341

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IMPACT AND FEEDBACK OF OCEAN WAVES ON THE ATMOSPHERE

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Abstract

The air-sea momentum transfer is known to be affected by the state of the ocean waves. For example, young wind waves are usually steeper than old wind waves and therefore extract more momentum from the air flow. As a result airflow over young windsea is rougher than over old windsea. The sea state dependent momentum transfer is thought to be of importance in rapidly varying circumstances occurring near depressions and fronts and in fetch limited circumstances near coasts. In this paper we review previous results regarding the impact of ocean waves on the atmospheric climate and we discuss new results regarding the impact of waves on the atmosphere for medium-range weather forecasting. It is concluded that the impact of ocean waves on the atmosphere depends in a sensitive manner on the resolution of the atmospheric model. In addition, analysis differences induced by the coupled wind-wave model give a considerable increase of the impact of ocean waves on the atmosphere. Overall, a modest positive impact on the scores of the atmospheric model is found while the improvements in the forecast performance of the wave model are even larger. As a consequence the coupled atmosphere, ocean-wave model was introduced into the ECMWF operational forecast system on June 29 1998. After this date a number of improvements of the wave model's advection scheme and integration scheme have been introduced and the angular resolution of the wave spectrum was doubled. Combined with the recent increase of spatial resolution of the atmospheric model from T_l319 to T_l511, this resulted in a dramatic increase of forecast skill of geopotential height and significant wave height, in particular in the Northern Hemisphere Summer time.

1. Introduction

Traditionally, many weather centres have applied their weather forecast to activities associated with the marine environment. Shipping, fisheries, offshore operations and coastal protection are, for example, all strongly dependent on weather and require marine weather forecasting extending to the limit of the medium-range forecasting period.

An important component of the marine weather forecast is the sea state. Wave forecasting using forecast low-level winds started in the mid-sixties. Although, due to lack of knowledge of the basic processes , the first ocean wave prediction models were certainly not perfect, it soon became evident that an important error source in forecast wave height was the wind speed error. This sensitive dependence of wave height results on the accuracy of the wind speed can be seen directly from the well-known empirical formula for the wave height in the special case of saturation under steady wind conditions. Denoting the significant wave height by H_s and the 10-metre wind speed by U_{10}, for saturated ocean waves we have the relation

$$H_s = \beta \frac{U_{10}^2}{g} \tag{1}$$

where g is the acceleration of gravity and β a constant. Eqn (1) follows from dimensional considerations and the assumption that the only relevant parameters in steady wind-generated, ocean gravity waves are g and U_{10}. Clearly, from (1) we infer that the relative error in wind sea wave height is twice the relative error in wind speed. It is therefore evident why ocean wave modellers have been critical regarding the quality of the generating wind field, often resulting in suggestions for improvement.

Ocean wave information can give benefits for atmospheric modelling and data assimilation. In particular, wave results have been used to diagnose planetary boundary problems and overactivity of the atmosphere, to obtain a consistent momentum balance at the ocean surface and to interpret satellite data. Firstly, this has resulted in an improved integration scheme for vertical diffusion (Janssen et al [l]), while the diagnoses of overactivity, as reflected by increased levels of kinetic energy during the forecast, provided the ECMWF modellers a norm to measure to what extent overactivity needed to be reduced (Janssen et al [2]). Secondly, the momentum exchange between atmosphere and the ocean surface could be treated more accurately if the wave-induced drag were taken into account in a two-way interaction (Janssen [3]); at present drag over the sea is modelled as a function of instantaneous wind speed only. Thirdly, the utilisation of satellite data from the scatterometer, Synthetic Aperture Radar (SAR) and Altimeter(i.e. winds) will benefit from both model generated and observed wave data for an optimal assimilation (Stoffelen and Anderson [4]; Hasselmann and Hasselmann [5]; Janssen, [6]). Finally, because of the strong interaction between wind and waves, wave information may be beneficial for the atmospheric state when assimilated into a coupled ocean-wave, atmosphere model.

In this paper we concentrate on the two-way interaction of wind and waves and we study its consequences for both atmospheric and wave model results. In Section 2 we briefly describe the physics of air-sea interaction which includes the role of the wave-induced stress in the process of air-sea momentum transfer and we compare the model of the sea-state dependent roughness with in-situ data. Next, in Section 3 we discuss the impact of the modified momentum, heat and moisture exchange on a few synoptic cases, where in particular we study the role of resolution of the atmospheric and wave model. Furthermore, the impact of ocean waves on the atmospheric climate during the Northern Hemisphere winter, studied by Janssen and Viterbo [7] is briefly reviewed. In Section 4 we describe the implication of results obtained at ECMWF for medium-range weather and wave forecasts. In these experiments both the analyses and forecasts were obtained with the coupled model and were compared with corresponding results from the uncoupled system. Although a number of data assimilation experiments were performed with the three-dimensional variational approach (3DVAR, cf. Andersson et al [8]), we mainly concentrate on the experiments with the four-dimensional variational (4DVAR, cf. Rabier et al [9]) assimilation scheme. We discuss differences in analysis and forecasts; considerable forecast differences in surface pressure and wave height are found in cases of rapidly developing lows. These differences have normally a small scale and, therefore, studying the scores we find a modest positive impact of two-way interaction on the atmospheric forecast model performance but there is larger improvement for wave height forecasts. The coupled atmosphere, ocean-wave model was introduced in operations on June 29 1998 and in Section 5 we discuss further developments of the wave model such as a revised integration scheme for the source functions and improvements to the advection scheme. We also increased the angular resolution of the spectrum by a factor of two. In the context of the new high-resolution atmospheric system the change in spectral resolution and the improved advection scheme has resulted in dramatic improvements of forecast skill of atmospheric and wave parameters. Finally, conclusions and a discussion of future developments are presented in Section 6.

2. Wind wave interaction

It has been known for some time that waves play an important role in the air-sea momentum transfer and it is of interest to address the consequences of the sea state dependence of the air-sea momentum transfer on the evolution of weather systems and on the atmospheric climate. During the last decade there has been considerable interest in the problem of the interaction of wind and ocean waves with emphasis on the sea state dependence of the momentum transfer across the air-sea interface. Experimental evidence has shown (see e.g. Donelan [10]; Smith et al [11]) that, especially for young wind sea (as occurs e.g. near fronts or near the coast for offshore winds) the airflow depends on the sea state. Comparing young and old wind sea cases, the drag coefficient may vary by a factor of two. This suggests that the interaction process of wind and waves should be regarded as a two-way process: the winds generate the waves which feedback on the atmospheric flow. Parallel to this development a theory of two-way interaction of wind and waves was developed, giving the sea state dependence of momentum, heat and moisture transfer both for pure wind sea and mixed wind sea, swell cases (Fabrikant [12]; Janssen [13, 3, 14]; Chalikov and Makin [15]). A parameterized version of the theory on sea-state dependent momentum transfer is included in cycle 4 of the WAM model (Komen et al [16]).

Since the late 1950's (Gelci et al [17]) it is known that the basic evolution equation for ocean waves is the so-called energy balance equation for the two-dimensional wave spectrum. Although the general structure of the source functions of the transport equation was presented more than 30 years ago (Hasselmann [18]) the early wave models did not attempt to compute the wave spectrum from first principles alone. The important role of nonlinear wave-wave interactions was not yet recognised. This changed by the 1970's when extensive wave growth experiments (Mitsuyasu [19, 20]; Hasselmann et al [21]) revealed the important role of the wave-wave interaction in controlling the shape of the wave spectrum. At that time a numerical implementation of the nonlinear interactions was computationally far too expensive, and thus wave modellers resorted to a simple parametrization which worked satisfactorily for locally-generated wind sea but was known to have defects in mixed wind sea-swell situations. The weakness of this approach was most pronounced in the case of strong and rapidly-varying winds such as occur in hurricanes. As a result, a large international group of scientists (the so-called WAM = Wave Modelling Group) decided to develop a new wave model based on first principles, without relying on the somewhat artificial distinction between wind sea and swell.

The foundations of this new model were laid by Klaus and Susanne Hasselmann at the Max-Planck Institut für Meteorologie (Hamburg), where the deep water version was developed. Since then many members of the WAM group have also contributed. For a complete account see Komen et al [16] where a thorough discussion on the energy balance equation, the physical foundations, the numerical aspects, verification of modelled results and assimilation of satellite data is given.

This newly developed model, called the WAM model, is based on the energy balance equation for the two-dimensional wave (surface variance) spectrum $F(f, \theta; \boldsymbol{x}, t)$, which is a function of wave frequency f, direction θ, position **x** and time t,

$$\frac{\partial}{\partial t}F + \nabla \cdot (\boldsymbol{c_g}F) = S_{in} + S_{nl} + S_{ds} \qquad (2)$$

Here, $\boldsymbol{c_g}$ is the group velocity of the surface waves. The source functions S describe different physical processes: S_{in} represents the direct atmospheric input through the surface wind, S_{nl} represents resonant four wave interactions (a process which conserves total energy and momentum) and S_{ds} is the dissipation term. Mean parameters such as wave height are derived as integrals of the wave spectrum. Operational running of the WAM model was only possible after an efficient algorithm for the computation of the nonlinear transfer became available (Hasselmann et al [22]) and when a reliable parametrization of the dissipation source

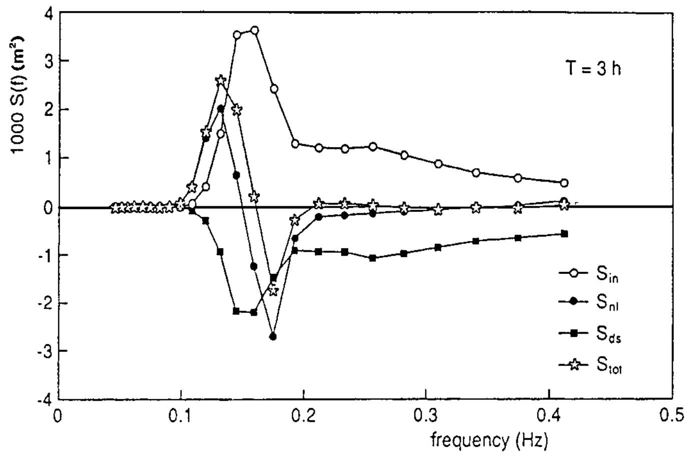

Figure 1: Energy balance for young windsea

function was found (Komen et al [23]). In addition, the determination of the surface stress (which is largely controlled by the wave-induced stress) was possible only after the work of Janssen [3, 14].

The WAM model does not impose any ad-hoc constraints on the spectral shape; the time evolution of the wave spectrum is computed by integrating the energy balance eqn (2). In order to appreciate the role of the respective source terms, in Fig 1 we have plotted the WAM model's directionally averaged source functions S_{in}, S_{nl} and S_{diss} as function of frequency for young wind sea 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 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. These results are in agreement with the findings from the Joint North Sea WAve project (JONSWAP, Hasselmann et al [21]).

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 (hence they control 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. However, ocean waves are in general not in an equilibrium state determined by a balance of the three source functions, because advection and unsteadiness are important as well. As a rule of thumb, of the amount of wave energy gained by wind about 90% is lost locally to the ocean by wave breaking, while the remaining 10% is either advected away or is spent in local growth. Fig 1 also shows a plot of the total source function, suggesting that, for young wind sea, there may be a considerable imbalance, in particular for the low-frequency waves.

It would be of considerable interest to develop a coupled atmosphere-ocean circulation system where the ocean waves are the agent that transfer energy and momentum across the air-sea interface in accordence with the energy balance equation. A first attempt towards this goal was reported by Bao et al [24], who studied in the context of a fully coupled model the development of hurricanes in the Gulf of Mexico. In this review we shall concentrate on just one aspect of the overall problem, namely the mutual interaction between wind and waves

The latest version of the WAM model (WAM Cy4) has been operational at ECMWF since June 1992.The first operational WAM model had a spatial resolution of 3 degrees, while the spectrum had 25 frequencies and 12 directions. No shallow water effects were included. The initial state for the wave forecast was constructed by forcing the WAM model with analysed ECMWF winds, and no wave data were used. Since August 1993 ERS-1(European Remote Sensing Satellite) altimeter wave height data are assimilated by means of an Optimum Interpolation Scheme (Lionello et al [25]). In July 1994, spatial resolution was increased by a factor of two while in December 1996 spatial resolution was further increased by a factor of three. Therefore, the present operational model has effectively a resolution of 55 km, while also shallow water effects such as bottom friction are switched on. In order to avoid too small advection time steps we changed from a regular spherical grid to an irregular spherical coordinate system in such a way that the distance between grid points is fixed. Hence near the poles we avoid violation of the CFL criterion while also a savings of the total number of grid points (and hence computation time) of 30% is achieved. The advection terms in the energy balance equation are evaluated in regular spherical coordinates obtaining the upwind energy fluxes from the nearest grid point on the irregular grid. In May 1996 we adopted the growth limiter of Hersbach and Janssen [26].

Validation results over the year 1995 are summarized in Janssen et al [27]. This study showed that the ECMWF wave forecasting system has reasonably good skill at that time, but when wave forecasts are compared to their verifying analysis a significant, positive systematic error is found in the later stages of the forecast. This systematic error is in particular a problem in the Tropics and the Southern Hemisphere where it can amount to 10% of the mean wave height.

In the ECMWF atmospheric model, the roughness length z_o of turbulent air flow over the ocean was until 1998 parametrized by means of the well-known Charnock relation (Charnock [28])

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

where α is the Charnock parameter, g is the acceleration of gravity and u_* the friction velocity. As a result, the drag coefficient of the air flow increases with increasing wind speed. This is in sharp contrast to turbulent air flow over a smooth surface, where the drag coefficient decreases with increasing wind speed. Clearly, the difference in behaviour of the drag coefficient must be due to the momentum transfer to the ocean waves, including the unresolved short gravity waves. Prescribing these unresolved gravity waves by a Phillips spectrum (Phillips [29]; Banner [30]), the rate of change of wave momentum P due to wind is directly related to the so-called wave-induced stress τ_w, or

$$\tau_w = \int df\, d\theta \frac{\partial}{\partial t} P \tag{4}$$

where the wave momentum depends on the two-dimensional wave-variance spectrum F(f, θ) (which is obtained from the energy balance eqn (2)),

$$P = \rho_w g F(f, \theta)/c \tag{5}$$

with c the phase speed ω/k of the waves, k is the wave number, ω is the angular frequency and ρ_w the water density. By means of the energy balance eqn (2) one may write (4) as

$$\tau_w = \rho_w g \int df d\theta S_{in}/c \tag{6}$$

where S_{in} represents the wind input term which is proportional to the 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. Since it is known that the high-frequency part of the wave spectrum depends on the sea state (for example, young waves are steeper than old waves) it follows that the wave-induced stress depends on the sea state. Therefore, the Charnock 'constant' α should depend on the sea state as well. Based on the so-called quasi-linear theory of wind-wave generation (Janssen [13]), which determines the slowing down of the wind because of the air-sea momentum transfer, Janssen [14] found the following dependence of the Charnock parameter α on the wave-induced stress τ_w

$$\alpha = \beta\left(1 - \frac{\tau_w}{\tau}\right)^{-\frac{1}{2}}, \beta = 0.01 \tag{7}$$

where $\tau = \rho_a u_*^2$. In the absence of surface gravity waves ($\tau_w = 0$), Eq (7) parametrizes the momentum loss due to processes which have not been considered so far, for example the growth of capillary waves or the interaction with currents, while in the presence of gravity waves ($\tau_w \neq 0$) the air flow becomes rougher because these waves also extract momentum from the air.

The validity of eqn (7) to determine the surface stress has been checked by Janssen [31] using HEXOS(Humidity EXchange Over the Sea) data (Smith et al [11]). During HEXOS, the wind speed U_{10}, friction velocity u_* and the one-dimensional frequency spectrum were measured simultaneously. Using the wind input term S_{in} from the WAM model the wave-induced stress may therefore be determined and for a known observed wind speed U_{10} the friction velocity u_* is obtained by solving

$$u_* = C_D^{\frac{1}{2}} U_{10}, C_D = (\kappa/\ln(10/z_o))^2 \tag{8}$$

iteratively. Here z_o is given by eqn (3) with Charnock parameter eqn (7) and κ is the von Karman constant. The comparison of modelled and observed friction velocity showed a good agreement.

Because of short fetch and/or short duration, the HEXOS data set and Donelan's data set consists of cases of relatively young wind sea which have steep waves and thus a high wave-induced stress. As a consequence relatively high Charnock parameters are obtained. On the open ocean, however, wind seas have long fetch and/or duration and they interact with swells generated by winds in remote areas. Under those circumstances ocean waves are expected to be less steep with relative low values of wave induced stress and therefore low values of the Charnock parameter are obtained. This is shown in Fig 2a where, for two areas namely the North Sea and the Western Atlantic area, we present the relation between the mean Charnock parameter α

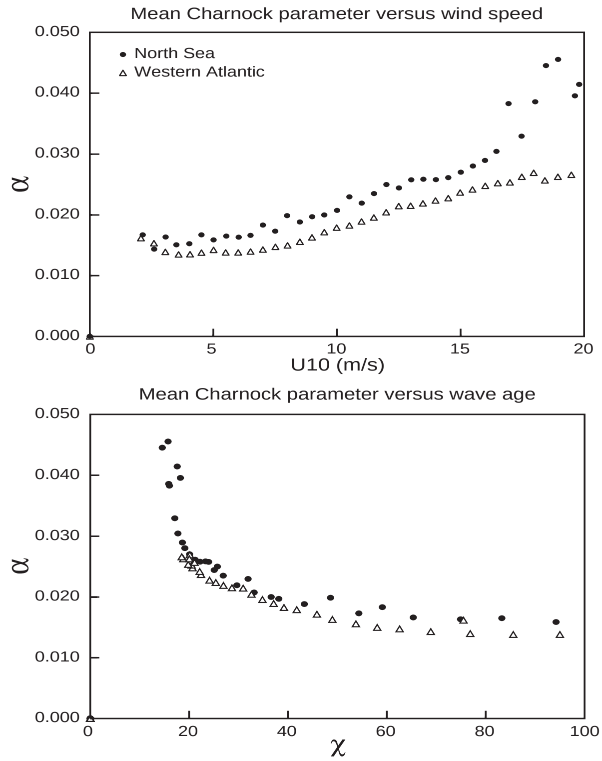

Figure 2: Charnock parameter as function of a) surface wind and b) wave age for the North Sea and the Western Atlantic over a 10-day period in November 2000. The Charnock parameter is averaged over .5 m/s wind speed bins.

(averaged over 1 m/s wind speed bins) and 10 m wind speed for the first 10 days of November 2000. The Charnock parameter was obtained from the coupled WAM-ECMWF model, details of which will be given shortly. It is clear from Fig 2a that the relationship between Charnock parameter and wind speed depends on the geographical location. In the western Atlantic the Charnock parameter is systematically lower compared to the North Sea, presumably because of the relative abundance of swell in the Western Atlantic, which through nonlinear transfer and the quasi-linear damping has a calming effect on the wind-generated waves that determine the wave-induced stress.

Therefore, Fig 2a suggests that there is no unique relation between Charnock parameter and surface wind speed. Normally, one searches for a relationship between the Charnock parameter and dimensionless wave parameters. In our approach eqn (7) suggests that the normalised wave-induced stress is the key variable to choose, but in practice the wave stress is hard to determine. For fetch- or duration limited wind sea, it is known (JONSWAP [21]) that the seastate, and therefore the wave stress, is determined by the wave age $\chi = c_p/u_*$, with c_p the phase velocity of the waves at the peak of the spectrum and u_* the friction velocity. It is therefore tempting to use the wave age as a proxy for the wave-induced stress, even for mixed windsea-swell conditions. In Fig 2b we have plotted for the same cases as in Fig 2a the mean Charnock parameter as function of mean wave age and it is evident that the wave age is a more appropriate parameter to characterise the roughness of the ocean in agreement with the findings of Donelan [10] and HEXOS (Smith et al [11]).

Nevertheless, there is still an ongoing debate whether the wave age is an appropriate parameter to characterize the dependence of the drag of air flow over the oceans on the sea state. For pure windsea, Donelan et al [32] find a relation between the enhancement of the drag coefficient and a measure for the sea state, namely the wave age. However, in general the sea state on the open ocean is confused because it consists of a mixture of windsea and swell. Therefore, a characterisation of the sea state in terms of wave age may not be a viable option. Thus, alternatives to the wave age parameter have been proposed (Monbaliu [33]; Anctil and Donelan [34]; J Janssen [35]). These proposals range from the use of a measure of the high-frequency part of the spectrum (closely related to the mean square slope) to the use of the orbital motion of the waves. Referring to Fig 2a-b, on the other hand, suggests that according to the coupled ECMWF-WAM model the choice of the wave age seems to be realistic after all.

Finally, although there is clear evidence that under fetch limited conditions the aerodynamic drag depends on the sea state, it is presently not understood why for the open ocean Yelland et al [36] were not able to detect a sea state dependence of the drag coefficient, while Banner et al [37], Eymard et al [38] and Hare et al [39] did find a significant increase of the Charnock parameter with increasing wind speed which is consistent with a sea state dependence of the drag. A more detailed discussion may be found in the Comments by Taylor and Yelland [40] and the Reply by Janssen [41].

From all this it may be concluded that, depending on the sea state, the drag coefficient may vary considerably keeping the surface wind speed fixed. Therefore, one may wonder whether two-way interaction has impact on the evolution of an individual depression, on medium-range forecasting in general and on the mean atmospheric circulation.

In the following, impact of ocean waves on the atmospheric model will be studied by means of a coupled atmospheric, ocean wave model. The coupling scheme is a simple one: ocean waves are driven by atmospheric winds, while the sea-state dependent slowing down of these winds is parametrized by the Charnock parameter given in eqn (7).

3. Impact on synoptic cases and on climatology

3.1 Impact on a single depression

Before we discuss numerical experiments investigating the impact of the sea-state dependent drag on the atmosphere we first would like to give a simple theoretical picture of the interaction of ocean waves with the large-scale atmosphere.

On average the time scale of impact of waves on the atmospheric circulation is of the order of 5 days. This time scale of impact is related to the decay of vorticity by surface friction. Using the quasi-geostrophic approximation (Pedlosky [42] the rate of change of vorticity may be related to the curl of the surface stress, while the surface stress is related to the geostrophic wind by means of an empirical geostrophic drag law. As a result, the decay law due to surface friction for the large-scale vorticity ζ becomes

$$\frac{\partial}{\partial t}\zeta = -\gamma\zeta, \quad \gamma \cong C_{Dg}U/\mathrm{D} \tag{9}$$

where U is the geostrophic wind speed, D is the equivalent depth and C_{Dg} the geostrophic drag coefficient,

$$C_{Dg} = [\kappa/\ln(0.2u_*/fz_o)]^2 \qquad (10)$$

with f the coriolis parameter. Note that according to the decay law (9) the impact of surface friction on the atmosphere is of a barotropic nature since the whole atmospheric column is affected. At this point it is important to argue why young wind sea is relevant at all for the evolution of a depression, since, as seen from the waves, the lifetime of young windsea is only 6-12 hours. However, for an observer moving with a depression the young windsea state may last several days or even longer, depending on its speed and curvature. For a geostrophic wind speed of 25 m/s and an equivalent depth of 5 km one finds for young wind sea (Charnock parameter α =0.09) a damping rate of 0.4 days^{-1} while, for old wind sea (α =0.018), a damping rate of 0.3 days^{-1} is found. Compared to growth rates of the baroclinic instability without surface friction (these are typically 0.6 days^{-1}) it is seen that the damping rates due to surface friction are a considerable fraction of the total growth rate of the vortex (depression). Different sea states may therefore give rise to differences in the total growth rate (which is about 0.1-0.2 days^{-1}) of the order of 50% so that significant differences occur over a timescale of 5 days. Because of this long timescale it should be noted that this simple picture of the impact of surface friction on the decay of a depression may be obscured by nonlinear wave-wave interactions and wave-mean flow interaction. In addition, it should be realized that enhanced surface roughness not always will lead to a decay of the depression, because heat fluxes are enhanced as well resulting in vortex stretching and therefore in a deepening of the low. Thus, whether due to the enhanced roughness (as caused by the ocean waves) there is filling up or a deepening of a low depends on whether momentum or heat fluxes determine the evolution of that particular low.

The impact of two-way interaction on the evolution of a single depression was studied by Doyle [43]. The atmospheric model used in this study is the Navy's Coupled Ocean Atmospheric Mesoscale Prediction System (COAMPS) and the wave model is the WAM model coupled through eqn (7). The three-dimensional model solves the compressible equations of motion. The model was used in a channel mode with the f-plane approximation. In the vertical the model has 32 layers with greater resolution in the lower troposphere (13 layers below 850 mb) to enable a good representation of the interaction of the wave-induced stress with the marine boundary layer.

For both atmospheric and WAM model the horizontal resolution is 30 km with periodic conditions in the zonal direction. A timestep of 90 s is used in the atmospheric model and 6 min for the ocean wave model. The simulation is integrated to 96 h which enables the study of the rapid development and early decay phases of an idealised cyclone.

The initial condition consists of a zonally symmetric jet stream based on mean winter conditions, where the initial temperature distribution is found from thermal wind balance. The initial moisture field is based on climatology as well. The initial perturbation of the baroclinically unstable jet was provided by a small longitudinal wind component below 250 mb, specified as a sinusoidal function that damps near the boundaries. The sea surface temperature pattern resembled the Gulf Stream pattern.

The sea level pressure at 60 hrs for the control run (i.e. constant Charnock parameter = 0.0185) and the coupled run (i.e. two-way interaction) is presented in Fig 3. It clearly shows that, due to the increased roughness in the coupled run, the central pressure of the low is 6 mb higher compared to the control run. The time evolution of the central pressure (not shown) confirms this picture. The increased roughness in the coupled run is accompanied by an increase in sensible heat flux of about 20%, the rainfall maximum increases by 34%, while the kinetic energy at the surface decreases by 20%. These results indicate that frictional effects of wind-generated ocean waves may influence the boundary layer structure in the vicinity of a marine cyclone.

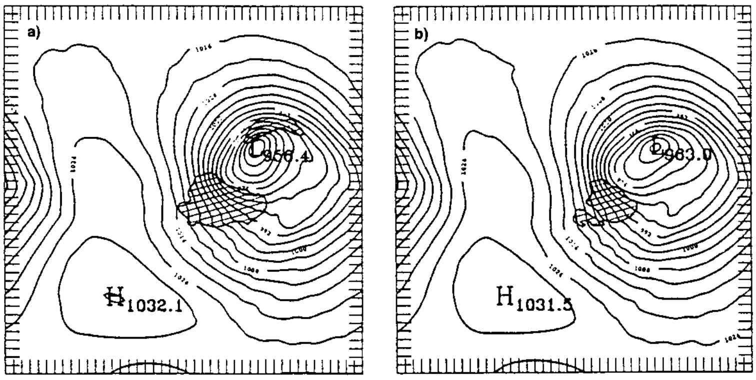

Figure 3: Simulated sea-level pressure for a) control and b) coupled simulations for the 60 h time. The isopleth interval is 4 mb. Regions of lowest model layer wind speed in excess of 25 m s^{-1} are denoted by shading. Tick marks are plotted along the borders every third grid point or 90 km. (From Doyle [43], with permission.)

Doyle's results are encouraging and do suggest that in extreme conditions the sea state dependent roughness makes a difference for the evolution of a low. On the other hand, Lionello et al [44] and Lalbeharry et al [45] found a reduced sensitivity of the development of (synthetic) lows on the sea state, while Bao et al [24] found in contrast to the previous works a deepening of a Gulf of Mexico hurricane by 6 mb.

There may be several reasons for these discrepancies. First, the Lionello et al [44] and Lalbeharry et al [45] atmospheric simulations had a coarse resolution, which limits the intensity of the low and therefore reduces the coupling between the ocean waves and the atmosphere. In addition, while all these authors used the same wave prediction system, the atmospheric models were clearly different. A different treatment of the physical processes in the boundary layer will undoubtedly lead to differences in sensitivity to the sea state. Moreover, most atmospheric models have horizontal diffusion in order to ensure numerical stability for feasible integration time steps. A side effect of horizontal diffusion is, however, a reduced level of activity in scales (say of the order of 200 km) that are relevant for the interaction of wind and waves. Finally, in the Bao et al [24] experiment, and also in the Lalbeharry et al [45] one, sensible and latent heat fluxes have the same sensitivity to the sea state as the momentum flux. Although normally (cf. Janssen and Viterbo [7], Lionello et al [44]) the dependence of heat flux on the sea state plays a minor role in the development of a low, under conditions where hurricanes develop air-sea temperature differences may be large. In that event increased roughness by the waves may result in enhanced heat fluxes giving vortex stretching and therefore a deeper low. However, it should be noted that field observations from HEXOS (DeCosmo [46]) do not support the sensitive dependence of heat fluxes on the sea state and or wind speed as chosen by Bao et al [24] and Lalbeharry et al [45].

To summarize the above discussion it seems clear that in order to obtain impact of the ocean waves on the atmospheric circulation a high spatial resolution atmosphere and wave model is required which has realistic levels of activity at scales of a few hundreds of kilometres and which has a realistic parametrisation of the physical processes in the boundary layer. In addition a sufficient number of layers in the vertical are needed to give a proper description of the physics in the boundary layer. For most operational, global weather forecasting systems these choices are not yet feasible, except perhaps for the ECMWF model which presently has a spatial resolution of 40 km and 60 layers in the vertical. Therefore it is of interest to study the impact of ocean waves on the atmospheric circulation using this state of the art operational weather prediction model.

A scheme to couple the WAM model and the ECMWF model was developed, which was similar in spirit to those developed by Weber et al [47] and Doyle [43] (see also Janssen [48]). The atmospheric model and the wave model are interacting with each other every coupling time step, i.e. the atmospheric model is run for a coupling time step, then the wave model is called using the lowest level winds produced by the atmospheric model after which the atmospheric model is run for the next coupling time step, now using the Charnock parameter (α) as determined by the WAM model during the previous interaction period. A schematic view is given in Fig 4. In the present set-up, following Beljaars [49], neutral exchange coefficients for momentum, heat and moisture differ from each other. This means that compared to momentum transfer, heat and moisture transfer have a reduced sensitivity to the sea state because they depend on the square root of the drag coefficient. The extended range integrations on which we report in Section 3.2 had identical exchange coefficients but we checked that this assumption had hardly any influence on the results for the impact on the atmospheric climate (in agreement with remarks of Lionello et al [44]).

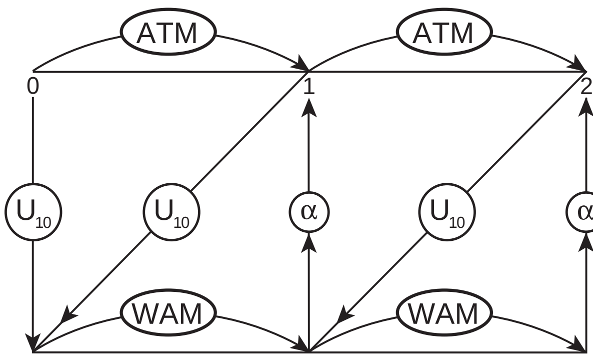

Figure 4: Schematic of the coupled system.

In the remainder of this paper we shall discuss the impact of two-way interaction on the atmospheric circulation. We measure this impact by comparing results of two-way interaction with results of one-way interaction experiments. The one-way interaction experiments are performed with the same software as the two-way experiments except that in the one-way experiment, which will be called the control from now on, the Charnock parameter takes the constant value 0.018, while in the two-way experiment (coupled for short) the Charnock parameter is determined from the wave model, thus giving a consistent energy balance at the surface.

At weather centres impact of a change in the forecast system is usually determined by a comparison of forecast scores of the experiment and the control. The forecast scores are obtained by comparing forecast

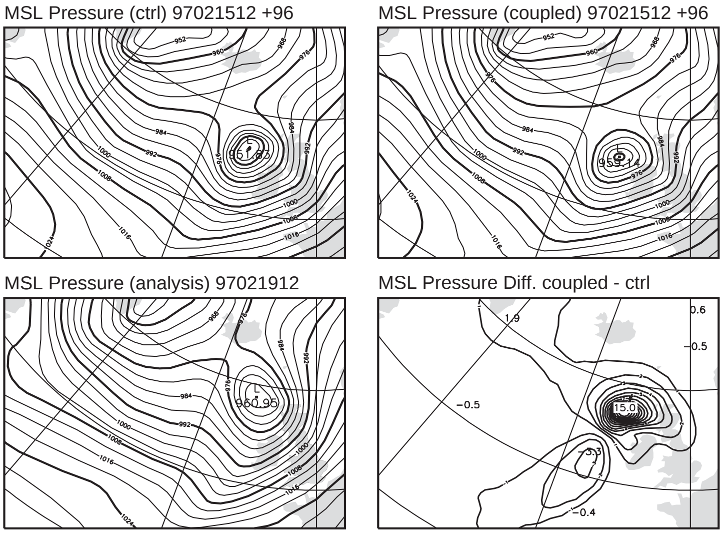

Figure 5: Comparison of 4 day forecast of surface pressure over the North Atlantic for 15 February 1997. Top left panel: control, top right panel: coupled, bottom left panel: operational analysis, bottom right panel: the difference between coupled and control. Version of coupled model is T213/L31-0.5D.

results with the corresponding verifying analyses and by plotting forecast scores such as systematic error, rms error and anomaly correlation as function of forecast time. Note that the anomaly correlation measures to what extent a forecast is beating climatology and it is defined as the correlation between the forecast anomaly and the analysis anomaly (both with respect to the climate). However, in case of positive correlation, when the forecast beats the climatology, one does not necessarily have a forecast that is regarded as useful by the forecaster. The relation between usefulness of the forecast and anomaly correlation is not straightforward and can only be established by means of the subjective step that forecasters assess by inspection of weather maps whether, for a given anomaly correlation, a forecast is useful or not. In this way weather forecasts were found to be useful when the anomaly correlation exceeded 60%. We have adopted a similar approach to the study of the impact of changes in the wave forecasting system.

In order to distinguish between different versions of the coupled code we introduce the following notation: Txxx/Lyyy-z.zD denotes the coupling of an atmosphere model, having a triangular Truncation of Txxx and yy vertical Levels, with a wave model which has a resolution of z.z Degrees. Most of the early results discussed in this paper are obtained with the T213/L31-1.5D version of the coupled system while we also discuss results from the T213/L31-0.5D version.

We have performed an extensive set of experiments with the coupled system. This ranges from experiments investigating interesting cases of rapid development of lows to tests where the coupled system is used both in the analysis and forecast. Results from the analysis and forecast experiments will be discussed in the next Section, while here we focus on experiments where the coupled system is only used in forecast mode, therefore control and experiment start from the same initial conditions.

We tested the forecast performance of the coupled system with its T213/L31-0.5D version by running an extensive set of 10 day forecasts on 'randomly' selected cases. As initial conditions the 15th of every month in the period of September 1996 until August 1997 was chosen. The control results were obtained with cycle 18R1 of the ECMWF atmospheric model.

The initial data for the atmospheric fields were taken from the analysis provided by the atmospheric model which was operational at the start of the forecast. The initial data of the WAM model were generated by using as initial condition a so-called JONSWAP spectrum 10 days before the starting date of the forecast and by running the wave model until the starting date of the experiment with analysed winds from the ECMWF archive.

Comparing scores of the anomaly correlation of 1000 and 500 mb height of the coupled and control runs we found hardly any differences, except over the North Pacific where the coupled run showed a small benificial impact of ocean waves on the atmosphere (not shown). Nevertheless, considerable synoptic differences between the coupled and control run are found occasionally. An example is the 4 day forecast from 15 February 1997, shown in Fig 5. This low was observed during FASTEX(Fronts and Atlantic Storm-Track EXperiment) and was called IOP 17. The left top panel of Fig 5 shows the surface pressure over the North Atlantic for the control run, the top right panel shows the coupled results, the bottom left panel shows the operational verifying analysis, while the bottom right panel shows the difference between coupled and control. The control day 4 forecast has a good quality, as may be judged from the comparison with the analysis although the fast moving low west of Scotland is misplaced and too deep by about 9 mb. The anomaly correlation over the North Atlantic area was larger than 90%. In the coupled run it is seen that the low near Scotland is less deep by 8 mb so that in this respect there is better agreement between the coupled run and the verifying analysis. However, because the differences are small scale and the scores are evaluated on a coarse grid with a resolution of 2.5 deg., there is hardly any difference in anomaly correlation between coupled and control run.

It is emphasized that these results depend in a sensitive manner on the resolution of the coupled system. To appreciate this point we show in Fig 6 results of the T106/L31-1.5D model for the same case. While in the high resolution run we notice differences of up to 15 mb, the low resolution results show only differences of at

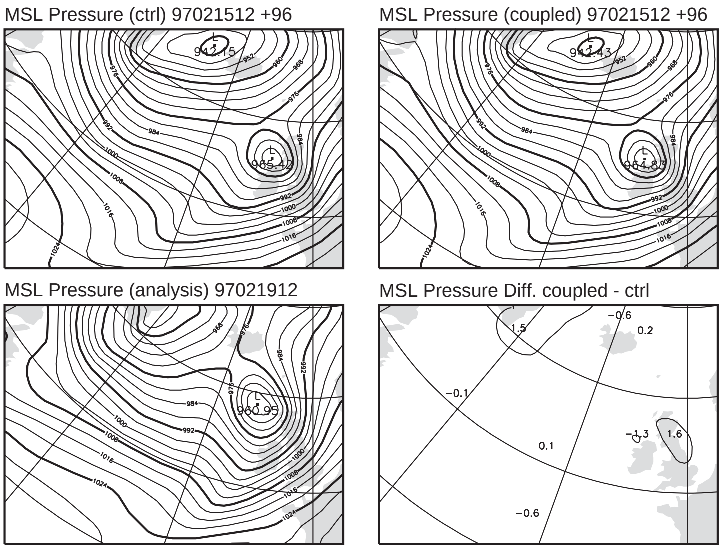

Figure 6: Idem as Fig 5 but now for T106/L31-1.5D version.

most 2 mb. A similar sensitive dependence on resolution was also noticed in other extreme events. An example, not shown here, is the so-called Braer storm of 11 January 1993.

3.2 Seasonal Integrations

We have seen that two-way interaction may have impact on synoptic cases and the question of interest is whether there is any systematic impact of waves on the atmospheric circulation. Hence this requires a study with the coupled model over time scales of a season. To that end, a number of 120 day runs with the T63/L19-3.0D version of the coupled model were performed and the mean over the last 90 days was compared with results from the model with one-way interaction only. We concentrate here on the winter season of 1990. In order to obtain reliable information on the impact of waves on the atmospheric circulation there is a need for ensemble integrations (see e.g. Ferranti et al [50]). The variability in the Northern Hemisphere is high, especially over the oceans. Therefore 15 coupled and control runs were performed for the winter season of 1990 starting from the analysis of 15 consecutive days, thus providing a reliable estimate of the meanstate of the atmosphere at a certain location. At the same time, information on the variability may be inferred from the scatter around the mean, thus a student t-test may be applied to test statistical significance of the mean difference between coupled and control run.

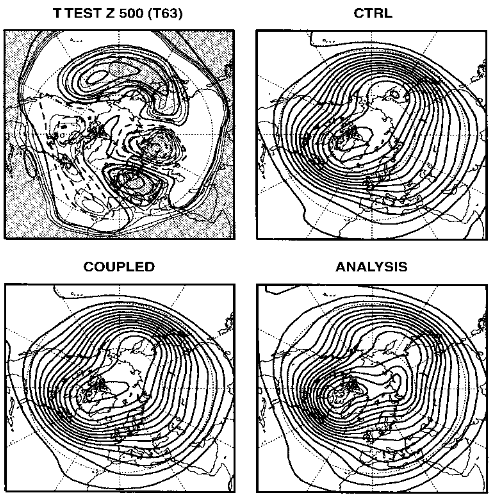

Figure 7: Ensemble mean of 500 mb geopotential height of coupled and control run and their differences. For comparison also the analysed climate is shown. Period is winter 1990 and area is Northern Hemisphere.

Results are discussed in detail by Janssen and Viterbo [7]. Here we only describe features in the 500 hPa height field. In Fig 7 we have plotted the ensemble mean of 500 hPa 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. Fig 8 shows a similar plot for the Southern Hemisphere. Contours in the mean 500 hPa height field are plotted every 60 m, while in the difference plot we have indicated by dark 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 in both hemispheres.

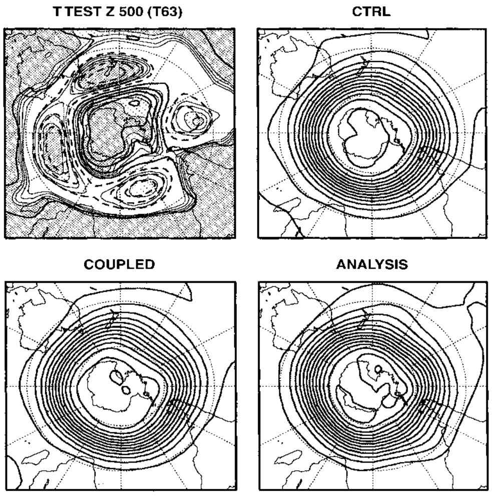

Figure 8: Same as Fig 7 but now for Southern Hemisphere

In the Northern Hemisphere we note differences over the Northern Pacific, Europe and Siberia. In the last two areas the coupled climate shows, when compared to the analysed climate, a considerable improvement. This is, however, unclear for the Northern Pacific. There are also differences in the low and high frequency variability (not shown). Over the North Atlantic a small shift in storm track is to be noted while the low

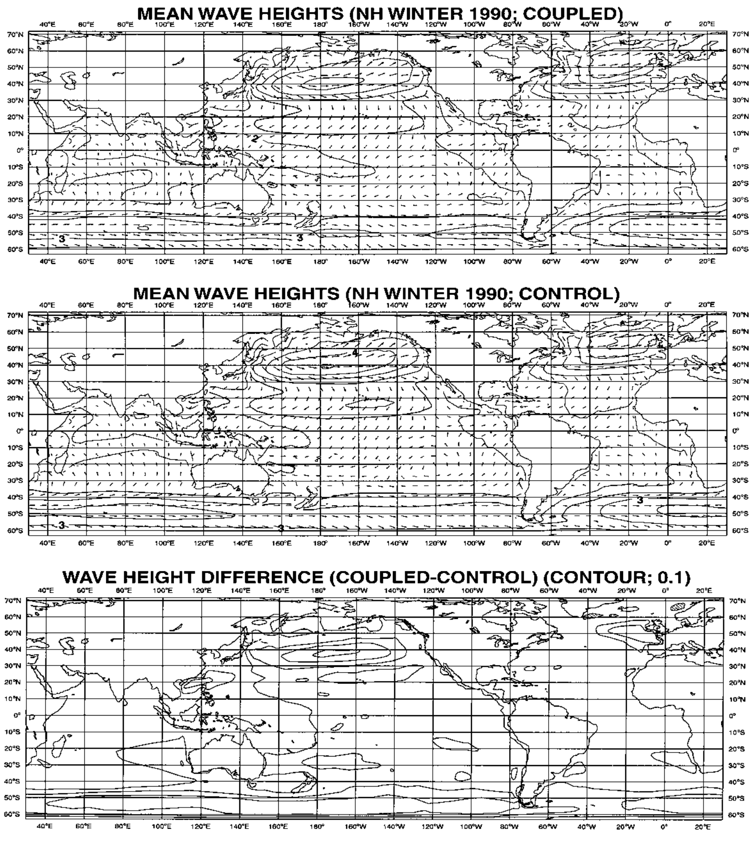

Figure 9: Ensemble mean of wave height for coupled and control run and their differences.

frequency variability over Europe drops by 25%. The impact of waves in the Southern Hemisphere appears to be large scale; the storm track has, in the coupled climate, a more "wavy" character. East of New Zealand the coupled climate shows improvement when compared to the analysed climate. It is also of interest to show the ensemble mean of wave parameters. Fig 9 shows the mean wave height for coupled run, control run and the difference. Differences are of the order of 20-30 cm which is about 10% of the mean wave height; they are large-scale, especially in the Southern Hemisphere, with generally a reduction in wave height. For the Southern Hemisphere (between 20^{o}S and 60^{o}S) the mean bias between coupled and control runs is -20 cm. Although this is not sufficient to alleviate the systematic error problem noted in the 10 day forecast verification (Section 2), it certainly works in the right direction. The reason for the lower wave height in the coupled run is a slowing down of the air flow caused by enhanced roughness during young sea states.As far as the 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 [7] also performed seasonal forecasts with the T21 version of the coupled system and in particular 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 big 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. A more detailed discussion on this issue may be found in Janssen and Viterbo [7].

4. Medium-range forecasting

In this Section we discuss some results from experiments performed at ECMWF where both the analysis and forecasts are obtained with the coupled model and we compare with corresponding analyses and forecasts of the uncoupled system. Hence, the atmospheric model was modified in the manner described in the previous Section to allow for two-way interaction between atmosphere and waves while also the analysis suite was changed in order to modify the first-guess in a manner consistent with the coupled physics. The early experiments were done with the 3DVAR assimilation system, and in the first-guess step it is then possible to assimilate Altimeter wave height data. Later, the coupled system was upgraded by using the 4DVAR assimilation scheme; in that event Altimeter data are assimilated in the final trajectory calculation.

The initial results were obtained with the T213/L31-1.5D version of the coupled system. The atmospheric model cycle was CY16R2. We performed analyses and forecasts over two periods of 21 days in February 1997 and July 1996. Altimeter data were not used in the analysis, so they may give an independent check on the quality of the wave height field. The next step was to upgrade the coupled system by increasing the wave model resolution from 1.5 deg to 0.5 degree and to allow for the assimilation of Altimeter data. Using CY16R3 of the atmospheric model we redid the first 15 days of February 1997. In all these experiments the assimilation scheme was 3DVAR. Finally, we combined the 4DVAR assimilation scheme and the coupled system and as a technical check we did 10 days of analyses and forecasts over a period in December 1997. A discussion of the quality of the coupled analysis and the forecast performance of the coupled system is given in Janssen et al [51]. Here, we briefly summarize the highlights

4.1 Discussion of early results

By visual inspection of synoptic charts the analysed 1000 mb geopotential height from the coupled and control run were compared. In the Northern Hemisphere we found differences in surface pressure of at most 1 mb while in data void areas such as the Southern Atlantic differences were occasionally more substantial, of the order of 5 mb or more . In view of the sensitive dependence of the atmospheric model to changes in its initial condition, these relatively small changes in analysed surface pressure may result in substantial differences during the forecast. Thus, compared to the 'randomly' selected forecasts that use the same initial conditions (cf. Section 3.1), the data assimilation experiments have differences between coupled and control that are typically larger by a factor of two.

As already pointed out, two-way interaction of wind and waves may either deepen or fill an atmospheric low. In addition, the coupling between wind and waves affects the first guess fields in the assimilation cycle, and it is not clear beforehand how this change in initial condition affects the fate of a low.

As a first synoptic example we discuss the impact of the sea state dependent roughness on the rapidly moving low from FASTEX which was discussed already in Section 3. This event started just south of New Foundland and arrived two days later west of Scotland. The day 2 forecast of this case is shown in Fig 10 and the surface

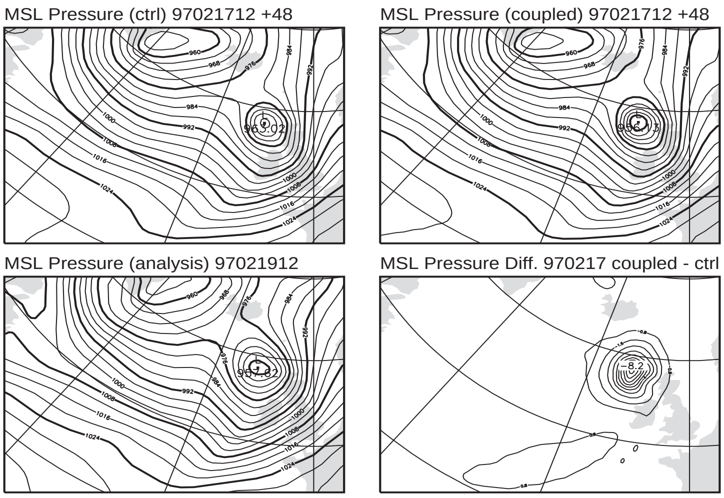

Figure 10: Comparison of 2 day forecast of the FASTEX IOP-17 event from control (top left panel) and coupled (top right panel) experiment. Differences between coupled and control are shown in the bottom right panel while the verifying analysis is from the coupled experiment. Date is 17 February 1997.

pressure in the coupled run is lower by 8 mb, in good agreement with the coupled analyzed pressure of that low. Such differences in surface pressure result in considerable differences in the strength of the surface wind and therefore also in wave height. In this case the maximum wave height increased from 9 to 13m. However, because of the small scale of the differences, there is hardly any change in anomaly correlation over the North Atlantic area; in fact, at day 2 both coupled and control experiment have anomaly correlations close to 100%.

The next example discussed here concerns the day 2 forecast from 24 December 1997 for the North Pacific. It is an example of large scale impact of two-way interaction on the atmospheric circulation. In this case the resolution of the wave model was 0.5 deg. while the analysis was performed with 4DVAR. Fig 11 shows the comparison of the coupled day 2 forecast with the control forecast and the control analysis. Substantial large scale differences in the surface pressure can be seen, and a better agreement between coupled forecast and analysis is noted. As a result, considerable improvements in the anomaly correlation for surface pressure were obtained for the whole 10 day forecast range. The different pressure distributions result in differences in wind field and wave height field. Fig 12 shows the comparison between coupled wave height forecast, control

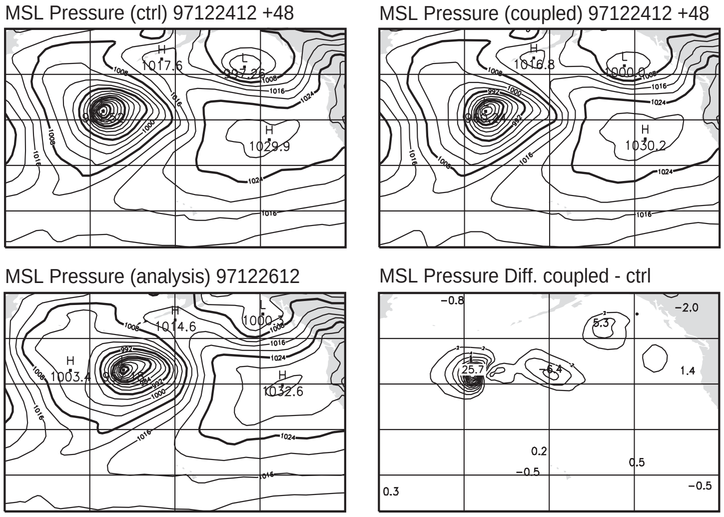

Figure 11: Day 2 forecast of surface pressure for 24 December 1997. Area is North Pacific.

forecast and verifying analysis on midnight of 27 December 1997. Differences between coupled and control wave height reach 4 m and the coupled forecast is in better agreement with the verifying analysis, while the control forecast is too high. This finding agrees with our remarks in Section 2 that the control forecasting system systematically has too high waves in particular in the later stages of the forecast range.

It is concluded that under extreme conditions there is impact of two-way interaction on storm systems, and as a consequence of the modified wind fields this affects the wave height fields as well. Normally, the impact on the weather patterns is of small scale.

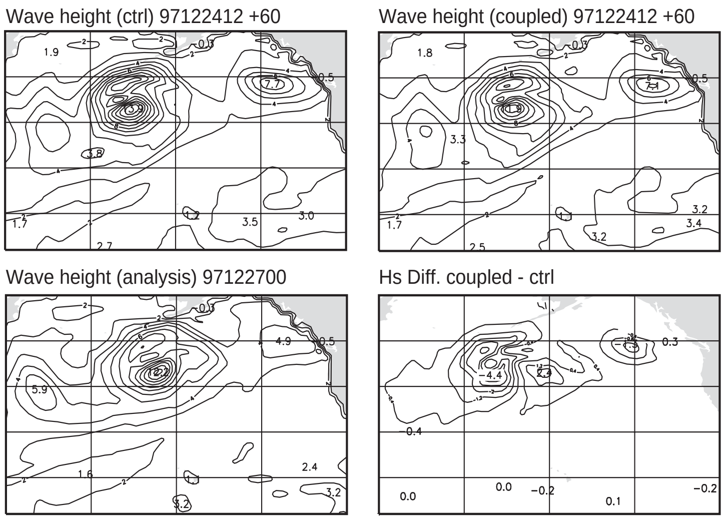

Figure 12: Difference between coupled (top right panel) and control (top right panel) 60 h forecast wave height field and the comparison with the operational wave height analysis (bottom right panel).

The question now is whether there is any systematic impact of the two-way interaction on the forecast performance of wind and waves. We therefore discuss some of the forecast scores of the experiments that have been performed. We remark that the wave scores are quite sensitive to the verifying analysis that is used and therefore we only present wave scores against their own analysis. As a consequence, the number of forecast cases is limited.

As an illustrative example of the impact of two-way interaction on the forecast performance of wave height we show in Fig 13 for the July period in 1996 anomaly correlation, standard deviation of wave height error, systematic error and mean of verifying analysis as function of forecast time. The area is the Southern Hemisphere which from a wave forecasting point of view is the most interesting area at this time of the year. In agreement with our expectations from the extended-range simulations we notice a reduction of the systematic error in the later stages of the forecast while we also observe a reduction in the standard deviation of error and improvements in the anomaly correlation of wave height. This is accompanied by a reduction in standard deviation of surface wind speed error (not shown) and a reduction of the rms error in 1000 mb wind for the full period in question.

Regarding impact of two-way interaction on the atmospheric circulation it should be noted that the early experiments showed a small impact only, in particular in the winter time. Also, the early experiments only showed impact in relatively small areas such as the North Atlantic and the North Pacific. In the summertime impact was somewhat more substantial as follows from Fig 14 which shows the scores of 500 mb geopotential height for Europe and the North Atlantic for the June-July period in 1996. A reason for the relatively large impact of ocean waves on the atmosphere in the summertime may be the following. In winter the atmospheric circulation is dominated by baroclinic activity, and physical processes such as surface friction play a relatively minor role, although, as we have seen, there may be a considerable small scale impact in cases of rapidly developing lows. On the other hand, in the summertime there is less baroclinic activity and therefore physical processes may play a more prominent role in the evolution of weather systems.

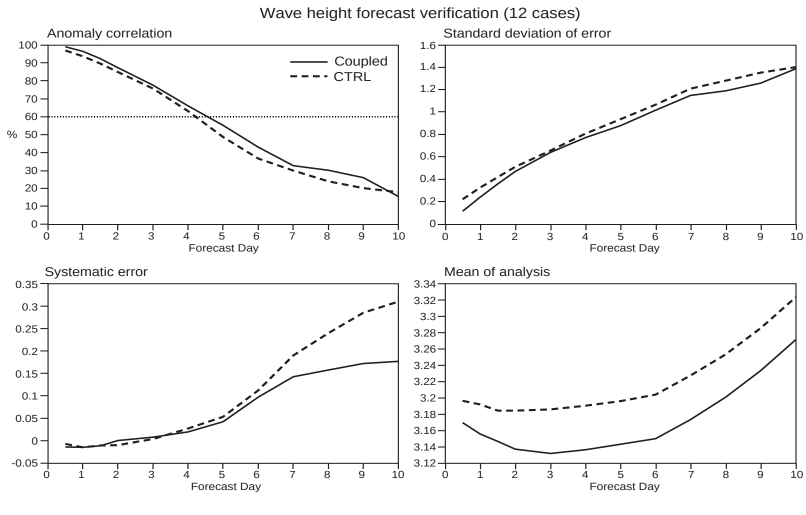

Figure 13: Wave height scores for June-July 1996 period in the Southern Hemisphere. Verification is against own analysis.

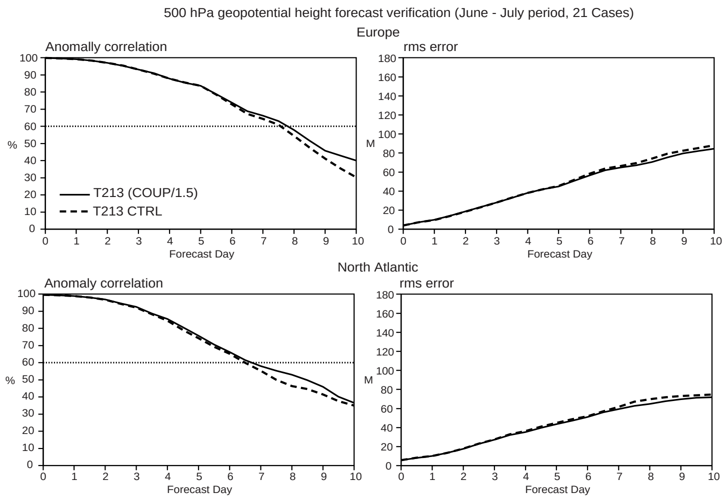

Figure 14: Scores of 500 mb height field for Europe and North Atlantic for the full June-July 1996 period.

In summary, it is concluded that although there may be considerable impact of two-way interaction on synoptic cases there is, because of the relatively small scale of the impact, only a modest positive impact on the scores of the atmospheric parameters. Furthermore, another reason for the relatively modest impact on the atmospheric scores may be that extreme events, where one would expect a substantial impact of two-way interaction, do not occur very frequently. The impact on the wave height scores seems to be more substantial. A reason for this is that the wave forecast depends to a large extent on the quality of the small scale features in the driving wind field.

4.2 Pre-operational and E-suite results

Because results were regarded as positively encouraging it was decided to develop an operational coupled ocean-wave, atmosphere prediction system. Development work took place on a Fujitsu VPP700 which is a parallel machine with vector processors. The parallel environment required that extensive work needed to be done in areas such as an efficient IO in order to minimize the slowing down of the forecasting system. Moreover, the decomposition of the work in the atmospheric part of the model was done in a manner that differed from the ocean-wave model. The reason for this is that in the wave model there are no land points, and it would be a waste of computer resources to execute wave calculations over land. The problem of the different distribution of the work over the processors was solved by exchanging global fields of wind and the Charnock parameter between the atmospheric part and the wave part, and each processor then reads that part of the global field that is needed to perform the necessary calculations.

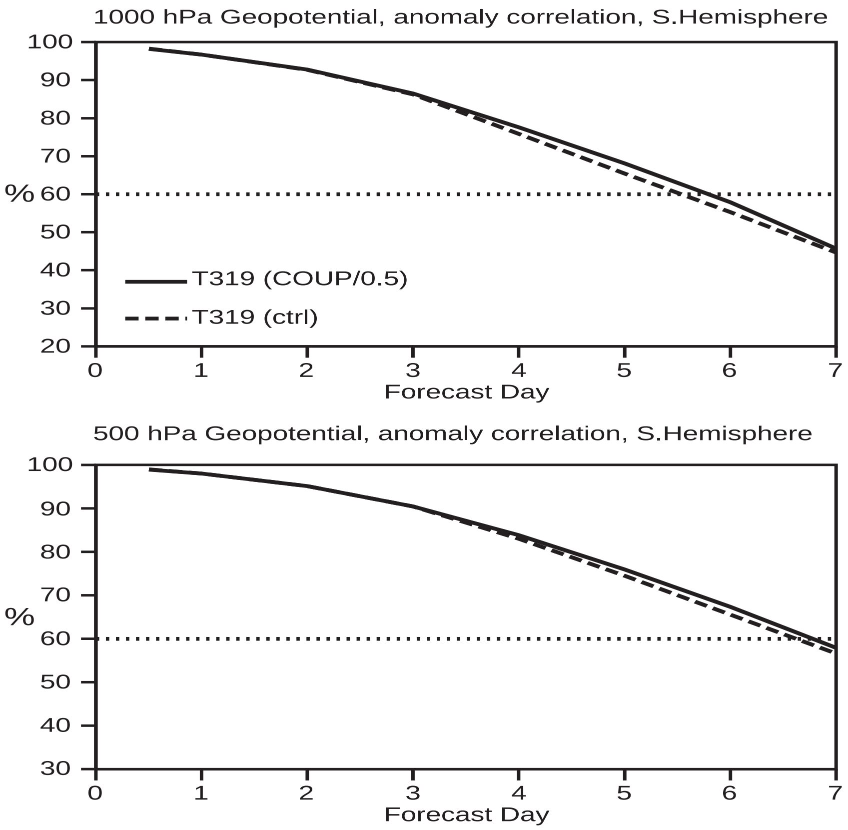

Figure 15: Scores of forecast 1000 and 500 mb geopotential for the Southern Hemisphere for 28 cases in the December 1997/January 1998 period.

During the extensive development work, ECMWF increased the resolution of the atmospheric model from T213 to T_l 319. Here the subscript l the denotes the important change from a reduced Gaussian grid to a linear Gaussian grid (Hortal [52]. Pre-operational testing of the coupled forecasting system was performed with the T_l319/L31-0.5D version on a summer period (17 cases in May 1997) and a winter period (28 cases in December 1997 and January 1998). The general impression was that the impact of the waves on the atmosphere had increased somewhat. This is illustrated for the 28 cases of the winter period in Fig 15 where scores are shown for 1000 and 500 mb geopotential height from the Southern Hemisphere, an area where previously no systematic impact was observed. The positive impact at 500 mb height is in agreement with the simple picture of impact we presented at the beginning of Section 3 namely that changes in surface friction have a barotropic signature so that they affect the whole atmospheric column(see also Janssen and Viterbo [7]). In order to get an idea about the statistical significance of the impact, we show in Fig 16 a scatter diagram of 1000 mb scores at day 5 of the forecast, together with the results from a statistical significance test. A high statistical significance (>99%) is obtained. A reason for the higher sensitivity of the T_l319 atmospheric model to the sea-state dependent roughness could be that this higher resolution model had clearly less horizontal diffusion at the sub-synoptic scales of 100-200 km, which are the scales that are relevant for wind-wave interaction.

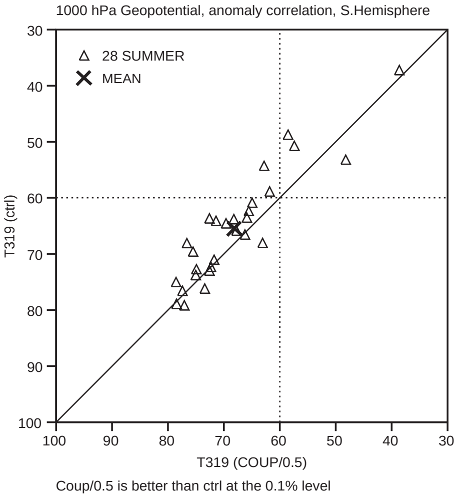

Figure 16: Scatter diagram of day 5 scores of forecast 1000 mb height for the same cases as shown in Fig 15. Results of statistical significance tests are given as well.

As a final test before operational introduction of the coupled forecasting system(CY18R6), a parallel run to the operational suite (e-suite for short) was performed which consisted of a large sample of 74 forecasts. The period extended from April 16 until June 28 1998. In addition to the two-way interaction of wind and waves CY18R6 also contained other changes to the atmospheric model such as the use of SSM/I humidity data, a different usage of radio sonde data in the analysis and a modified variational quality control. Therefore, a comparison of the CY18R6 e-suite with the operational results is not only testing the benefits of two-way interaction. But surface parameters such as the 10m wind speed over the oceans and significant wave height are likely affected most by the wind-wave coupling so we will only show scores from the e-suite for these parameters.

The systematic and standard deviation of error of forecast wave height (as obtained by comparing with the verifying analysis on a 1.5 deg grid) for the Tropics and the Southern Hemisphere are shown in Fig 17a and 17b. First, a considerable reduction in the standard deviation of error is found of about 5%. In addition, while the control (from operations) has a substantial growth of the systematic error through the forecast range, the e-suite shows a much reduced systematic error growth. Therefore, forecast and analysis are in this respect more balanced. Having a sea-state dependent drag coefficient removes a long-standing problem of systematic forecast error growth in the ECMWF wave forecasting system (Janssen et al [27]). In 1994, systematic wave height errors in the 5-10 day forecasts of the Tropics and the Southern Hemisphere were about 20% of the area averaged wave height. However, changes in the ECMWF atmospheric general circulation model in April 1995 (and continuing), and the change from the use of ERS-1 to ERS-2 Altimeter wave height data in the wave assimilation in May 1996 reduced systematic errors to 5-10% of the average wave height. With the introduction of an operational coupled ocean-wave, atmosphere model at ECMWF on 29 June 1998, the systematic forecast error of wave height is 2-3% and has virtually disappeared (see for a more complete discussion Janssen et al [2].

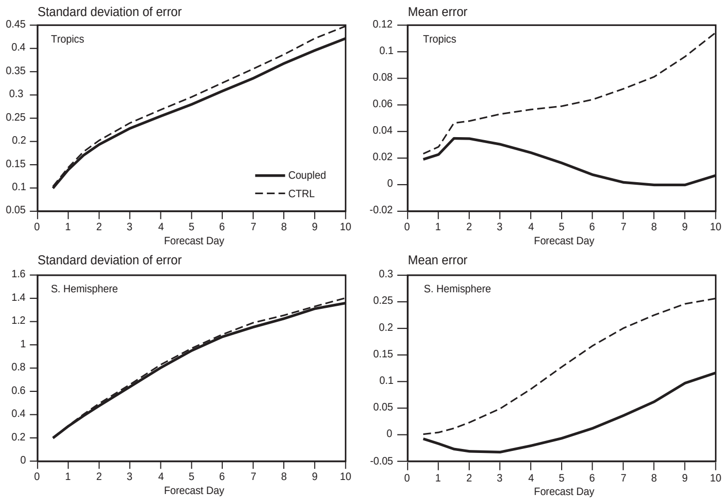

Figure 17: Systematic error and standard deviation of error of forecast wave height as function of forecast time for a) Tropics and b) the Southern Hemisphere. Period is 16 April until 28 June 1998 (74 cases).

. The reduction of systematic forecast error of wave height in the Tropics is an interesting problem because of the combination of local wind-generated waves, called windsea, and remotely-forced wind-generated waves that have propagated long distances from the extra-tropics and are known as swell. In the Tropics swell is the main component of the sea state, and the reduction in systematic error growth in that area is very likely caused by a better simulation of windsea in the extra-tropics. Because the e-suite was run during part of the Austral winter, the storms in the Southern Hemisphere give the dominant contribution to the swells in the Tropics. Hence, most likely the improvement in forecast performance in the Southern Hemisphere, as seen in Fig 17, is the major reason for the reduction of systematic error growth of wave height in the Tropics.

When compared to the verifying analysis, forecast surface wind speed has shown similar improvements in all areas. Also, when compared to observations the quality of the ECMWF wind field improved as well with the implementation of two-way wind-wave interaction in the ECMWF forecast-analysis system on June 29 1998. The root mean square (rms) difference computed between the ECMWF 6-hour and ERS-2 scatterometer winds shows that a 20 cm/s (about 10% of total error) reduction occurred on June 1998 (Fig 18). The bias is reduced by the same amount, although not as clearly visible in Fig 18.

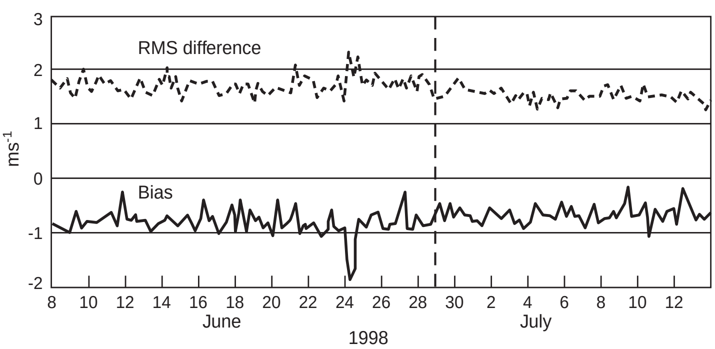

Figure 18: Bias (ERS-2 minus ECMWF) and rms difference between the background

The results of this subsection are summarised as follows. From the pre-operational testing with the T_l319 version of the coupled system it is clear that there is somewhat larger impact of the sea-state dependent roughness on the atmospheric circulation, as illustrated by the improvements in forecast skill over the Southern Hemisphere during the Austral summer. The wave forecast performance has improved as well, in particular there is less systematic error growth during the forecast, in agreement with the results in Section 3.2 on simulations of the wave climate. Finally, the quality of the surface winds has improved subtantially because of the introduction of two-way interaction.

5. Further developments

After the operational implementation of the coupled ocean-wave, atmosphere forecasting system on 29 June 1998, a considerable effort was spend in validation of the new forecasting system against buoy observations, Altimeter wind and wave height data and the verifying analysis. This work confirmed the findings of the previous Section that there was indeed a reduced systematic error growth in wave height with forecast time and a reduction in the standard deviation of error in wave height. In addition, errors in surface wind speed were reduced as well. A detailed account of this verification effort is reported in Janssen et al [2].

Furthermore, we continued the investigation on the benefits of sea state information for atmospheric modelling. An obvious candidate is information on the mean square slope (mss) of the sea surface. This parameter is of importance for a correct evaluation of the ocean surface albedo for low zenith angles, and it may be of help in an improved usage of satellite data such as from ATOVS, SSM/I, Scatterometer and Altimeter. In order to see whether modelled mss has any value, a comparison was made with the well-known observations of Cox and Munk [53]. These observations show a linear dependence of the root mean square slope(rmss) on the surface wind speed. The model results were obtained by averaging the rmss from a global field over a wind speed bin of 10 cm/s, and it was found that in the wind speed range of 5 to 15 m/s the WAM model overestimates rmss compared to the observations, while for larger wind speeds there was a good agreement. The overestimation of modelled rmss was found to be caused by the numerical scheme the WAM model uses to integrate the energy balance equation (Komen et al [16]). In the standard WAM model, the energy balance eqn (2) is solved over a finite frequency range where the high frequency cut-off value f_c is given by

$$f_c = \mathrm{Max}(2.5 * f_{mean}, 4 * f_{pm}). \quad (11)$$

Here, f_{mean} is the mean frequency of the spectrum (approximately given by $1.1 f_p$ with f_p the peak frequency), while f_{pm} is the Pierson-Moskovitch frequency which is inversely proportional to the local wind speed.

There are several reasons for the introduction of a cut-off frequency. An important reason is that in the derivation of the nonlinear transfer the surface elevation is expanded around a zero-mean level. Although this is a valid assumption for long waves, this linearisation is not allowed for the short, high-frequency waves that travel on top of the long gravity waves, because most of the time the ocean surface does not correspond with a zero mean level. As a consequence, theory does not properly describe nonlinear transfer rates for high-frequency waves and in practice an upper limit on the integration of the energy balance equation has been imposed, given by $2.5 * f_{mean}$. The reason for the alternative upper bound to the frequency range, $4 * f_{pm}$, was given by the consideration that in cases of swell with relatively light winds, one would like to describe dynamically the wind sea component as well. However, the nonlinear transfer is then applied to frequencies which are outside its range of validity with the consequence that levels in the high frequency part of the spectrum become too large. In order to check this point we calculated the rmss using an integration scheme with cut-off value

$$f_c = 2.5 * f_{mean} \quad (12)$$

and in the wind speed range between 5 and 15 m/s a close agreement with the observations of Cox and Munk [53] was obtained.

The reduction of mean square slope in the intermediate wind speed range of 5-15 m/s also has consequences for the determination of the Charnock parameter, since less steep waves will give rise to a smaller value. This is shown in Fig 19 where we have plotted the average of the Charnock parameter for the old and the new choice of cut-off frequency. In the intermediate wind speed range the Charnock parameter decreases by as much as 30% with the new choice, while for large winds there is hardly any difference. This reduction in Charnock parameter gives better agreement with stress measurements during TOGA(cf. Zeng et al [54] who show that for low winds a mean value of about 0.013 is found) while the increase of the Charnock parameter with wind speed is in qualitative agreement with stress measurements by means of the eddy correlation technique by Hare et al [39] during FASTEX. As a reference, we have also shown in Fig 19 the mean Charnock parameter from the HEXOS parametrisation (Smith et al [11]) for mature seas which shows a close agreement with the Charnock parameter using the frequency cut-off eqn (12).

It was therefore thought worthwhile to validate the new choice for cut-off frequency in terms of wave and atmospheric scores. We anticipated that the change of the wave model's integration scheme gives rise to a reduced drag at the surface and therefore the systematic wave height error in the forecast increases, when forecast is verified against the analysis. However, the wave analysis uses ERS-2 altimeter wave height data which when compared to buoy data are too low by 7%. There is therefore a need to correct the Altimeter data

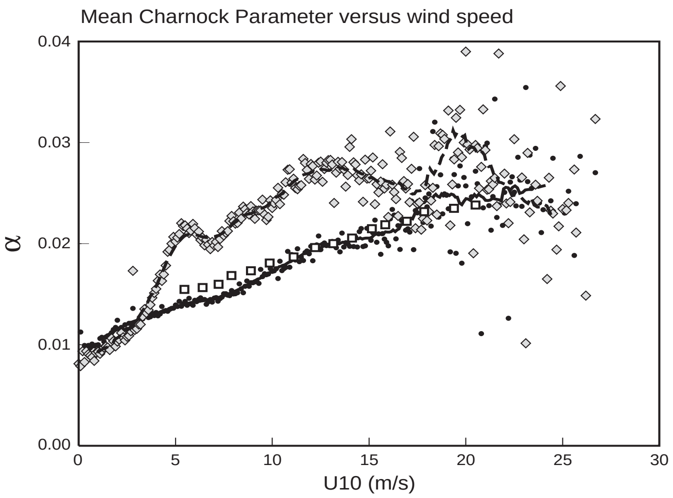

Figure 19: Average Charnock parameter as function of wind speed. Area is the Globe. Results from the old integration scheme are denoted by diamonds, the new integration scheme corresponds to the dots, and the average Hexos results are denoted by the full square.

and this was done in the manner described in Janssen [55]. This correction gives higher analysed wave height and may therefore compensate for the increase of systematic forecast wave height error.

We studied the impact of the new WAM integration scheme on 2 two-week periods in August 1998 and January 1999. We verified the analysed winds and first-guess wave heights against Altimeter data. Although the quality of the surface winds did not change, the first-guess wave height improved, in particular over the Southern Oceans, which gave a reduction in standard deviation of error of about 6%. The overall impact on the scores of geopotential height was neutral in the Northern and Southern Hemisphere with fairly large improvements on small areas such as Europe, the North Pacific and North America. Surface winds showed an improvement and as a consequence forecast skill of wave height increased over areas such as the North Pacific.

Finally, over the last two years there has been a major effort to increase resolution of the atmospheric system both in the vertical and in the horizontal. Presently, there are 60 layers in the vertical while horizontal resolution was increased in November 2000 from T_l319 to T_l511, corresponding to a spatial resolution of about 40 km. The change in horizontal resolution was accompanied by a considerable decrease of the effects of horizontal diffusion, hence giving much more realistically looking kinetic energy spectra. Variability at scales of 100-200 km has increased quite dramatically by a factor of 2-4 resulting in a more realistic simulation of surface wind speed and significant wave height in coastal areas such as Adriatic Sea, North Sea, the Baltic Sea and the east coasts of the USA, East Asia and Australia. The increase of spatial resolution therefore gave a tremendous benefit for wave forecasting.

The increase in spatial resolution was accompanied by a better representation of the wave spectrum: angular resolution was increased by a factor of two while also 5 frequency bins were added, hence the spectrum is presently represented by 30 frequencies and 24 directions. At the same time the propagation scheme was improved by introducing a first order interpolation scheme to determine the relevant energy fluxes, while in agreement with the changes in the surface drag the growth curves needed in the data assimilation scheme (Lionello et al [25]) were modified as well. These wave model changes were tested in the context of the T_l 511/60L atmospheric model and a considerable improvement of forecast skill in atmospheric and ocean wave parameters was obtained. This is illustrated for a 24 day period in August 2000. Fig 20 shows the anomaly correlation of 500 mb geopotential height for the Northern and Southern Hemisphere and in particular in the Northern Hemisphere a large improvement in forecast skill is obtained. Similar large improvements in forecast skill of the waves were found.

This sensitive dependence of atmospheric performance to relatively minor changes in the ocean wave model may come as a surprise. However, it is in keeping with our findings that wind-wave interaction is sensitive to the resolution of the atmospheric model. Higher resolution models have higher (and more realistic) activity on the sub-synoptic scales of 100-200 km, which are the scales that are important for wind-wave interaction. In addition, physical processes, and therefore the sea-state dependent roughness, play a more important role in the summertime, which is nicely reflected by the scores shown in Fig 20.

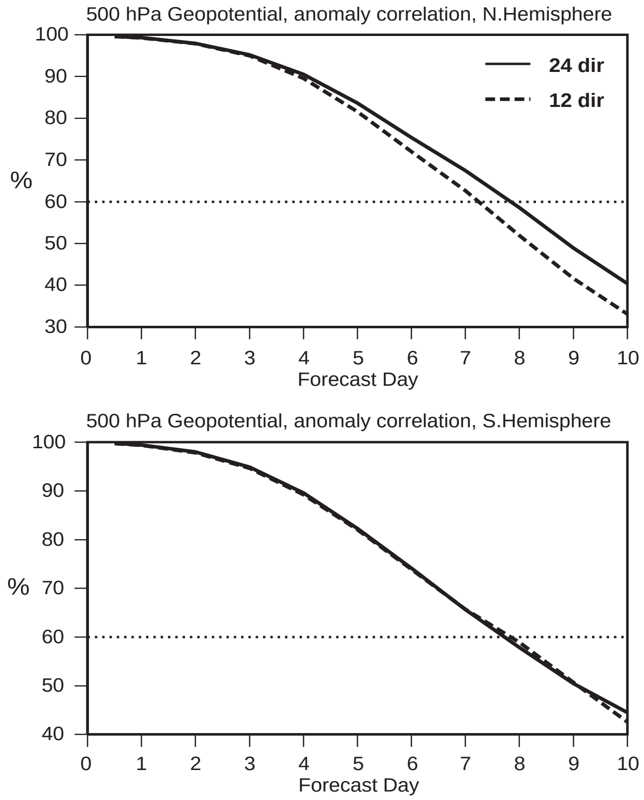

Figure 20: Anomaly correlation of 500 mb geopotential for the Northern and Southern Hemisphere for the last 24 days in August 2000. Here impact of wave model changes on the T1511/60L version of the ECMWF coupled system is shown.

6. Conclusions

In this paper we have studied the impact of two-way interaction between surface winds and ocean waves on the performance of atmospheric and ocean-wave medium-range forecasts. First, we described the evidence for impact of the sea state on the atmospheric circulation around 1995. The work of Doyle [43] on the impact on a single depression and the work of Janssen and Viterbo [7] on the impact on the atmospheric climate suggested some prospects for two-way interaction. However, these studies also revealed that to be successful the atmospheric model needed to have a sufficiently high resolution to give a proper representation of the atmospheric state at the scales that are relevant for wind-wave coupling.

We continued the review by giving a historical account of the development work at ECMWF on two-way interaction. The initial experiments with the T213 atmospheric model (resolution of about 100 km) showed there was significant impact on individual lows. But the impact was small scale and therefore there was only a small systematic improvement on the forecast skill of the atmospheric system, in particular in the summertime. Impact on forecast skill of the waves was somewhat larger. It was evident, that in agreement with results on the wave climate of Janssen and Viterbo [7], a reduced growth in systematic forecast error of wave height would result because of two-way interaction. Subsequent experimentation was mainly concentrated on the impact of increasing resolution of the coupled ocean-wave, atmosphere forecasting system. For example, the T_l319 system showed more sensitivity for the coupling of wind and waves while with the T_l511 system, which was introduced in operations on 21 November 2000, a fairly large sensitivity to changes in the wave model was found. As a consequence, we found a big improvement in forecast skill of atmospheric and wave parameters.

Operational experience with wave modelling at many weather institutes indicates that wave forecasts provide a useful tool for marine forecasters. In addition to these practical benefits, it is known that the wave forecasting at ECMWF has provided useful diagnostic information to improve the performance of the atmospheric forecast model, as discussed by Janssen et al [2]. In this paper we have shown that there are additional benefits for atmospheric and wave forecasting when surface winds and waves are coupled. Additionally, in the context of Ensemble Prediction a coupled wind-wave system has benefits because the ensemble products for waves are then readily available for applications such as ship routing, while at the same time an a priori estimate of wave height forecast error may be obtained. Consequently, the coupled wind-wave system was introduced in operations on 29 June 1998, both in the deterministic suite and in the Ensemble Prediction suite.

Further benefits of sea state information are expected in the future. The available observational data on waves (from Altimeter and SAR) could provide useful information for atmospheric data assimilation, because of the strong interaction of wind and waves. Four-dimensional variational data assimilation in a coupled system would provide consistent wind and wave fields, with the wave information helping to improve the atmospheric wind and temperature fields through the depth of the troposphere. Furthermore, the sea state might be of help in a more appropriate interpretation of satellite data from Altimeter, Scatterometer, SAR, SSM/I and ATOVS.

Also, the model for two-way interaction as developed by Janssen [14] was a first attempt to account for effects of waves on the momentum loss of air at the sea surface. A basic assumption was that the wave stress and the surface wind are in the same direction. Although this is a valid assumption most of the time because most of the wave stress is carried by the short waves (which are in the wind direction), in cases of a mixed sea state, consisting of wind waves and swell, this assumption may be violated. In these circumstances, swells (propagating in a direction that differs from that of the wind) may affect the direction in which the short waves travel, thus giving the possibility that the wave stress is not in the wind direction. Furthermore, the generation of spray may be relevant in the development of hurricanes. A first attempt to assess its importance was reported by Bao et al[24]. No doubt, more sophisticated air-sea interaction models will be developed in the near future, and it will be of interest to study what benefits these models will give for atmospheric and wave forecasting.

Finally, anticipating the development of one model for our geosphere, the ocean waves could provide the necessary interface between the ocean and atmosphere. It is of interest to study in this context the consequences of the modified surface stresses on the tropical ocean circulation and the sea surface temperature distribution (Palmer and Anderson [56]). Work in this direction is presently in progress.

Though Hasselmann [57] could only dream about the advantages of having ocean waves as an interface between the ocean and the atmosphere, it seems that presently the benefits of ocean wave modelling for atmospheric modelling are almost realized, while it will be of interest to see to what extent two-way interaction affects the ocean circulation.

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REFERENCES

[1] Janssen, P A E M, A C M Beljaars, A Simmons and P Viterbo: The determination of the surface stress in an atmospheric model. Mon Wea Rev, 120, 2977-2985, 1992.

[2] Janssen, P A E M, J-R Bidlot and B Hansen: Diagnosis of the ECMWF ocean-wave forecasting system. ECMWF Technical Memorandum no. 318, 2000.

[3] Janssen, P A E M: Wave-induced stress and the drag of airflow over sea waves. J Phys Oceanogr, 19, 745-754, 1989.

[4] Stoffelen, A and D Anderson: The ECMWF contribution to the characterisation, interpretation, calibration and validation of ERS-1 scatterometer backscatter measurements and winds and their use in numerical weather prediction models. ECMWF Technical Report, 1995.

[5] Hasselmann, K and S Hasselmann: On the nonlinear mapping of an ocean wave spectrum into a SAR image spectrum and its inversion. J Geophys Res, C96, 10713-10729, 1991.

[6] Janssen, P A E M: Wave modelling and Altimeter wave height data. ECMWF Technical Memorandum no. 269, 1999.

[7] Janssen, P A E M and P Viterbo: Ocean waves and the atmospheric climate. J Climate, 9, 1269-1287, 1996.

[8] Andersson, E, J Haseler, P Unden, P Courtier, G Kelly, D Vasiljevic, C Brankovic, C Cardinali, C Gaffard, A Hollingsworth, C Jakob, P Janssen, E Klinker, A Lanzinger, M Miller, F Rabier, A Simmons, B Strauss, J-N Thepaut and P Viterbo: The ECMWF implementation of the three-dimensional variational assimilation (3D-VAR). III: Experimental results. Q J Roy Meteor Soc, 124, 1831-1860, 1998.

[9] Rabier, F, H Jarvinen, E Klinker, J-F Mahfouf and A Simmons: The ECMWF operational implementation of four dimensional variational assimilation. Part I: Experimental results with simplified Physics. Q J Roy Meteor Soc, 126, 1143-1170, 2000.

[10] Donelan, M: The dependence of the aerodynamic drag coefficient on wave parameters. Proc of the First Int Conf on Meteorology and Air-Sea Interaction of the Coastal Zone, The Hague, The Netherlands, American Meteorological Society, 381-387, 1982.

[11] Smith, S D, R J Anderson, W A Oost, C Kraan, N Maat, J De Cosmo, K B Katsaros, K Davidson, K Bumke, L Hasse, H M Chadwick: Sea surface wind stress and drag coefficients: The HEXOS RESULTS. Bound-Layer Meteorol, 60, 109-142, 1992.

[12] Fabrikant, A L: Quasilinear theory of wind-wave generation. Izv Atmos Ocean Phys, 12, 524-526, 1976.

[13] Janssen, P A E M: Quasi-linear approximation for the spectrum of wind-generated water waves. J Fluid Mech, 117, 493-506, 1982.

[14] Janssen, P A E M: Quasi-linear theory of wind wave generation applied to wave forecasting. J Phys Oceanogr, 21, 1631-1642, 1991.

[15] Chalikov, D V and V K Makin: Models of the wave boundary layer. Bound-Layer Meteor, 56, 83-99, 1991.

[16] Komen, G J, L Cavaleri, M Donelan, K Hasselmann, S Hasselmann, P A E M Janssen: Dynamics and modelling of ocean waves. Cambridge University Press, 1994.

[17] Gelci, R, H Cazalé and J Vascal: Prévision de la Houle. La méthode des densités spectro angulaires. Bull Inform Comité Central Oceanogr d'Etude Côtes, 9, 416-433, 1957.

[18] Hasselmann, K: Grundgleichungen der Seegangsvoraussage. Schiffstechnik, 7, 191-195, 1960.

[19] Mitsuyasu, H: On the growth of the spectrum of wind-generated waves. 1. Rep Res Inst Appl Mech, Kyushu Univ, 16, 459-465, 1968.

[20] Mitsuyasu, H: On the growth of the spectrum of wind-generated waves. 2. Rep Res Inst Appl Mech, Kyushu Univ, 17, 235-243, 1969.

[21] Hasselmann, K et 15 al: Measurements of wind-wave growth and swell decay during the JOint North Sea WAve Project (JONSWAP). Dtsch Hydrogr Z, Suppl A, 8 (12), 1973.

[22] Hasselmann, S, K Hasselmann, J H Allender and T P Barnett: Computations and parametrizations of the nonlinear energy transfer in a gravity wave spectrum. Part II. Parametrizations of the nonlinear energy transfer for application in wave models. J Phys Oceanogr, 11, 1373-1391, 1985.

[23] Komen, G J, K Hasselmann and S Hasselmann: On the existence of a fully developed windsea spectrum. J Phys Oceanogr, 14, 1271-1285, 1984.

[24] Bao J-W, J M Wilczak, J-K Choi and L H Kantha: Numerical Simulations of Air-Sea Interaction under High Wind Conditions Using a Coupled Model: A Study of Hurricane Development. Mon Wea Rev, 128, 2190-2210, 2000.

[25] Lionello, P, H Günther and P A E M Janssen: Assimilation of altimeter wave data in a global ocean wave model. J Geophys Res, C97, 14453-14474, 1992.

[26] Hersbach, H and P A E M Janssen: Improvement of the short-fetch behavior in the Wave Ocean Model(WAM). J Atmos Oceanic Technol, 16, 884-892, 1999.

[27] Janssen, P A E M, B Hansen and J Bidlot: Verification of the ECMWF Wave Forecasting System against Buoy and Altimeter Data. Weather and Forecasting, 12, 763-784, 1997.

[28] Charnock, H: Wind stress on a water surface. Quart J Roy Meteor Soc, 81, 639-640, 1955.

[29] Phillips, O M: The equilibrium range in the spectrum of wind-generated ocean waves. J Fluid Mech, 4, 426-434, 1958.

[30] Banner, M L: Equilibrium spectra of wind waves. J Phys Oceanogr, 20, 966-984, 1990.

[31] Janssen, P A E M: Experimental evidence of the effect of surface waves on the airflow. J Phys Oceanogr, 22, 1600-1604, 1992.

[32] Donelan, M A, F W Dobson, S D Smith and R J Anderson: On the dependence of sea surface roughness on wave development. J Phys Oceanogr, 23, 2143-2149, 1993.

[33] Monbaliu, J: On the use of the Donelan wave spectral parameter as a measure for the roughness of wind waves. Bound-Layer Meteorol, 67, 277-291, 1994.

[34] Anctil, F and M A Donelan: Air-water momentum flux observations over shoaling waves, J Phys Oceanogr, 26, 1344-1353, 1996.

[35] Janssen, J A M: Does wind stress depend on the sea-state or not? A statistical error analysis of HEXMAX data, Bound-Layer Meteorol., 83, 479-503, 1997.

[36] Yelland, M J, B I Moat, P K Taylor, R W Pascal, J Hutchings and V C Cornell: Wind stress measurements from the open ocean corrected for air flow distortion by the ship. J Phys Oceanogr, 28, 1511-1526, 1998.

[37] Banner, M, W Chen, E Walsh, J Jensen, S Lee and C Fandry: The Southern Ocean Wave Experiment. Part I: Overview and main results. J Phys Oceanogr, 29, 2130-2144, 1999.

[38] Eymard, L, et 12 al: Surface fluxes in the North Atlantic current during CATCH/FASTEX, Quart J Roy Meteor Soc, 125, 3562-3599, 1999.

[39] Hare, J, P Persson, C Fairall and J Edson: Behavior of Charnock's relation for high wind conditions, Preprint volume of the 13th Conference on Boundary Layers and Turbulence, January 10-15, 1999, Dallas, Texas, pp. 252-255, 1999.

[40] Taylor, P and M Yelland: Comments on: On the effect of ocean waves on the kinetic energy balance and consequences for the inertial dissipation technique. Accepted for publication in J Phys Oceanogr, 2001.

[41] Janssen P A E M: Ocean waves do affect the kinetic energy balance. Accepted for publication in J Phys Oceanogr, 2001.

[42] Pedlosky, J: Geophysical Fluid Dynamics, Springer-Verlag, New York. 1987.

[43] Doyle, J D: Coupled ocean wave/atmosphere mesoscale model simulations of cyclogenesis. Tellus, 47A, 766-778, 1995.

[44] Lionello, P, P Malguzzi and A Buzzi: Coupling between the Atmospheric Circulation and the Ocean Wave Field: An Idealized Case. J Phys Oceanogr, 28, 161-177, 1998.

[45] Lalbeharry, R, J Mailhot, S Desjardins, and L Wilson: Examination of the impact of a coupled atmospheric and ocean wave system. Part II.: Ocean wave aspects. J Phys Oceanogr, 30, 402-415, 2000.

[46] DeCosmo, J: Air-sea exchange of momentum, heat and water vapor over whitecap sea states. Ph. D. thesis, University of Washington, 212 pp, 1991.

[47] Weber, S, H von Storch, P Viterbo and L Zambresky: Coupling an ocean wave model to an atmospheric general circulation model. Climate Dynamics, 9, 63-69, 1993.

[48] Janssen, P A E M: Results with a coupled wind wave model. ECMWF Technical Report No. 71, 1994.

[49] Beljaars, A C M: The parametrization of surface fluxes in large scale models under free convection. Quart. J Roy Meteor Soc, 121, 255-270, 1995.

[50] Ferranti, L, F Molteni, C Brankovic and T N Palmer: Diagnosis of extratropical variability in seasonal integrations of the ECMWF model. J Climate, 7, 849-868, 1994.

[51] Janssen, P A E M, J D Doyle, J Bidlot, B Hansen, L Isaksen and P Viterbo: The impact of ocean waves on the atmosphere. Proceedings of a Seminar on Atmosphere-surface Interaction, September 8-12, 1997, ECMWF, Reading, UK, pp 85-112, 1999.

[52] Hortal, M: The development and testing of a new two-time-level semi-Lagrangian scheme (SETTLS) in the ECMWF forecast model. ECMWF Technical Memorandum no. 292, 1999.

[53] Cox, C and W Munk: Measurements of the roughness of the sea surface from photographs of the sun's glitter. J Opt Soc Am, 44, 838-850, 1954.

[54] Zeng, X, M Zhao and R E Dickinson: Intercomparison of Bulk aerodynamic algorithms for the computation of sea surface fluxes using TOGA CORE and TAO data, J Climate, 11, 2628-2644, 1998.

[55] Janssen, P A E M: ECMWF wave modeling and satellite altimeter wave data, in Satellites, Oceanography and Society, D. Halpern ed., Elsevier, pp. 35-56, 2000.

[56] Palmer, T N and D L T Anderson: The prospects for seasonal forecasting - a review paper, Q J R Meteorol Soc, 120, 755-793, 1994.

[57] Hasselmann, K: Waves, dreams and visions, John Hopkins APL Technical Digest, 11, 366-369, 1990