Impact of sea state on atmosphere and ocean

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ABSTRACT

Ocean waves represent the interface between the ocean and the atmosphere, and, therefore, wave models are needed to compute not only the wave spectrum, but also the processes at the air-sea interface that govern the fluxes across the interface.

This was one of the reasons for developing a wave prediction model, called the WAM model, that determines the sea state dependence of the air-sea fluxes. As a first step, the study of the two-way interaction between ocean waves and atmospheric circulation was undertaken. Modest improvements in forecasting waves and winds were obtained, and, as a consequence, since the 29th of June 1998 ECMWF has been producing weather and ocean wave analyses and forecasts, using a coupled IFS-WAM forecasting system.

In this paper, we briefly discuss our recent experience with two-way interaction. Nowadays, there is a substantial impact on weather and wave forecasting, and reasons for this increase in impact are given. The next task is to couple the ocean waves and the ocean circulation. It is argued that prediction of the ocean circulation would benefit from the inclusion of currents in the calculation of fluxes. In addition, it is shown that the thickness of the ocean mixed layer is to a large extent determined by the energy flux associated with breaking ocean waves. Finally, the momentum flux is controlled to a lesser extent by breaking waves.

1 Introduction.

Ocean waves represent the interface between the ocean and the atmosphere, the two most important systems governing the dynamics of climate and global change. A realistic description of the physical processes occurring at the ocean-atmosphere interface is essential for a reliable determination of the air-sea fluxes of momentum, sensible and latent heat, CO_2 and other trace gases, and aerosols. It is known that the wave field is intimately involved in these exchange processes, and, therefore, wave models are needed to compute not only the wave spectrum, but also the processes at the air-sea interface that govern the fluxes across the interface.

In the context of this extensive programme, a wave prediction system, called the WAM model, was developed that determines the sea state dependence of the air-sea fluxes (Komen et al, 1994). As a first step, the study of the two-way interaction between ocean waves and atmospheric circulation was undertaken. This interaction takes place on a relatively short time scale of a few days. Modest improvements in medium-range forecasting of waves and winds were obtained and, as a consequence, since the 29th of June 1998 ECMWF has been producing weather and ocean wave analyses and forecasts, using a coupled IFS-WAM forecasting system. On a seasonal time scale, depending on spatial resolution, however, a more substantial impact of ocean waves on the atmospheric climate was found (Janssen and Viterbo, 1996).

In this paper, we briefly discuss our recent experience with two-way interaction. Nowadays, this interaction has a substantial impact on weather and wave forecasting, one of the reasons being considerable increases in the spatial resolution of the atmospheric model which allows a more realistic representation of the small scales. These are the spatial scales that matter for small-scale air-sea interaction.

The next step in the development of one model for our geosphere is to study the impact of the sea state on the ocean circulation. There are a number of ways in which the sea state could affect the evolution of the ocean state. In the usual description of the ocean the momentum of the ocean waves is not taken into account, nevertheless a considerable list of authors (Hasselmann, 1970; Weber, 1983; Jenkins, 1987a; Xu and Bowen, 1994; McWilliams and Restrepo, 1999) have pointed out that in a rotating ocean the ocean waves excert a wave-induced stress on the Eulerian mean flow which results in a force equal to $\vec{u}_s \times \vec{f}$, where $\vec{f}$ is the Coriolis parameter, and $\vec{u}_s$ equals the Stokes drift. This additional force has a considerable impact on the Ekman turning of the surface current (Weber, 1983; Jenkins, 1987b). However, it is not clear whether the introduction of this additional force is appropriate, for the simple observation that conservation of momentum seems to be violated. This will be discussed in a different context in more detail.

Rather, we concentrate on the role ocean waves have in the transfer of momentum and energy to the ocean. In the first place, the roughness at the sea surface and hence the atmosheric flux provided by the atmosphere to the ocean waves is sea state dependendent. This sea-state dependence of the surface stress has a systematic impact on the temperature distribution of the ocean (Burgers et al, 1995). Regarding momentum and energy transfer to the ocean it is noted that in growing circumstances the ocean waves retain a small part of the momentum and energy (which is spent on wave growth). When ocean waves become swell the excess momentum and energy is lost because dissipation by wave breaking dominates the wind input. Furthermore, the momentum transfer is dominated by the high-frequency part of the spectrum (which are in equilibrium with the wind), hence there are only small differences between the atmospheric stress and the momentum flux to the ocean.

The energy flux is commonly parametrized as being proportional to u_*^3 (with u_* the air-friction velocity) where the proportionality constant is of the order of 5. We show that this is indeed a good approxiation in the generation phase of ocean waves. However, when ocean waves propagate out of the storm area, energy fluxes caused by breaking waves are much larger than as found from the above empirical rule. Finally, it is also argued that prediction of the ocean circulation would benefit from the inclusion of currents in the calculation of fluxes.

2 Impact on the atmosphere.

In this section a brief description of the impact of sea-state dependent drag on the atmospheric circulation is given. The basic idea is described in Janssen (1982) and Janssen (1989) while a parametrization of the sea-state dependent roughness is developed in Janssen (1991). This parametrization is included in WAMCy4 (Komen et al, 1994). A review of the impact on atmospheric circulation is given in Janssen et al (2002).

The basic idea is that momentum transfer from air to sea depends on the sea state because steep waves extract more momentum from the air flow than gentle, smooth waves. Steep waves typically occur in the early stages of wind-wave generation and when a frontal system passes, hence momentum transfer depends on the sea state. In order to account for this effect one needs to calculate the wave-induced stress τ_w which depends on the two-dimensional wave spectrum. Therefore the determination of the wave stress requires the solution of the energy balance equation

$$\frac{\partial}{\partial t}F + \vec{v}_g . \frac{\partial}{\partial \vec{x}}F = S_{in} + S_{nl} + S_{diss}, \tag{1}$$

where $F = F(\omega, \theta)$ is the two-dimensional wave spectrum which gives the energy distribution of the ocean waves over angular frequency ω and propagation direction θ. Furthermore, $\vec{v}_g$ is the group velocity and on the right hand side there are three source terms. The first one, S_{in} describes the generation of ocean waves by wind and therefore represents the momentum and energy transfer from air to ocean waves. The third term describes the dissipation of waves by processes such as white-capping, while the second terms denotes nonlinear transfer by resonant four-wave interactions. The nonlinear transfer conserves total energy and momentum and is important in shaping the wave spectrum and in the down-shift towards lower frequencies. In order to appreciate the role of the respective source terms, in Fig. 1 we have plotted the directionally averaged source functions S_{in}, S_{nl}, and S_{diss} (as developed for WAMCy4) as function of frequency for young windsea when a 20 m/s wind is blowing for just 3 hours. This figure shows a typical picture of the energy balance for growing ocean waves, namely the intermediate frequencies receive energy from the airflow which is transported by the nonlinear

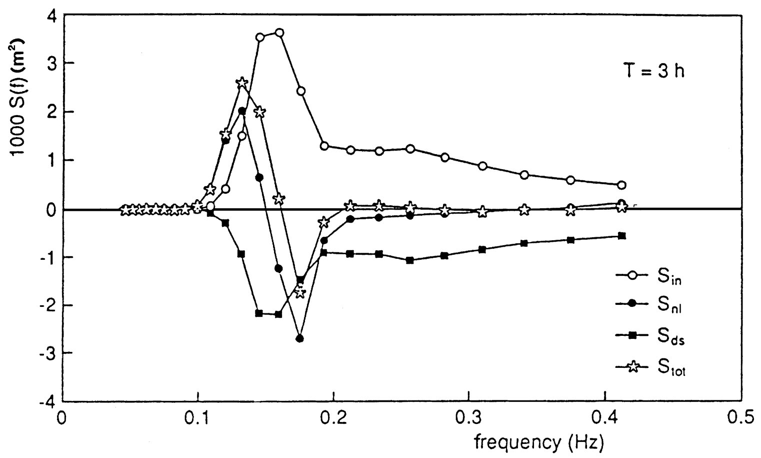

Figure 1: Energy balance for young windsea for a 20 *m/s wind speed.*

interactions towards the low and high frequencies where it is dissipated by processes such as white capping. The consequence is that the wave spectrum shows a shift of the peak of the spectrum towards lower frequencies, while a considerable enhancement of the peak energy of the spectrum is also noticed in the early stages of wave growth.

At the same time Fig. 1 illustrates the role ocean surface waves play in the interaction of the atmosphere and the ocean, because on the one hand ocean waves receive momentum and energy from the atmosphere through wind input (controlling to some extent the drag of air flow over the oceans), while on the other hand, through wave breaking, the ocean waves transfer energy and momentum to the ocean thereby feeding the turbulent and large-scale motions of the oceans. The energy-conserving nonlinear transfer plays no direct role in this interaction process, although it determines to a large extent the shape of the wave spectrum, and therefore controls energy and momentum fluxes in an indirect way. In equilibrium conditions, the fluxes received by the ocean waves from the atmosphere through the wind input term would balance the fluxes from ocean waves to ocean via wave breaking. However, ocean waves are in general not in an equilibrium state determined by the balance of the three source functions, because advection and unsteadiness are important as well. As a rule of thumb, of the amount of energy gained by wind, about 90% is lost locally to the ocean by wave breaking, while the remaining 5% is either advected away or is spent in local growth. As illustrated in Fig.1, which also shows a plot of the total source function, for young windseas there may, however, be a considerable imbalance, in particular for the low-frequency waves. On the other hand, when wind waves leave a storm area the magnitude of the wind input source function decreases dramatically, while the waves are still sufficiently steep so that white capping is still important. Since dissipation dominates, wave energy will decay and as a consequence momentum and energy flux to the ocean may be larger than the amounts received by the waves from the atmosphere.

It would be of considerable interest to develop a coupled atmosphere- ocean circulation system where the ocean waves are the agent that transfers energy and momentum across the air-sea interface in accordance with the energy balance equation. In this section we shall concentrate on just one aspect of the overall problem, namely the mutual interaction between wind and waves. In the next section we discuss the possible impacts on the ocean circulation.

In the wind-wave interaction problem we only need to know the wave-induced stress τ_w which follows from an integration of the input source function of the energy balance equation (1)

$$\tau_w = \rho_w g \int d\omega d\theta \; S_{in}/c, \tag{2}$$

where c is the phase speed of the gravity waves and ρ_w the water density. Here, it should be realized that wave momentum P and energy density F of the waves are related by $P = F/c$ and the wave stress is the rate of change of total wave momentum by wind input. Because waves grow exponentially fast the source function S_{in} is proportional to the wave spectrum itself. The wave-induced stress is mainly determined by the high-frequency part of the wave spectrum because these are the waves that have the largest growth rate due to wind.

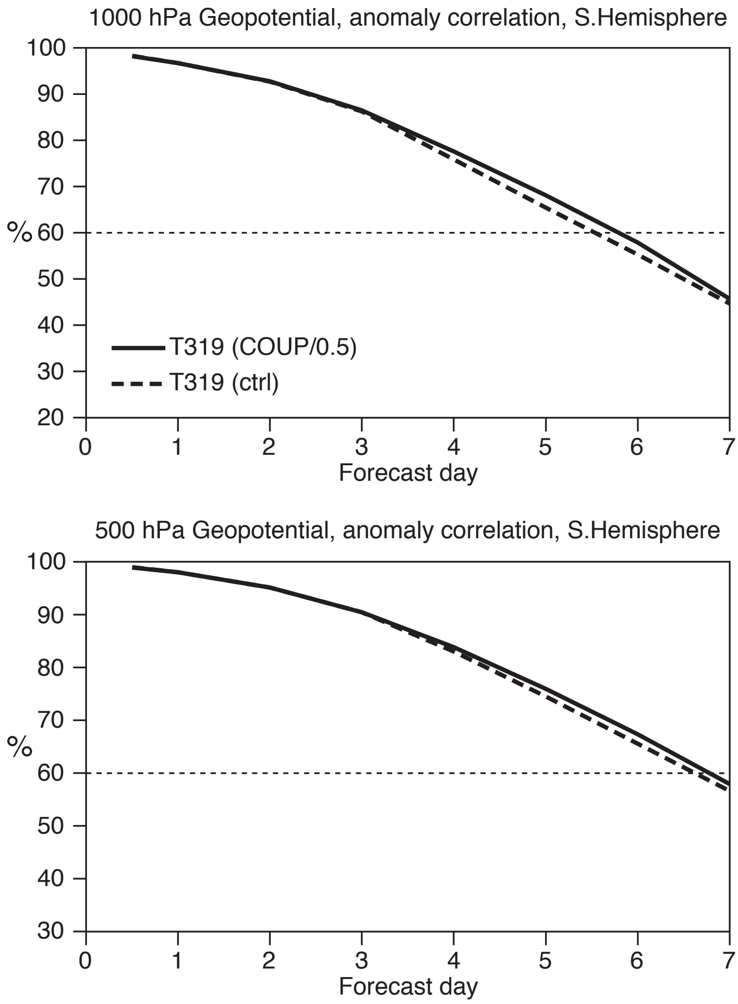

Figure 2: Scores of forecast 1000 *and* 500 *mb geopotential for the Southern Hemisphere for 28 cases in the December 1997-January 1998 period.*

Since it is known that the high-frequency spectrum depends on the stage of development of the windsea (for example, young wind waves are steeper than old wind waves) it follows that the wave-induced stress depends on the sea state. Therefore, young wind waves represent a rougher surface than gentle old windsea. The roughness z_0 therefore depends on the sea state and following the work of Janssen (1991) one finds for the roughness length a Charnock relation,

$$z_0 = \alpha u_*^2/g, \tag{3}$$

where the Charnock parameter depends on the sea state according to

$$\alpha = \frac{\beta}{\sqrt{1-\tau_w/\tau}},\ \beta = 0.01, \tag{4}$$

with $\tau = \rho_a u_*^2$ is the surface stress and u_* the friction velocity.

At ECMWF we have developed a coupled ocean-wave, atmosphere model in such a way that the wave model is called as a subroutine from the IFS system. This system was introduced in operations on the 29th of June 1998. Presently, every atmospheric time step wind fields, air density fields and a gustiness factor are passed from the atmospheric model to the wave model. Then the wave model integrates one time step and determines the two-dimensional wave spectrum. The wave-induced stress is obtained from Eq. (2) which is followed by a determination of the Charnock parameter field. The loop is closed by passing the Charnock field to the atmospheric model which then continues with the next time step.

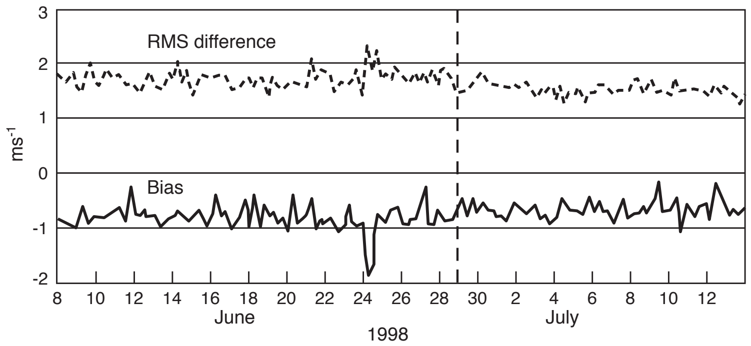

Figure 3: Bias (ERS-2 minus EC FG) and rms difference between the background ECMWF surface winds and the ERS-2 Scatterometer wind measurements.

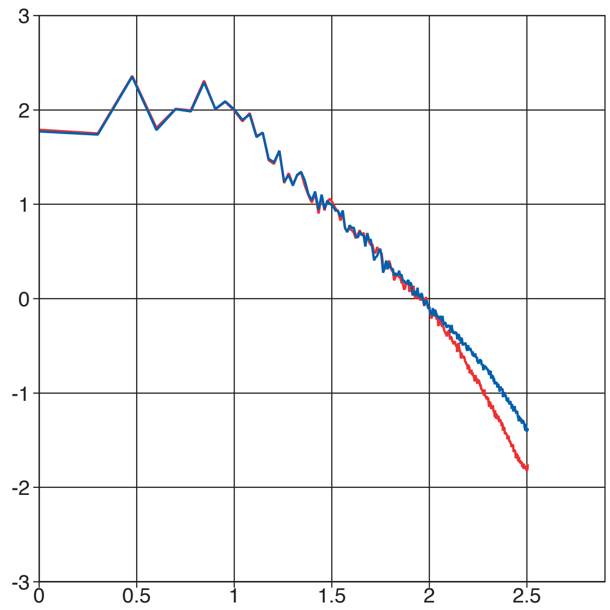

Figure 4: Comparison of surface kinetic energy spectrum as function of total wave number for T_l511 (blue) and T_l319 (red).

With this system we have performed a number of impact studies the results of which will be briefly described in the following sections.

2.1 Impact studies: medium-range forecasting.

Initial experiments showed a modest impact of sea-state dependent roughness on atmospheric scores. We illustrate this in Fig. 2 for 28 analyses and 10 day forecasts for the Southern Hemisphere summer time. Shown are anomaly correlation of the 1000 and 500 mb geopotential field. The resolution of the atmospheric model was T_l319 while the wave model had a spatial resolution of 0.5 deg. No impact of this size in the Northern Hemisphere scores were found.

When the two-way interaction of winds and waves was introduced in operations on the 29th of June 1998 there was a pronounced improvement of the quality of the surface wind field. Routinely, first-guess (FG) winds are compared with scatterometer winds (from ERS-2 in this case). As shown in Fig.3 which displays timeseries of bias (ERS-2-FG) and the rms difference, there is a considerable reduction of 10% in the rms error after the introduction of two-way interaction.

However, currently the impact of two-way interaction of wind and waves is more substantial. The main reason for this is an increase of atmospheric resolution from T_l319 to T_l511 (or from 65 to 40 km) which allows for a more realistic representation of small spatial scales. It is emphasized that these are the scales that matter for air-sea interaction. That the present atmospheric system has more realistic levels of kinetic energy at the small scales is illustrated by Fig. 4 where we have compared surface kinetic energy spectra from the T_l319 version of the IFS with the T_l511 version. At high wave numbers (small scales) energy levels from the high-resolution model are higher by at least a factor of two.

The more sensitive dependence of the T_l511 version of the IFS on the sea-state dependent drag became evident when we performed experiments with a doubling of angular resolution of the wave spectrum from 12 to 24 directions. Trials with the T_l319 version showed an improvement of forecast skill between 1 and 2 hours. However, when experiments were performed with the T_l511 version of the IFS a substantially larger impact of the increase of angular resolution was found. This is illustrated in Fig. 5 which compares forecast performance in Northern and Southern Hemisphere for 24 cases in August 2000. Note, that as a rule of thumb we usually find more impact in the summer time, presumably because physical processes near the surface play a more important role in the evolution of the weather. In winter time the atmospheric circulation is dominated by baroclinic activity, and physical processes such as surface friction play a relatively minor role, although, there may be a considerable small scale impact in cases of rapidly developing lows (Doyle, 1995; Janssen et al, 2002).

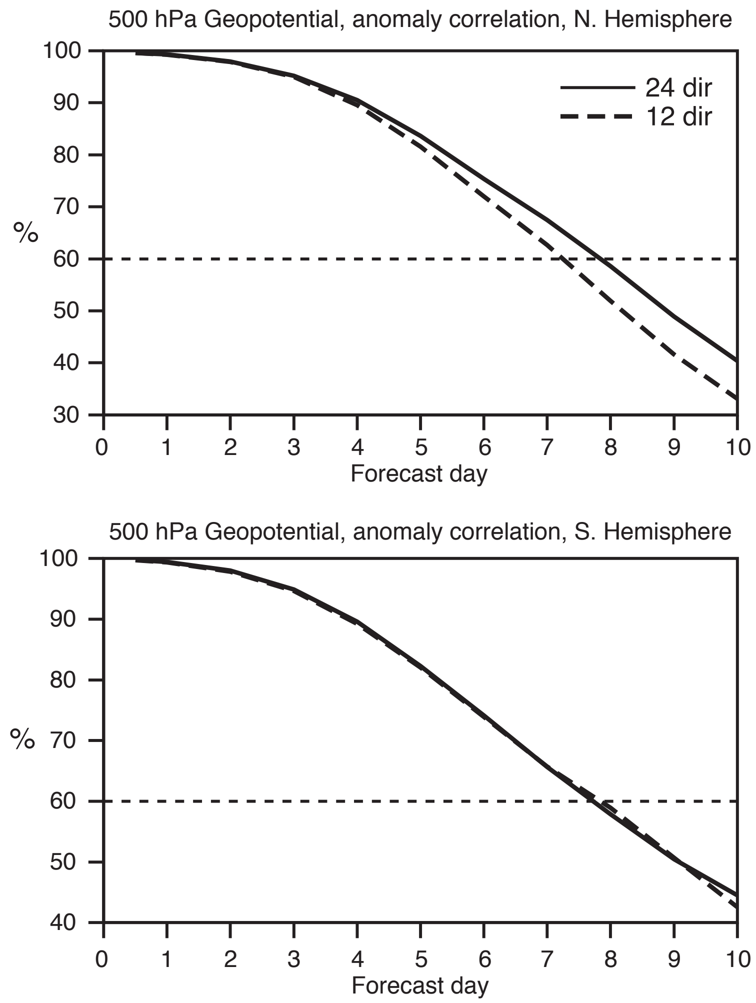

Figure 5: Anomaly correlation of 500 mb geopotential height for the Northern and the Southern Hemisphere for the last 24 days in August 2000. Here, the impact of increased angular resolution on the forecast performance of the T_l511 IFS forecast system is shown.

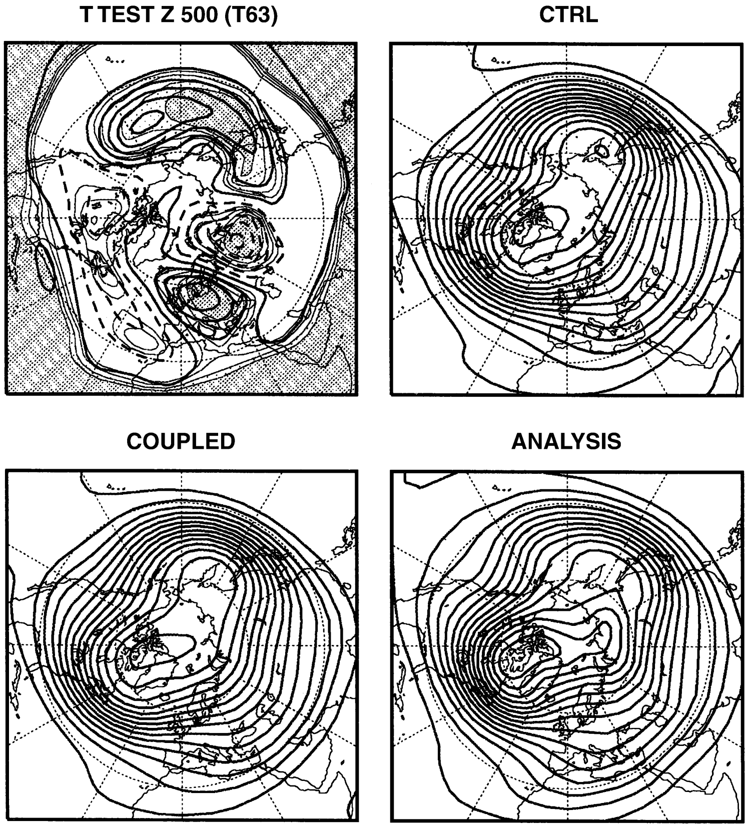

Figure 6: Ensemble mean of coupled and control run and their differences. For comparison the analysed climate is also shown. Period is winter 1990 and area is Northern Hemisphere. The shading indicates a measure of significance. Heavy shading means that there is a probability of 95% *that the difference is significant.*

2.2 Impact studies: seasonal integrations.

Janssen and Viterbo (1996) studied the impact of two-way interaction on the seasonal time scale. In order to obtain reliable information on the impact of waves on the atmospheric circulation there is a need for ensemble forecasting, because the variability of the weather, especially over the oceans is high. Therefore 15 coupled and control runs were performed for the winter season of 1990 starting from the analysis of 15 consecutive days. The atmospheric resolution was $T63$, and the wave model had a resolution of 3 deg, while the length of the runs was 120 days. By taking a time average over the last 90 days, followed by an ensemble average a reliable estimate of the mean state of one season could be provided. At the same time, information on the variability may be inferred from the scatter around the mean, and thus a student t-test may be applied to test statistical significance of the mean difference between coupled and control run. In Fig. 6 we have plotted the ensemble mean of 500 mb height field and their differences for the Northern Hemisphere while for comparison purposes, we also display the 90-day mean of the corresponding ECMWF analysis. Contours in the mean are plotted every 60 m, while in the difference plot we have indicated by heavy shading the probability of 95% (or more) that the two fields in question are not equal. Significant differences are noted in the storm track areas of the Northern Hemisphere (and, not shown, also for the Southern Hemisphere). We note differences over the Northern Pacific, Europe and Siberia. In the last two areas the coupled climate shows, when compared to the analysis, a considerable improvement. There are also improvements in low-frequency variability over the North Atlantic (not shown).

As far as impact of ocean waves on the atmospheric climate is concerned it should be emphasized that also here resolution of the atmospheric model plays a crucial role. Janssen and Viterbo (1996) also performed seasonal forecasts with the $T21$ version of the coupled system and particularly in the Southern Hemisphere a much reduced impact of the sea-state dependent drag on the atmospheric circulation was found. This should not come as a surprise when it is realized that with $T21$ the mean wind speeds are reduced by as much as 50%, therefore giving a much weaker coupling between wind and waves.

2.3 Impact studies: ocean circulation.

The study by Janssen and Viterbo (1996) also revealed that there were quite large changes in the surface stress in the warm pool area east of Indonesia. Because this area plays a prominent role in understanding certain issues in El Nino prediction, it was thought to be of interest to generate stresses over a one year period in order to investigate the impact of the sea-state dependent momentum transfer on ocean circulation. The long period of one year was thought to be necessary because of the long response times of the ocean circulation.

The stress fields were supplied to Dave Anderson (then at Oxford University) and Gerrit Burgers (KNMI) who forced their tropical ocean model with the coupled and control fluxes. Both models gave considerable differences in the temperature distribution of the surface layer of the ocean (Burgers et al, 1995). An integration period of 6 months gave already a good idea of the kind of impact, which was typically of the order of $1\,\mathrm{deg} K$. However, the difference patterns of the two models were surprisingly different. One model showed differences with fairly small spatial scale of the order of 2000 km, while the difference pattern in the other model covered the whole tropical Pacific.

Note that such experiments most likely exaggerate the size of the impact, because there may be an important feedback from the ocean to the atmosphere. The present ECMWF seasonal forecasting system consists of a coupled atmosphere, ocean circulation model. The atmospheric model is coupled to the ocean waves model in two-way interaction mode. Coupling of wind and waves gave a beneficial reduction in the drift in the mean temperature, but the size of the reduction was relatively modest ($0.2\,\mathrm{deg} K$ out of a drift of about $1\,\mathrm{deg} K$ in 6 months) (T. Stockdale, private communication 2003).

3 Towards a more realistic air-sea interface for ocean modelling.

Ocean waves are the agent that takes care of the momentum and energy transfer from atmosphere to the ocean. Through the process of wave dissipation by, for example, white-capping wave energy is transferred to the ocean column which then generates turbulence and large scale motion of the ocean. It is pointed out that it only makes sense to determine the fluxes through the waves if the circumstances are sufficiently non-stationary or inhomogeneous. If the surface gravity waves were in equilibrium with the wind, the air-sea interface would be transparent, because the energy and momentum received from the wind would be immediately transferred to the ocean column. However, whether unsteadiness and inhomogeneity (or, in other words, wave growth and energy advection) play a role, can only be decided by a determination of these fluxes in actual circumstances. Therefore, the fluxes into the ocean are determined for the month of January 2003 using WAMCy4 and operational analyzed wave spectra from ECMWF. In particular, there are considerable deviations from the atmospheric fluxes in the case of the energy flux.

However, first we discuss the need to include ocean currents in the determination of the atmospheric-ocean fluxes, and to provide the proper boundary condition for the atmospheric flow.

3.1 Impact of ocean currents on atmospheric fluxes.

As was shown by Pacanowski (1987) inclusion of currents in the determination of the atmospheric fluxes result in considerable impact on the ocean circulation and the temperature distribution in the equatorial region. This can be readily seen as the wind speeds in that area are typically $6 m/s$ while the surface currents can reach values of up to $1 m/s$, hence differences in the momentum flux may be up to 30%.

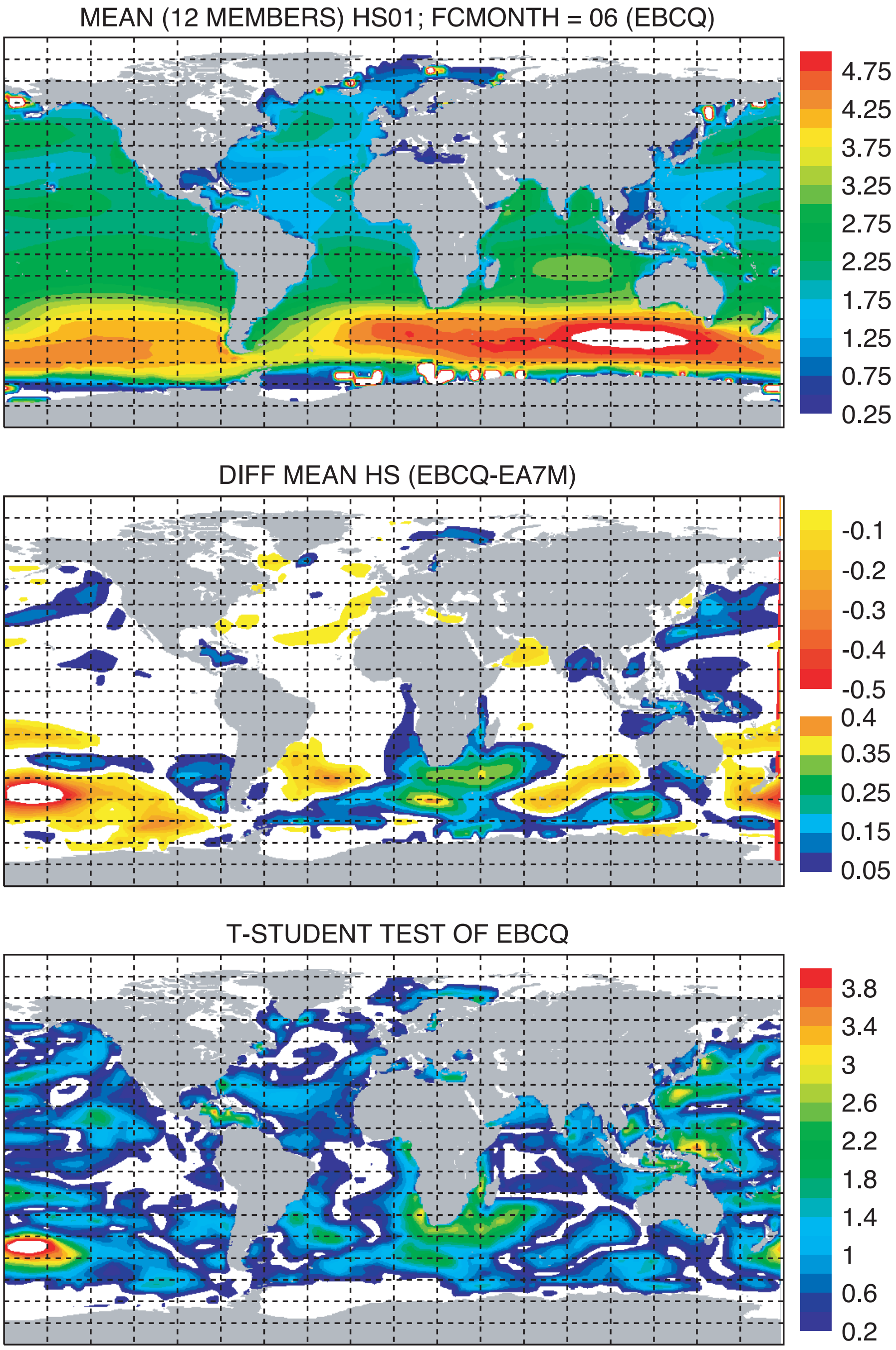

Figure 7: Impact of currents on fluxes as reflected by differences in the monthly mean wave height field. Top panel: mean of sixth month experimental forecast for July averaged over the 12 July's over the period 1991-2002. Middle panel: Difference with control. Bottom panel: significance parameter T; if $T > 2$ then the difference is significant with probability of 95%*.*

Again, feedback from the ocean circulation to the atmosphere was not taken into account so that the results of Pacanowski (1987) may overstate the case. In order to investigate the size of the impact of the inclusion of currents in the flux determination we ran 6 months forecasts with the ECMWF seasonal forecasting system over the period of 1991 until 2002. The forecasts started in January and July. There was a beneficial reduction of the drift in SST in the Equatorial Pacific, but the size of the reduction was only $0.1\,\deg K$. Nevertheless, there may be considerable differences in physical parameters such as the surface wind speed or significant wave height. This is shown in Fig. 7 which gives the monthly wave height field for the 6 month experimental forecast for July averaged over the twelve year period. In addition is shown the difference with the corresponding field of the control forecast, and the result of a significance test. There are considerable differences in the wave height field of the southern Hemisphere extra-tropics, and in the warm pool area east of Indonesia. The student t-test in the bottom panel reveals all of the major current systems of the ocean, except perhaps the Gulf stream in the North Atlantic.

In the work of Pacanowski (1987) and our experiments the ocean surface current was approximated by means of the current at 5 m depth. However, it is known that this is a poor approximation; ocean waves play an important role in the top layer of the ocean, resulting in an additional surface drift of about 2.5% of the surface wind speed. In addition, because of Ekman turning, there may be considerable differences between the direction of the surface current and the one at 5 m depth. These effects can, however, only be taken into account by using sea state information within the context of a coupled ocean-circulation, atmosphere model.

3.2 Surface layer mixing and ocean waves.

The work of Terray et al (1996) and Craig and Banner (1994) has highlighted the prominent role of breaking waves and its contribution to the surface current. For example, in the field considerable deviations from the usual balance between production and dissipation of turbulent kinetic energy are found which are caused by the energy flux produced by breaking waves. When observed turbulent kinetic energy dissipation, ε, and depth z are scaled by parameters related to the wave field, an almost universal relation between dimensionless dissipation and dimensionless depth is found. Here, dimensionless dissipation is given by $\varepsilon H_S/\Phi_{aw}$, with H_S the significant wave height and Φ_{aw} the energy flux from wind to waves, while the dimensionless depth is given by z/H_S.

In the Craig and Banner model (1994) the difference between production and dissipation of turbulent kinetic energy is balanced by a flux of turbulent energy, following the work of Mellor and Yamada (1982). In particular, by choosing a wave-height dependent mixing length, Terray et al (1999) found a good agreement between modelled dissipation and current profile on the one hand and observations on the other hand.

In order to be able to give a realistic representation of the mixing processes in the surface layer of the ocean it is clear that a reliable estimate of energy and momentum fluxes to the ocean column is required. A first attempt to estimate these fluxes from modelled wave spectra and knowledge about the generation and dissipation of ocean waves was given by Komen (1987). Weber (1994) studied energy and momentum fluxes in the context of a low-resolution coupled ocean-wave atmosphere model (WAM-ECHAM), and she concluded that there is no need to use a wave prediction model to determine, for example, the energy flux. A parametrization of the type $\Phi_{aw} = m\rho_a u_*^3$ (with u_* the air friction velocity and m a constant) would suffice. It will be argued that this conclusion depends on an approximation used by Weber to estimate the energy flux.

Let us first define the momentum and energy flux. The total wave momentum $\vec{P}$ depends on the variance spectrum $F(\omega,\theta)$ and is defined as

$$\vec{P} = \rho_w g \int_0^{2\pi} \int_0^{\infty} d\omega d\theta\ \frac{\vec{k}}{\omega} F(\omega,\theta), \tag{5}$$

which agrees with the well-known relation that wave momentum is simply wave energy divided by the phase speed of the waves. The momentum fluxes to and from the wave field are given by the rate of change in time of wave momentum, and one may distinguish different momentum fluxes depending on the different physical processes. For example, making use of the energy balance equation (1) the wave-induced stress is given by (cf. 2)

$$\vec{\tau}_{aw} = \rho_w g \int_0^{2\pi} \int_0^{\infty} d\omega d\theta \; \frac{\vec{k}}{\omega} S_{in}(\omega,\theta), \tag{6}$$

while the dissipation stress is given by

$$\vec{\tau}_{wo} = \rho_w g \int_0^{2\pi} \int_0^{\infty} d\omega d\theta \; \frac{\vec{k}}{\omega} S_{diss}(\omega,\theta), \tag{7}$$

Similarly, the energy flux from wind to waves is defined by

$$\Phi_{aw} = \rho_w g \int_0^{2\pi} \int_0^{\infty} d\omega d\theta \; S_{in}(\omega,\theta), \tag{8}$$

and the definition for Φ_{wo} follows immediately from the above one by replacing S_{in} by S_{diss}. It is important to note that while the momentum fluxes are mainly determined by the high-frequency part of the wave spectrum, the energy flux is to a larger extent determined by the low-frequency waves.

In an operational wave model, the prognostic frequency range is limited by practical considerations such as restrictions on computation time, but also by the consideration that the high-frequency part of the dissipation source function is not well-known. In the ECMWF version of the WAM model the prognostic range of the wave spectrum is given by the condition

$$\omega < \omega_c = 2.5\omega_{mean} \tag{9}$$

where ω_{mean} is a conveniently defined mean angular frequency. In the diagnostic range, $\omega > \omega_c$, the wave spectrum is given by Phillips' ω^{-5} power law. In the diagnostic range it is assumed that there is a balance between wind input, dissipation and nonlinear transfer. In practice this means that all energy and momentum going into the high-frequency range of the spectrum is dissipated, and is therefore directly transferred to the ocean column.

As a consequence, the momentum flux to the ocean, $\vec{\tau}_{oc}$, is given by

$$\vec{\tau}_{oc} = \vec{\tau}_a - \rho_w g \int_0^{2\pi} \int_0^{\omega_c} d\omega d\theta \; \frac{\vec{k}}{\omega} \left(S_{in} + S_{nl} + S_{diss}\right), \tag{10}$$

where $\vec{\tau}_a$ is the atmospheric stress, whose magnitude is given by $\tau_a = \rho_a u_*^2$. Note that ocean momentum flux $\vec{\tau}_{oc}$ only involves the sum of the three source functions of the energy balance equation and therefore it only involves the total rate of change of wave momentum. Any wave model that is forced by reliable atmospheric stresses and that produces wave height results that compare well with, for example, buoy wave height data and Altimeter wave height data, will produce reliable estimates of the ocean momentum flux $\vec{\tau}_{oc}$.

Ignoring the direct energy flux from air to currents, because it is small (cf. Phillips, 1977), the energy flux to the ocean, Φ_{oc}, is given by

$$\Phi_{oc} = \Phi_{aw}^{tot} - \rho_w g \int_0^{2\pi} \int_0^{\omega_c} d\omega d\theta \; \left(S_{in} + S_{nl} + S_{diss}\right), \tag{11}$$

where Φ_{aw}^{tot} is the total energy flux transferred from air to ocean waves. This total energy flux is fairly well-known, because empirically the wind input to ocean waves is well-known, even in the high-frequency part of the spectrum (cf. Plant, 1982). Furthermore, there is now a consensus that the high-frequency part of the spectrum obeys an ω^{-5} power law (Banner, 1990; Birch and Ewing, 1986; Hara and Karachintsev, 2003, to mention but a few references). Hence, fairly reliable estimates of the energy flux Φ_{oc} may be provided by means of a wave model provided the model has a wind input term that agrees with the observations of wave growth and provided modelled wave heights compare well with observations.

Before results of time series for momentum and energy flux for a simple case are presented, we have to make one remark on the numerical implementation of (10) and (11). The energy balance equation is solved by means of an implicit integration scheme (cf. Komen et al, 1994). To be consistent with the numerical treatment of the energy balance, the momentum and energy flux have to be treated in a similar spirit, i.e. including the implicit factors of the integration scheme.

Let us now illustrate the sea-state dependence of the momentum and energy flux for the simple case of the passage of a front. To that end we take a single grid-point version of the ECMWF version of the WAM model and force the waves for the first day with a constant wind speed of 18 m/s, which is followed by a drop in wind speed to 10 m/s and a change in wind direction by 90 deg. In Fig. 8 we have plotted time series of atmospheric stress (τ_a), the momentum flux to the ocean (τ_{oc}), the total air-wave energy flux (Φ_{aw}^{tot}) and the energy flux into the ocean (Φ_{oc}). The momentum fluxes have been normalized by τ_a, while the energy fluxes have been normalized by $m\rho_a u_*^3$, with $m = 5.2$ which is a convenient mean value. During the first day we deal with the case of wind-generated gravity waves, hence windsea, and, in particular, the difference between atmospheric stress and the momentum flux to the ocean is small, most of the time at best 2%. This is a well-known property of windsea (JONSWAP, 1973). For windsea, the difference between total energy flux Φ_{aw}^{tot} and the energy flux into the ocean Φ_{oc} is somewhat larger. When the front passes at $T = 24hrs$ there is a sudden drop in wind, hence in atmospheric stress. However, the waves are still steep and experience an excessive amount of dissipation in such a way that wave energy decreases. As a consequence, considerable amounts of momentum and energy are dumped in the ocean column, much larger than the amounts one would expect from the local wind. Therefore, in cases of rapidly varying circumstances, the fluxes are seen to depend on the sea state. This is in particular true for the energy flux Φ_{oc} and to a much lesser extent for the momentum flux τ_{oc}.

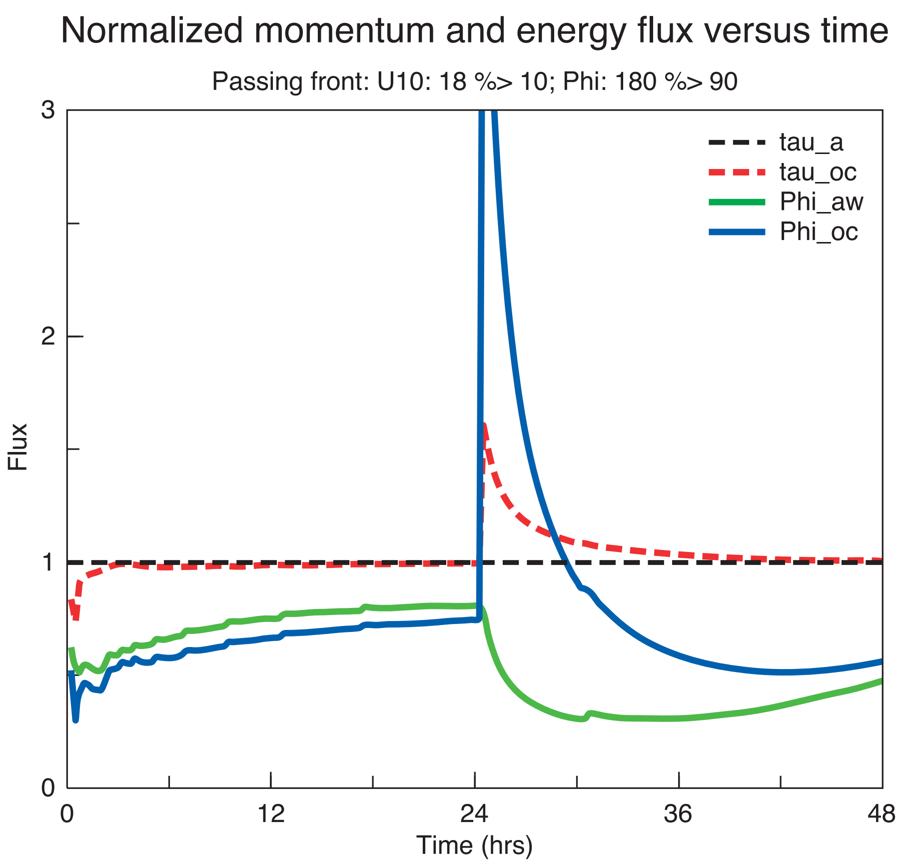

Figure 8: Evolution in time of normalized momentum flux and energy flux to the ocean for the case of a passing front after 24 hrs. The momentum flux has been normalized with $\rho_a u_*^2$, while the energy flux has been normalized with $m\rho_a u_*^3$, where $m = 5.2$.

This different behaviour of momentum flux and energy flux is caused by a combination of two factors. By definition momentum flux is mainly determined by the high frequency part of the spectrum while we have assumed that in the unresolved part of the spectrum there is a balance between wind input and dissipation. Hence, for windsea there is almost always a balance between atmospheric momentum flux and the flux into the ocean. This holds to a lesser extent for the energy flux because this flux is partly determined by the low frequency part of the wave spectrum as well.

We finally remark that in the work of Weber (1994) the energy flux into the ocean was approximated by the relation $\Phi_{oc} \simeq \langle c \rangle \tau_{aw}$, where $\langle c \rangle$ is the mean phase velocity. This generally overestimates the energy flux by at least a factor of two and as a consequence she finds fairly high values of m, $m \simeq 14$. In addition, in interesting cases such as the passage of a front the energy flux approximated in this manner will follow the wind. For example, in the frontal case of Fig. 8 the energy flux to the ocean would decrease dramatically at $T = 24$ hrs, while, in fact, it should hardly change. Therefore, it is not surprising that with this approximation the energy flux Φ_{oc} and wind are closely related.

4 Conclusions.

In this paper we have reviewed the impact of the sea state on atmospheric circulation, from the medium-range to seasonal forecasting time scales. An important finding is that with the recent increase of atmospheric resolution from T_l319 to T_l511 we have experienced a more pronounced impact of ocean waves on the atmosphere in the medium-range. The same remark applies to seasonal forecasting time scales. Apparently, a realistic representation of the small scales is important for air-sea interaction.

Furthermore, we have discussed possible benefits of sea state information for coupled atmosphere, ocean circulation modelling such as relevant for seasonal forecasting. These possible benefits have always been ignored by the ocean modelling community. This is surprising when it is realized from the physical point of view that the ocean surface layer is to a large extent controlled by the physics of breaking waves. This will have impact on the magnitude and direction of the surface drift, and hence on the atmospheric circulation and fluxes. In addition, the energy flux into the ocean is sea state dependent, and hence the thickness of the mixed layer. In particular, the mixed layer depth will be thin in areas where there is hardly any variability in the wind, such as in the Trade winds, whereas the mixed layer will be deep in the extra-Tropics.

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