbulence are introduced by the term involving the shear in the Stokes-drift profile.

In the next Figure we show an approximate solution to the TKE equation which illustrates that wave breaking enhances turbulence up to a depth of a few wave heights, while Langmuir turbulence acts in the deeper parts of the ocean. For comparison, results for Monin-Obukhov similarity (from balance of turbulent shear production and turbulent dissipation) are shown as well.

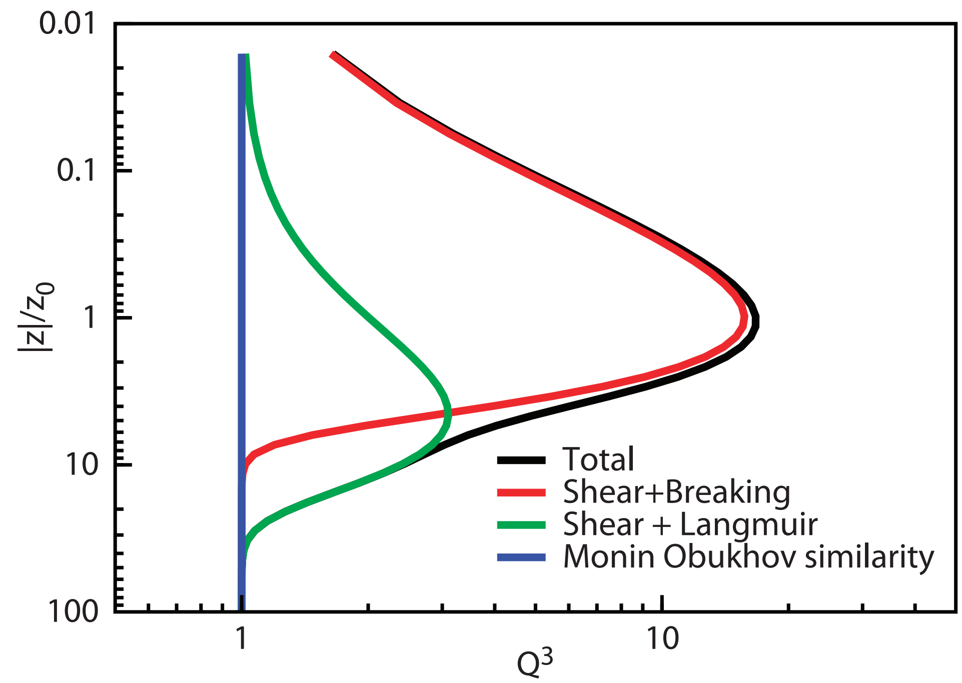


Figure 8: Profile of Q^3 in the ocean column near the surface, with Q a dimensionless turbulent velocity. The contributions by wave dissipation (red line) and Langmuir turbulence (green line) are shown as well. Finally, the profile according to Monin-Obukhov similarity, which is basically the balance between shear production and dissipation, is shown as the blue line and is constant because of the constant stress assumption.

The following Figure shows a comparison between the profile of modelled dissipation and a fit to observations of turbulence dissipation. The law-of-the wall follows from

$$\varepsilon = \nu_m S^2,$$

which for a constant stress, i.e. $\nu_m S =$ const, gives an inverse dependence on the distance from the surface.

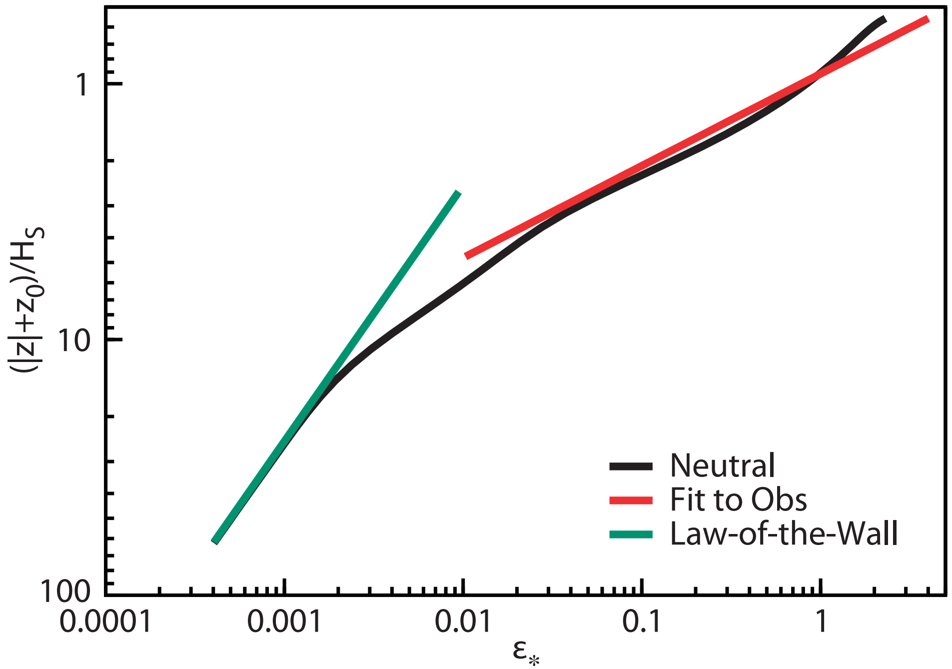


Figure 9: Dimensionless dissipation $\varepsilon_* = \varepsilon H_S/\Phi_{oc}$ versus $(z+z_0)/H_S$

IMPACT ON MEAN SST FIELD

We introduced the sea state dependent upper ocean mixing in the NEMO model.

The ocean circulation equations are very similar to the hydrostatic equations for the atmosphere, except, of course no clouds, but rather salinity.

A number of methods are used to advance these equations (discretized on an Arakawa C grid) in time. The non-diffusive parts are treated by a leap frog scheme, while for the diffusive parts a forward/backward time differencing scheme is used. By introducing a semi-implicit computation of the hydrostatic pressure gradient term the stability range of the leap frog scheme can be extended by a factor of two.

Show results from standalone runs, forced by ERA-interim fluxes and seastate. Averages are over a 20 year period. The control run is one where the dimensionless energy flux is a constant, given by $m = 3.5$.

The default version of NEMO already has sea state effects on upper ocean mixing. However, since no wave information is available the dimensionless energy flux m is a constant, given by $m = 3.5$. With a top-layer of 10m thick, this gives rise to too much mixing, as the layer where wave-induced mixing operates has only a thickness of the significant wave height $H_S \simeq 2.5m$. In the experiment we have corrected for this, thus giving more realistically but reduced amount of mixing.

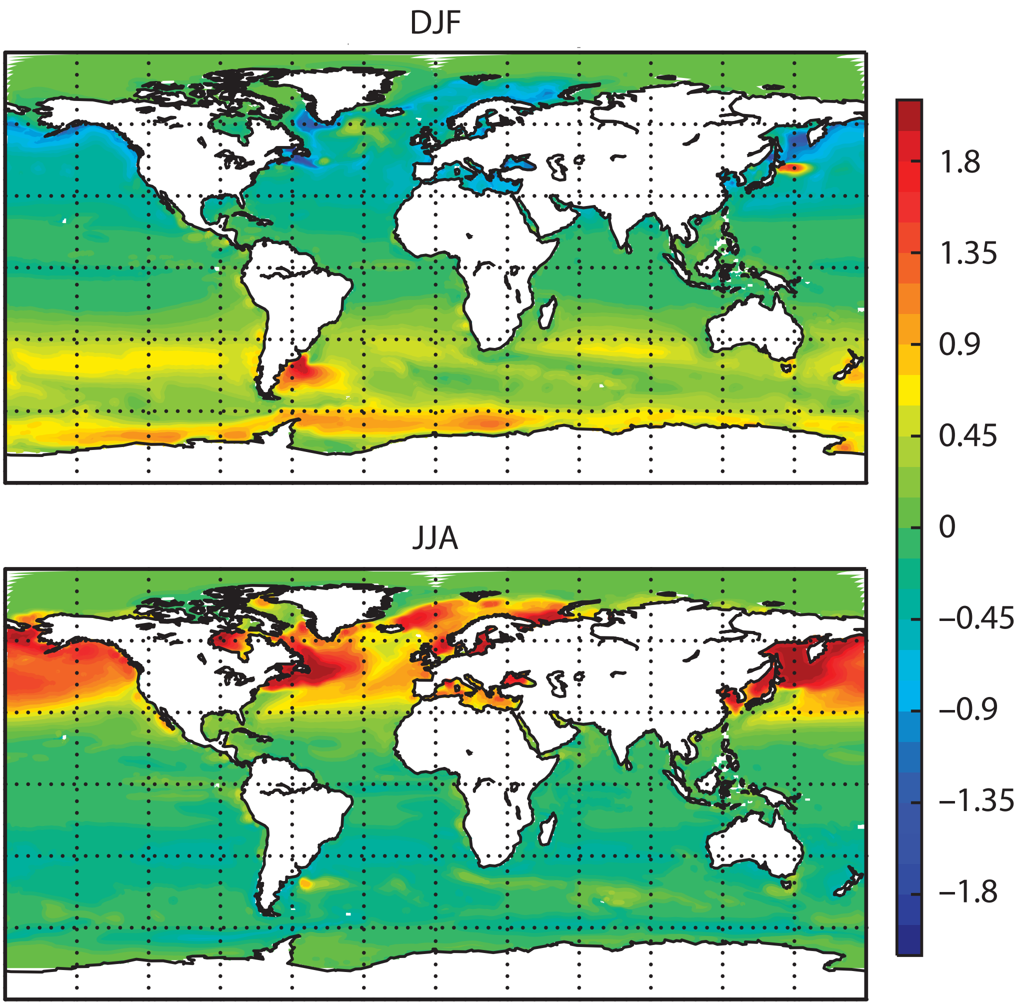


Figure 10: 20-yr average SST difference between CTRL (fsei) and a run with TKE mixing dictated by the energy flux from the ERA-Interim WAM model plus the other wave effects (fxl2). The differences are most pronounced in the summer hemisphere, where the mixing is reduced, leading to higher SST. The colour scale is ±2 K.

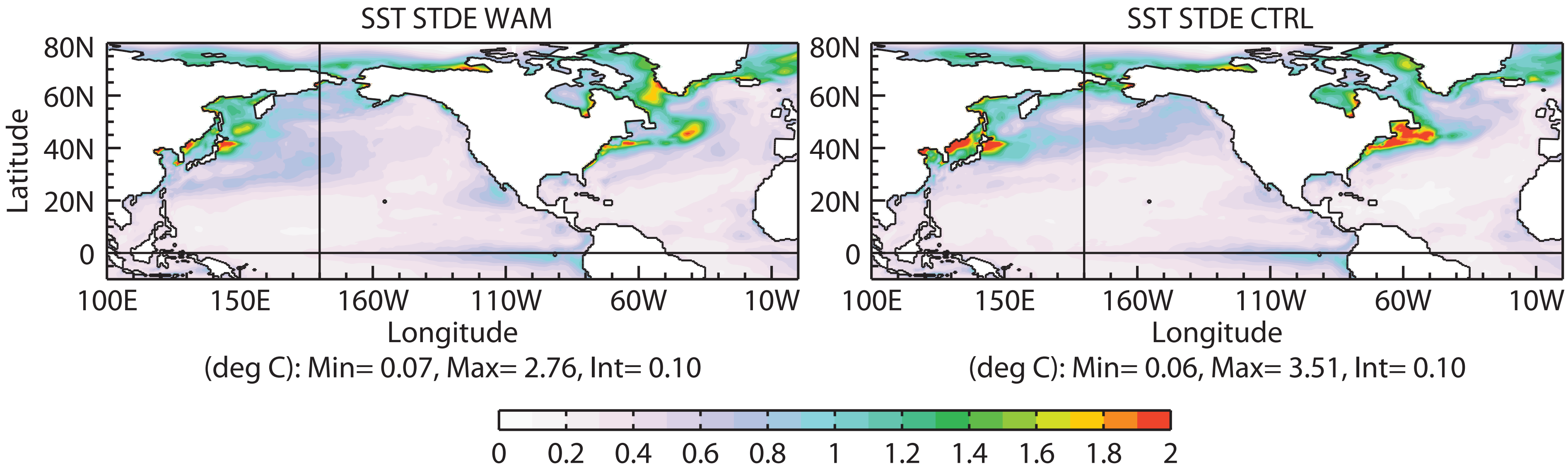


Figure 11: Standard deviation of errors in modelled SST, obtained from a comparison with OIv2 SST analyses.The left panel, labeled WAM, shows the STD errors when all sea state effects are switched on, while the right panel shows the STD errors obtained from CTRL.

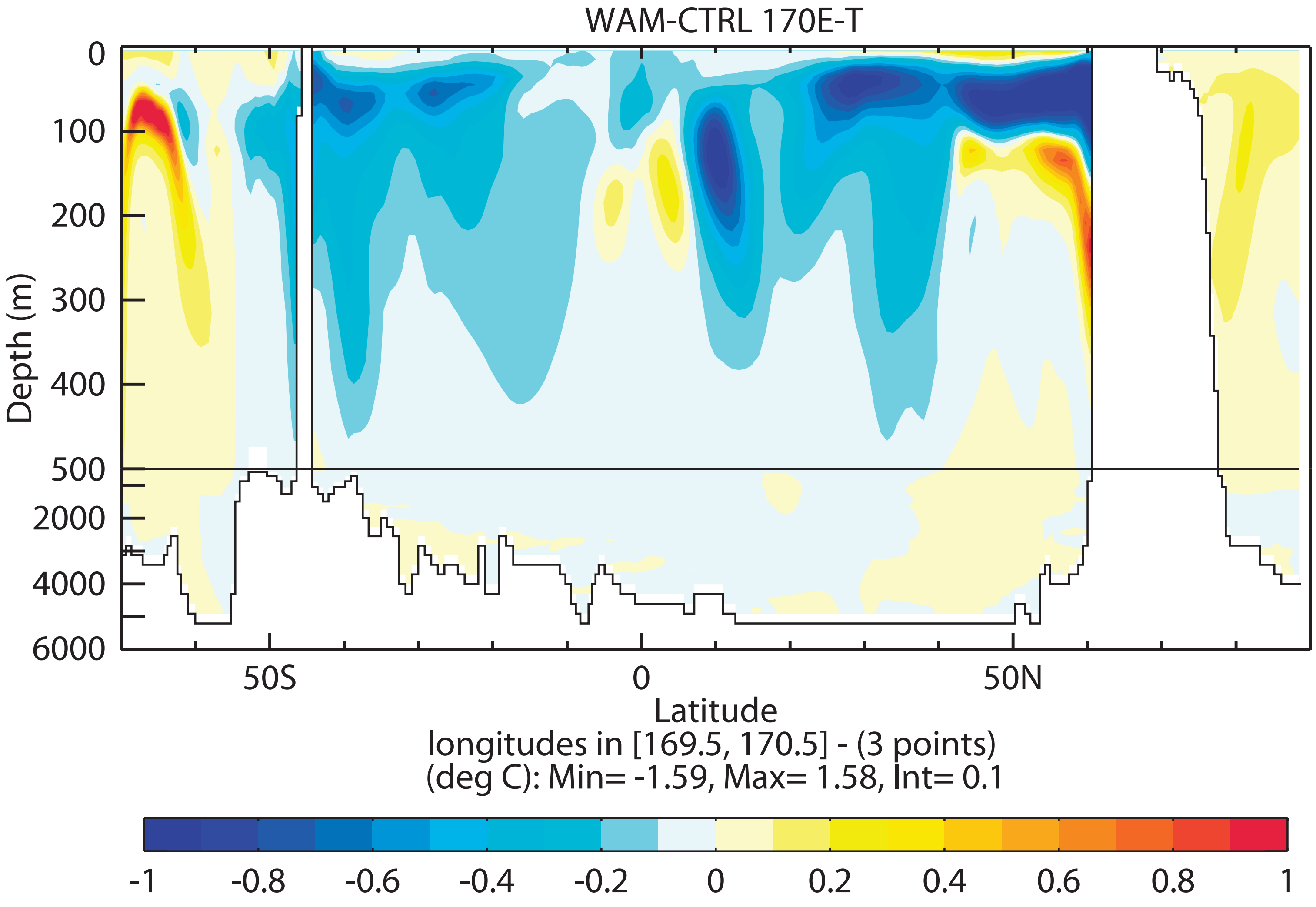


Figure 12: Panel a: Cross section at 170 E of the difference between temperature of WAM and CTRL. Impact of sea state dependent mixing is seen down to 400 m depth.

IMPACT ON COUPLED RUNS

Next, we study results from coupled seasonal forecast runs. Again, the control run is one where the dimensionless energy flux is a constant, given by $m = 3.5$.

As the control gave substantial biases, and seasonal forecasting skill is very sensitive to systematic errors we used an additional control run with a reduced value of m, $m = 0.56$, which had very similar systematic errors as the experiment with seastate effects included. Surprisingly, in certain areas this had a negative impact on forecasts skill.

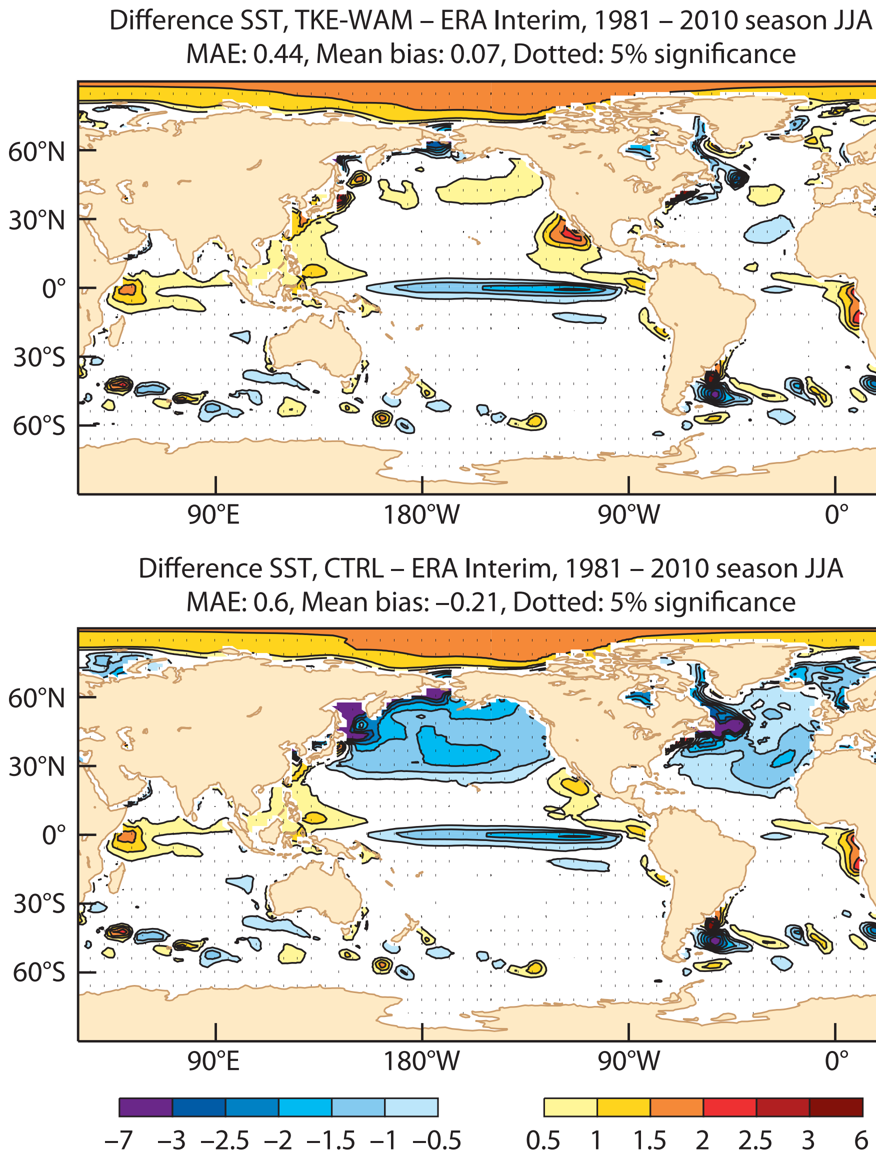


Figure 13: Systematic differences in SST with respect to the OIv2 analysis for JJA with start dates in May. Top panel the experiment TKE-WAM, bottom panel the CTRL experiment.

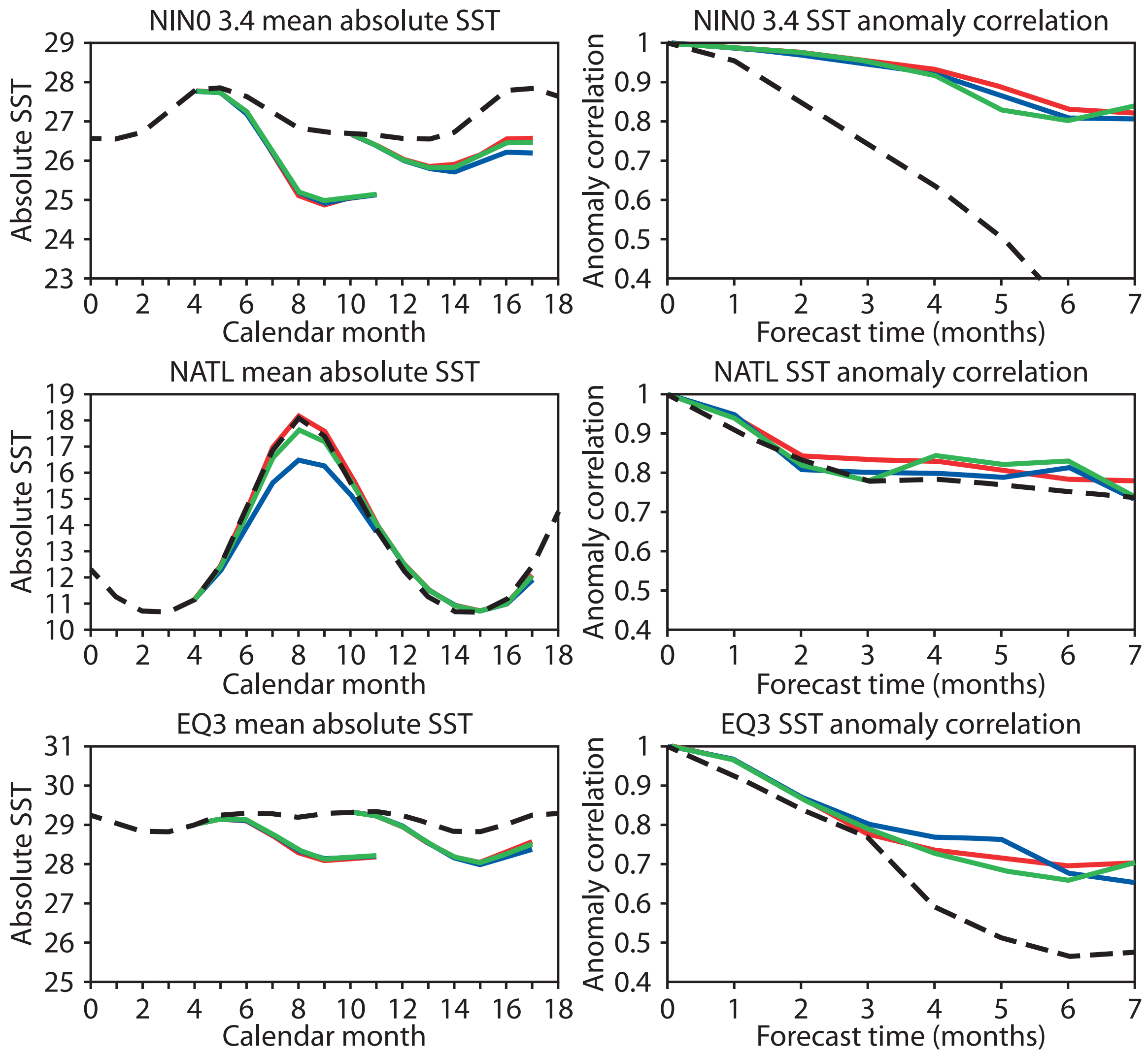


Figure 14: Absolute SST (left column) and correlation (right column) for the seasonal forecasts with CTRL ($m = 3.5$) (blue), TKE-WAM (red) and TKE-20 ($m = 0.56$, green) in selected areas.

COUPLING FROM DAY 0

For a long time, in the operational medium-range/monthly ensemble forecasting system (ENS) the interaction between atmosphere and ocean was only switched on at Day 10 in the forecast. In the Autumn of 2013 a new version of ENS has been introduced in operations where the coupling starts from Day 0. Also, sea state effects on upper ocean mixing and dynamics have been switched on.

Coupling from Day 0 has beneficial impacts on

- hurricane forecasting (already shown)
- the MJO
- and the statistical properties of ENS.

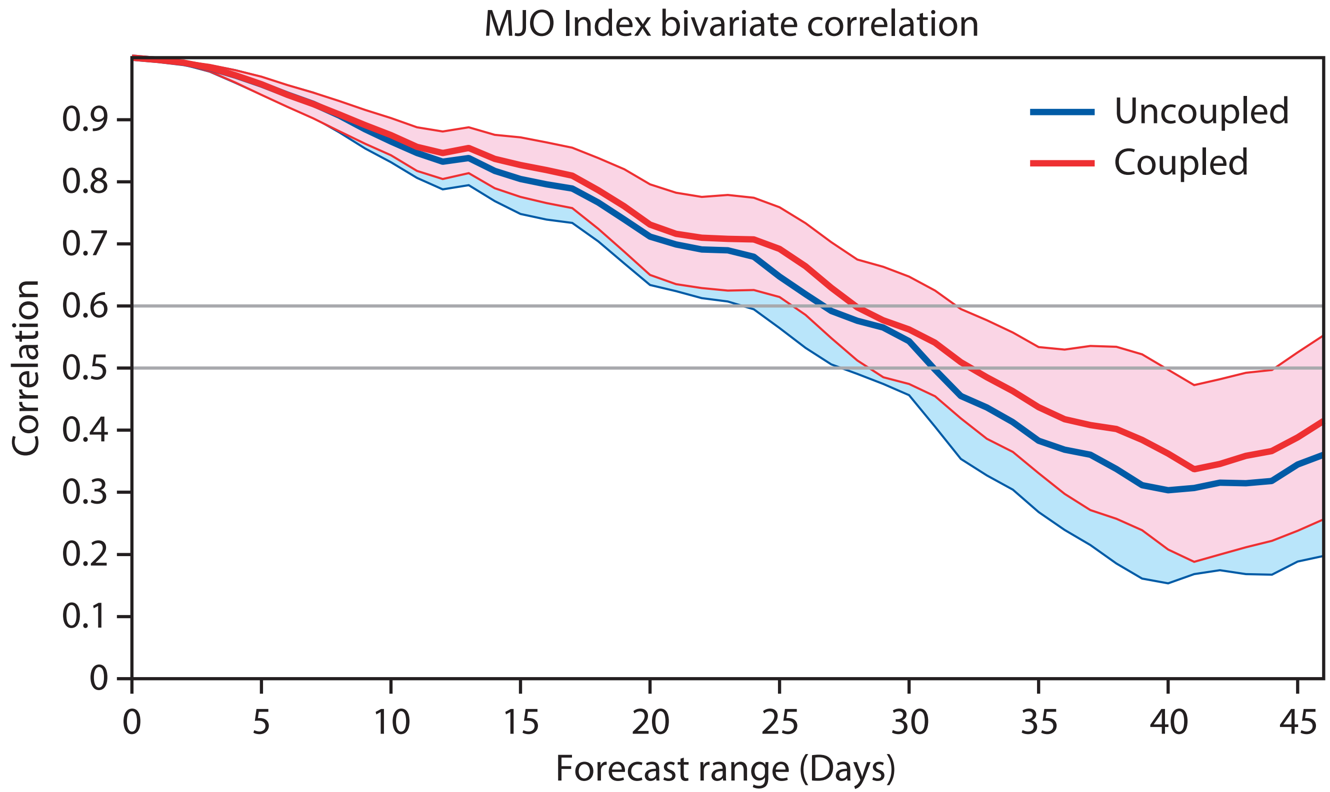


Figure 15: MJO Bivariate correlation for the control runs (legA uncoupled, blue curve) and coupled from day 0 integrations (leg A coupled, red curve). The shaded areas represent the 5% level of confidence using a 10,000 re-sampling bootstrap procedure.

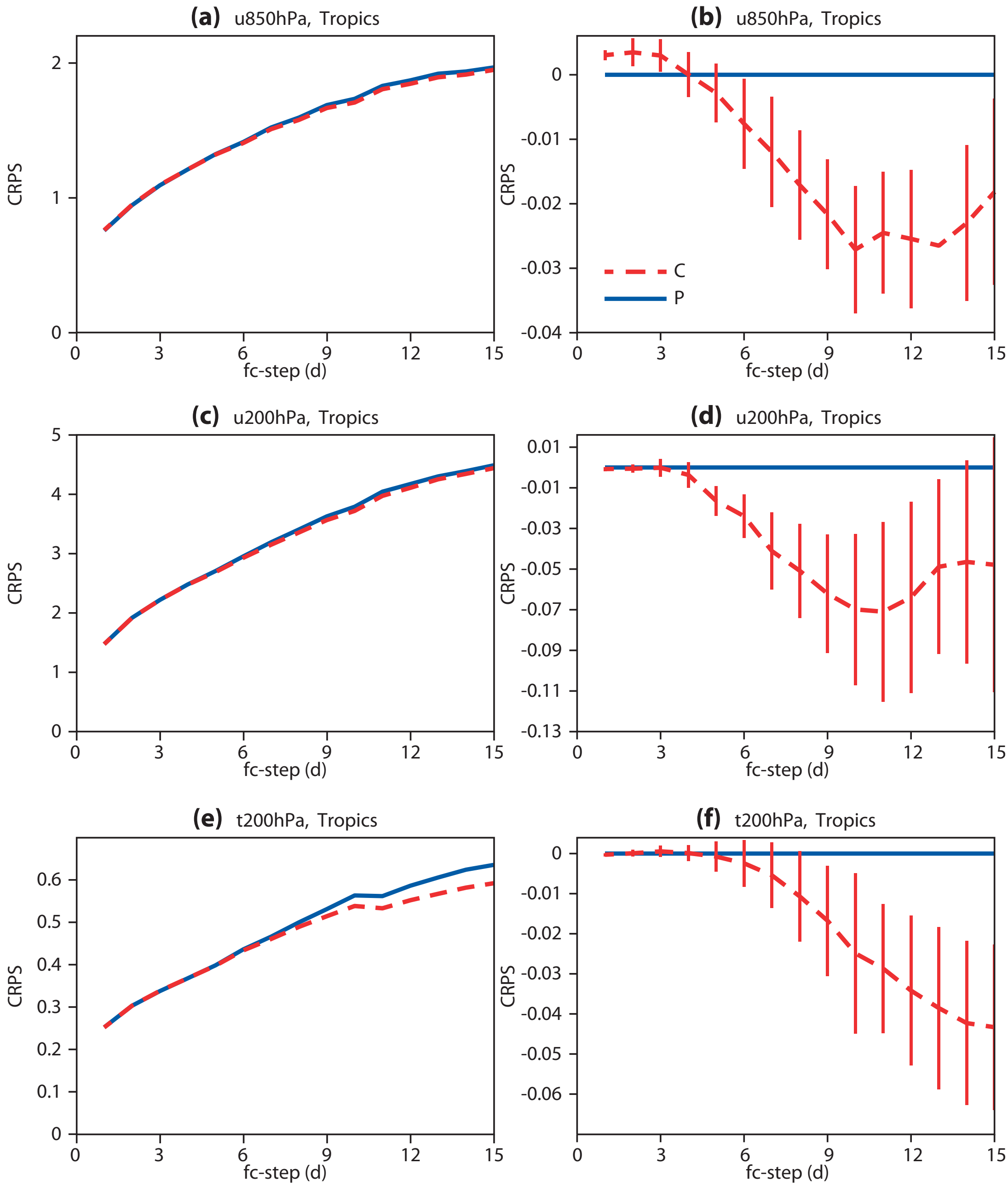
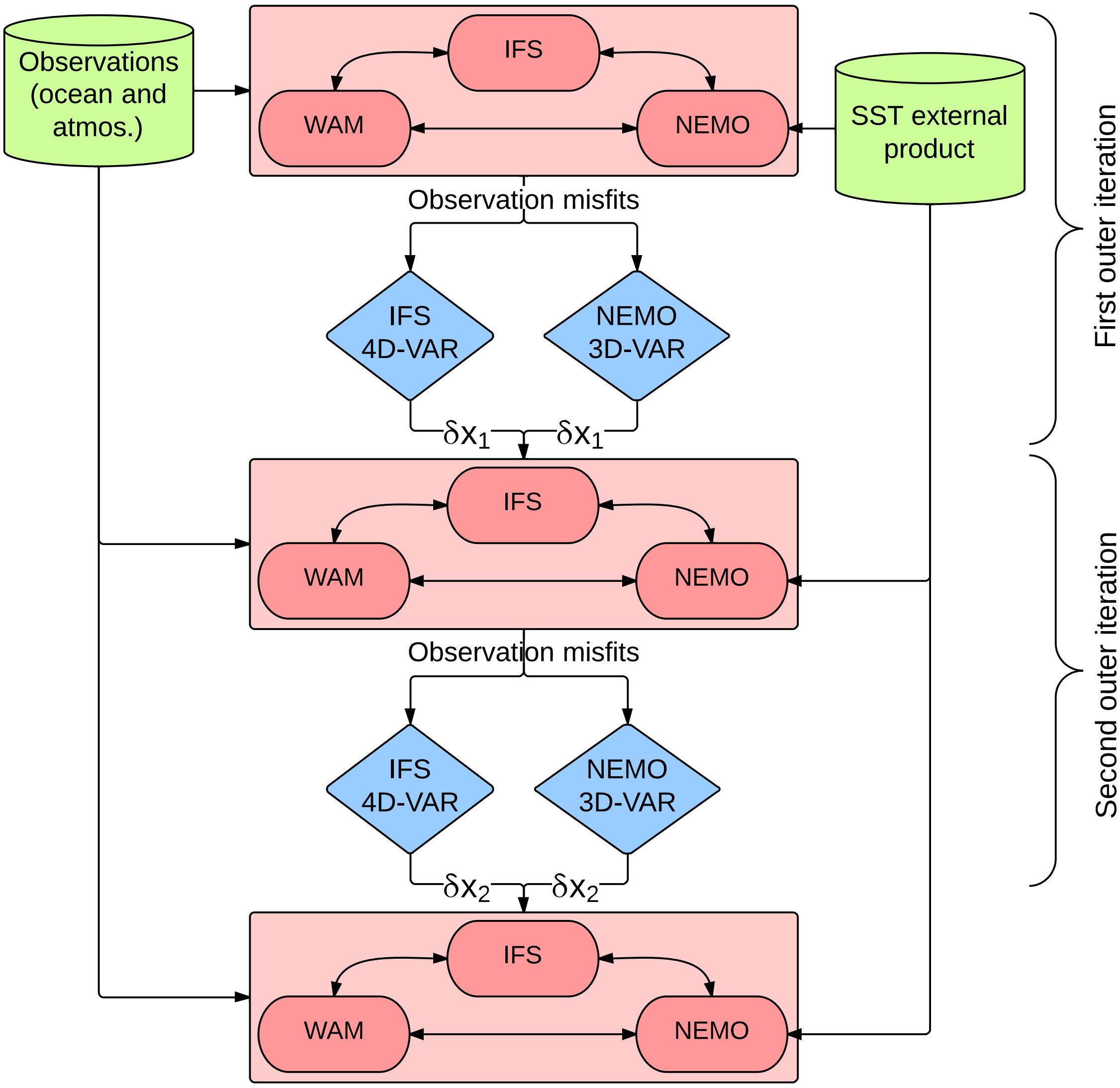


Figure 16: Impact of coupling on CRPS for tropics: (a,b) zonal wind at 850 hPa, (c,d) zonal wind at 200 hPa and (e,f) temperature at 200 hPa.

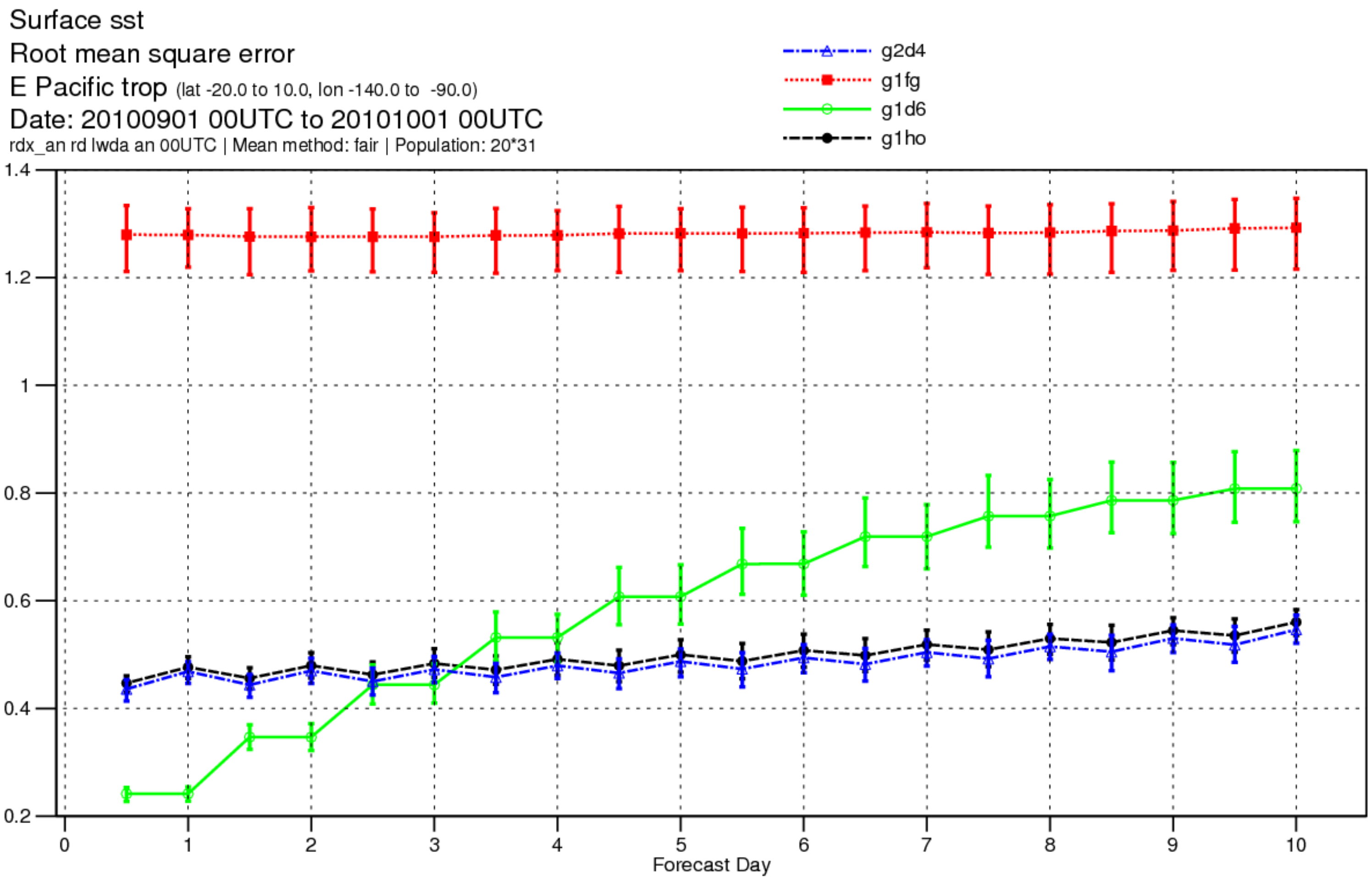
WEAKLY COUPLED DATA ASSIMILATION



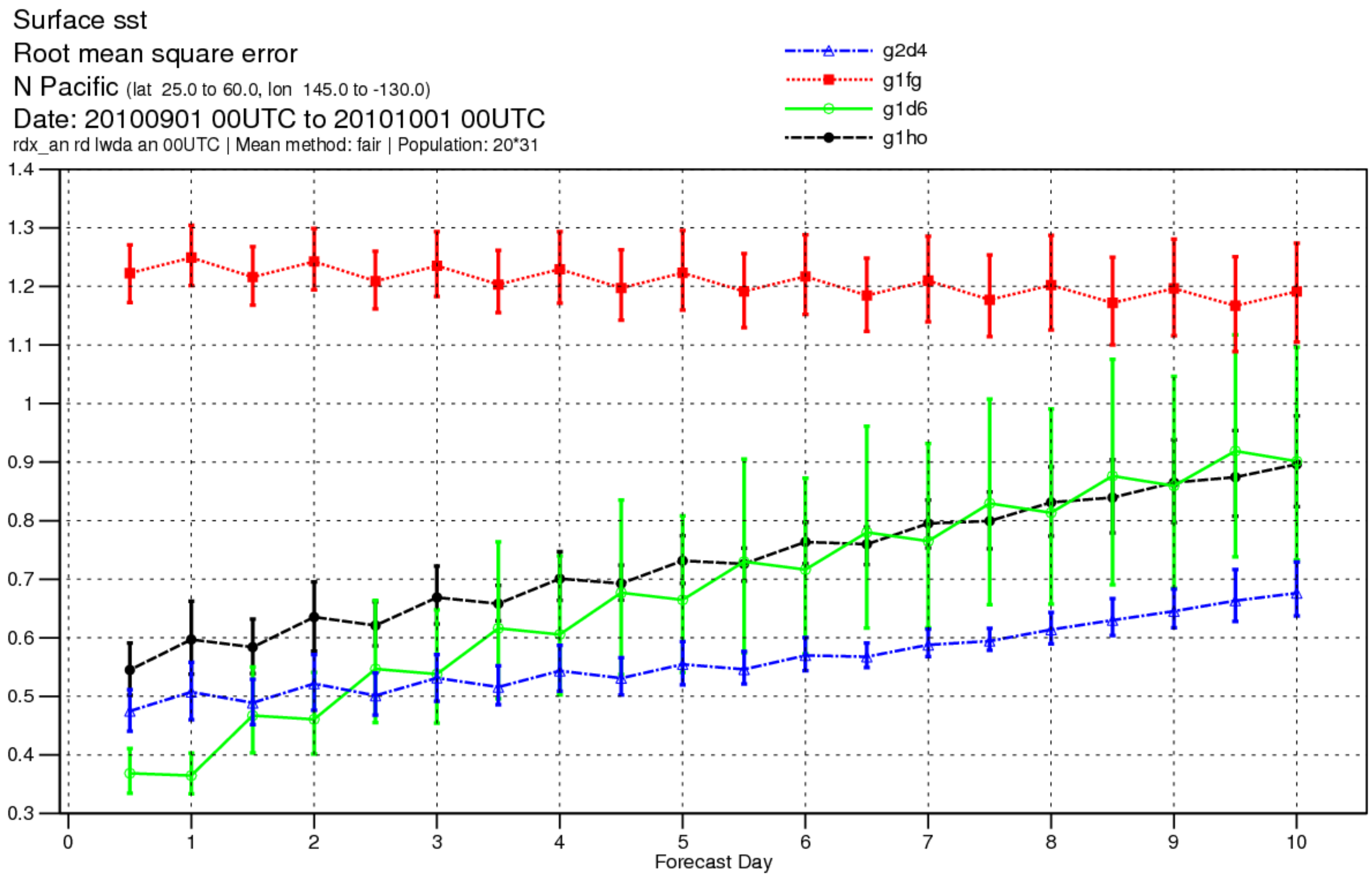
Data assimilation and sea state effects

The success of data assimilation depends on the quality of the model. The reference model used so far has too much mixing in the North Pacific in the Summer time. Hence, the introduction of sea state dependent mixing is expected to have a pronounced impact on SST scores.

Show results in a set up where at analysis time SST observations are nudged into the model state.



Comparison of SST error from weakly coupled analysis (including sst nudging) with present practice for East Tropical Pacific. Blue, with Sea-state dependent mixing. Black, without.



Comparison of SST error from weakly coupled analysis (including sst nudging) with present practice for North Pacific. Blue, with sea state dependent mixing. Black without.

CONCLUSIONS

- There might be a need to have a systematic study on numerics of coupled systems.
- Wave breaking enhances the upper ocean mixing and even affects the average SST field over a 20 year period, while there is a hint that it might have impact on predictability.
- There is a trend towards the introduction of more complicated EMS's. Not only forecasting but also data assimilation needs to be done in the context of a coupled system.
- Work on the development of a weakly coupled data assimilation system is in progress.