THE PHYSICS OF OCEAN WAVES

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1 Introduction

Wind-generated surface waves on the ocean are important from a physical and a geophysical point of view. Knowledge of the sea state (climatology and prediction) is also of great practical use. Ocean waves have been the subject of many studies. This has resulted in a steady progress of our understanding (see, for example: Phillips, 1977; WMO, 1988; and Komen et al, 1994) and in the availability of advanced measurement tools and advanced numerical models, such as the WAM model (WAMDI, 1988).

In this lecture some of the salient features of these developments will be presented. We will begin with a discussion of basic phenomenology. Simplified cases will be discussed, to illustrate an important message, namely that wave spectra are usually quite complex in realistic situations.

Next we will address the dynamics of wave evolution, and we will show how our understanding of basic wave dynamics and air/sea interaction can be used to compute these complex spectra. This comprises a discussion of the implementation of the WAM model and its validation. We will also discuss the use of a wave model and wave observations for validation of surface wind estimates over the ocean, as obtained, for example, in the ECMWF reanalysis project (ERA).

In the last part we briefly review what is known about the role of waves in air/sea interaction.

2 Phenomenology

2.1 Observations

Our knowledge of waves is based on many observations. *Visual observations* of wind speed and direction, significant wave height, wave period, and wave direction (wind sea and swell) have always been of importance, and will continue to be so. In addition, there are valuable *instrumental observations* from buoys and platforms. Instruments can be used for a determination of the wave spectrum. Certain areas, such as the North Sea, are covered reasonably well with instruments. Instrumental observations in other areas, notably in the Southern Hemisphere are extremely scarce. Fortunately, a wealth of ocean wave observations has become available from *satellites* such as Geosat, Topex-Poseidon and ERS-1 and ERS-2. In particular, the radar altimeter has proven to be quite reliable for measurement of the significant wave height. Synthetic aperture radars (SAR) are able to provide two-dimensional information about the wave spectrum.

Conventional and satellite observations are complementary. Satellites have a global coverage, which is a great advantage, but their repeat cycle (the time lapse between two observations at the same position) is rather low. Conventional observations, on the other hand, have excellent temporal resolution, but they are point measurements, so their spatial coverage is restricted by the number of available instruments.

Figure 1 shows a scatter plot in which for a large number of cases the wave height is given as a function of the wind speed. (The cases are model simulations rather than observations, but they are sufficiently realistic to be used as an illustration of the point.) It is important to note that there is no clear relationship. The solid line is an attempt to draw a relation. But as you can see, it fails. At low wind speed waves are higher then expected. These are cases in which advection of energy (swell) is important. At high wind speed most wave heights lie below the curve. This is because they are not yet fully grown.

To illustrate this we will discuss wave generation in a number of simple situations.

2.2 Duration- and fetch-limited wave growth

Imagine an infinite flat windless ocean. At a given time, say $t = 0$, a uniform wind starts blowing. Due to instability (Kawai, 1979) waves begin to form. They grow with time. The significant wave height H_S, everywhere equal, will be a simple function of two variables, the wind speed u and time t:

$$H_S = H_S(u, t) \tag{1}$$

Wind-generated surface waves are gravity waves, i.e. waves in which gravity is the restoring force. Therefore, one may expect that their behaviour scales with the strength of the gravitational acceleration g. This allows one to rewrite (1) as

$$\frac{H_S g}{u^2} = f\left(\frac{tg}{u}\right). \tag{2}$$

The function f has been determined from observations. Wave modellers can not agree on the exact shape of f but typically $H_S g/u^2$ grows as a function of tg/u to reach a maximum level of $H_S g/u^2 = 0.22$ at $tg/u = 2 \times 10^5$. For longer durations the growth becomes negligible, so that the waves don't grow higher for that given wind speed. For a typical strong wind of 20 m/s, it takes about two days until the waves reach their maximum height of about 9 m. Younger waves will be less high. This gives a qualitative explanation for the scatter at the high-wind end of figure 1.

As an alternative to duration-limited growth one often discusses fetch limited growth. In this case one considers a half plane ocean, with a constant off-shore wind of infinite duration. The wave height then is a function of wind speed and distance to shore, the so-called fetch. This situation can be treated in a similar way, i.e. the relevant quantities can be made non-dimensional, and one can consider the nondimensional wave height as a function of nondimensional fetch.

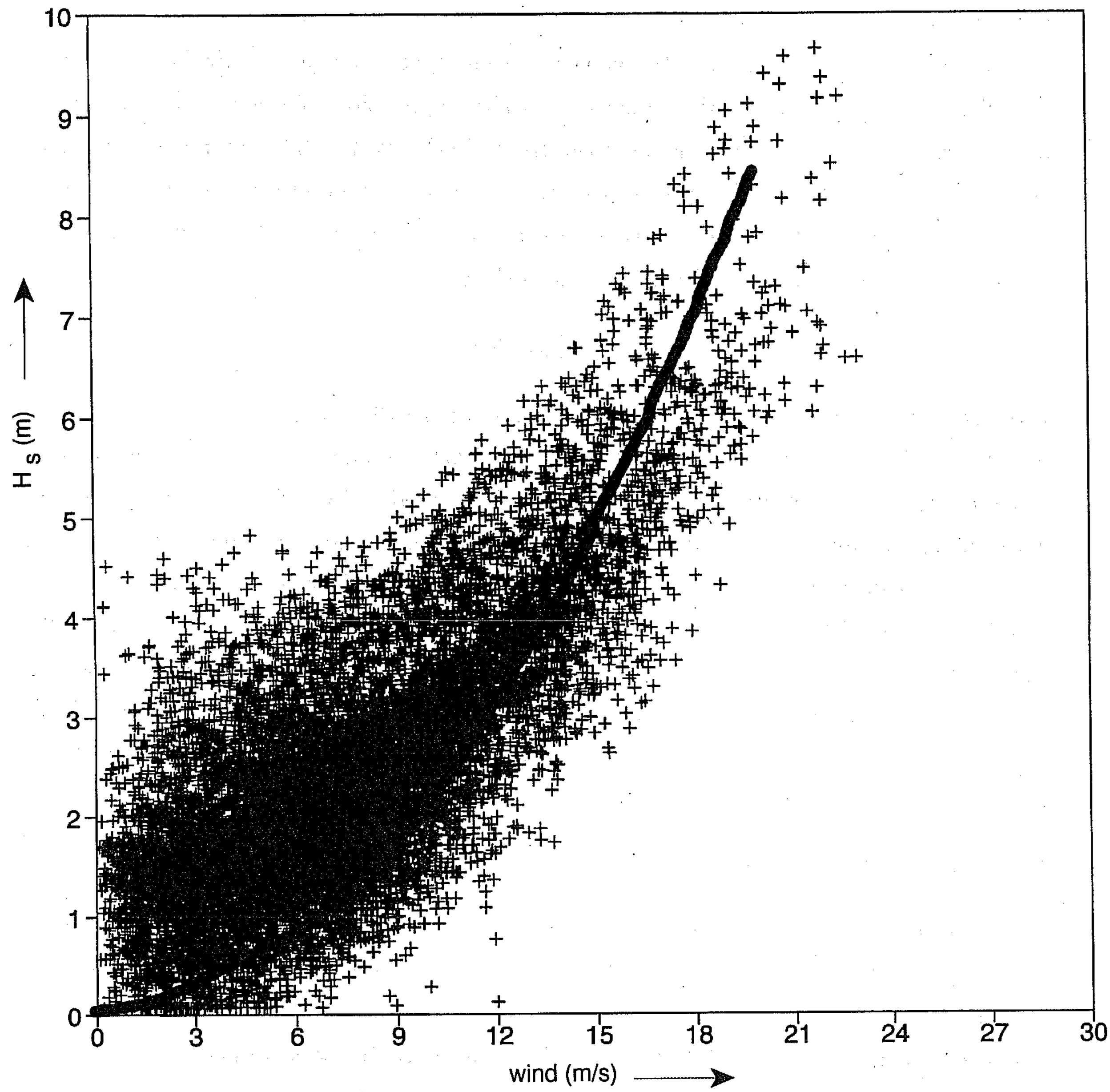

Fig. 1. The significant wave height as a function of wind speed for a large number of (simulated) situations. There is no relation.

This fetch-limited growth curve has the same asymptotic value as the duration-limited growth curve. At a wind speed of 20 m/s one needs a fetch well over 1000 km to reach asymptotics.

The concepts of fetch- and duration-limited growth have only limited value, because in reality they are never encountered: winds are not homogeneous and constant, and coasts are not straight.

Another problem arises from the choice of wind speed parameter. Scaling with u is disputable, because there is another quantity with the dimension of velocity, which is probably more relevant for wave growth, namely the friction velocity u_*. This parameter is defined as the square root of the (kinematic) surface stress τ. It can be determined from high-resolution measurements of the horizontal and vertical wind speed fluctuations. From such measurements it was found that the drag coefficient C_D, defined by

$$u_*^2 = C_D u^2 \tag{3}$$

varies roughly between 1×10^{-3} (low wind speeds) and 3×10^{-3} (high speeds). This implies that u-scaling and u_*-scaling can not both be correct. When u scaling is correct - as assumed above - one obtains an asymptotic wave height which is proportional to u^2. In the case of u_* scaling it will be proportional to u_*^2. Suppose one has two scaling relations, one based on u-scaling, the other on u_*-scaling, and giving the same asymptotic wave height at low wind speed. Then they will give asymptotic wave heights that differ by a factor of three at high wind speed. So, not both can be in agreement with observations. For practical applications the difference is unacceptably large. There are arguments and observations that favour u_*-scaling (Janssen, Komen, de Voogt, 1987), but even until today the superiority of u_*-scaling has not been proven convincingly from wind and wave observations.

2.3 Spectral evolution

Of course, the wave height is not the only quantity that matters. Individual, monochromatic, waves are characterised by their height (which equals 2× the amplitude); period; length and direction. Period and length are related through the dispersion relation (see next section), so one may choose height, period and direction as the basic characteristics. In general, the sea surface consists of a superposition of many waves. To characterise these one has to specify how the energy is distributed over different directions. This is done with the help of the two-dimensional wave spectrum. The total energy is proportional to the square of the significant wave height. This allows one to write:

$$H_S^2 \sim \int F_{(2)}(f,\theta)\mathrm{d}f\mathrm{d}\theta. \tag{4}$$

Here $F_{(2)}$ is the two-dimensional wave spectrum, $f = 1/T$ is the frequency (the inverse period) and θ is the wave direction. There exist different ways of representing the two-dimensional wave spectrum. In one representation the spectral level is contoured in the $f - \theta$ plane. The other representation is a polar one, in which the distance to the origin is proportional to the frequency and the angle with the horizontal axis corresponds with θ.

The one-dimensional spectrum is also often used. This is defined by integration over angles:

$$F_{(1)}(f) = \int F_{(2)}(f, \theta) \mathrm{d}\theta. \tag{5}$$

The shape of the spectrum provides the basic mathematical and physical characterisation of the sea state. One may now wonder what determines this spectral shape. A first answer to this question came from the Jonswap experiment (Hasselmann et al, 1973). In this experiment the one-dimensional wave spectrum was measured under conditions that were close to the ideal situation of fetch-limited growth. Three important features were observed: 1. all spectra were sharply peaked; 2. the peak-frequency decreased with increasing fetch (and the corresponding wave length increased); 3. the high frequency tail approached an asymptotic universal level, but an overshoot occurs before this level is reached (figure 2). Other measurements provided evidence that the spectrum reaches a stationary shape for very large fetches. It was also found that waves travel in the wind direction in the mean, but that there is some angular spread about this direction. Duration-limited growth is thought to have a similar spectral characteristic. This relatively simple picture has stimulated the development of so-called parametric wave models in which the spectral shape was parametrised in terms of a small number of parameters.

As already noted, the concepts of fetch- and duration limited growth have only limited value, because in reality they are never encountered. When the conditions are not ideal, spectra are no longer simple. This became first apparent in the SWAMP (1985) study, when the behaviour of many wave models was compared in situations that were still idealised but which differed from fetch- and duration-limited wave growth. The simplest such case is the slanting fetch situation in which the coastline is still straight and the wind is constant but blowing under an angle. This was found already to introduce a significant asymmetry in the spectrum. Another simple situation is one in which the wind is turning. This gives rise to a complex spectral response. Things become even more complicated when the wind field has a realistic spatial and time-varying structure. In such situations one may distinguish between advected waves and locally growing waves. Advected waves are called swell when their propagation direction differs significantly from the local wind direction or when their propagation speed is significantly larger than the local wind speed. In extreme situations one may locally have large waves without (or even against) the wind. Such swells can easily be observed as breakers near a beach. They account for the high wave heights at low wind speeds in figure 1.

In general, wave spectra can have very complex shapes indeed. This is nicely illustrated in figure 3 (Gerling, 1991), which shows a time sequence of spectra for realistic conditions. The central question then is whether we can understand and predict this type of spectral evolution on the basis of the fundamental laws governing ocean wave dynamics. The answer is yes, as we will see in the next section.

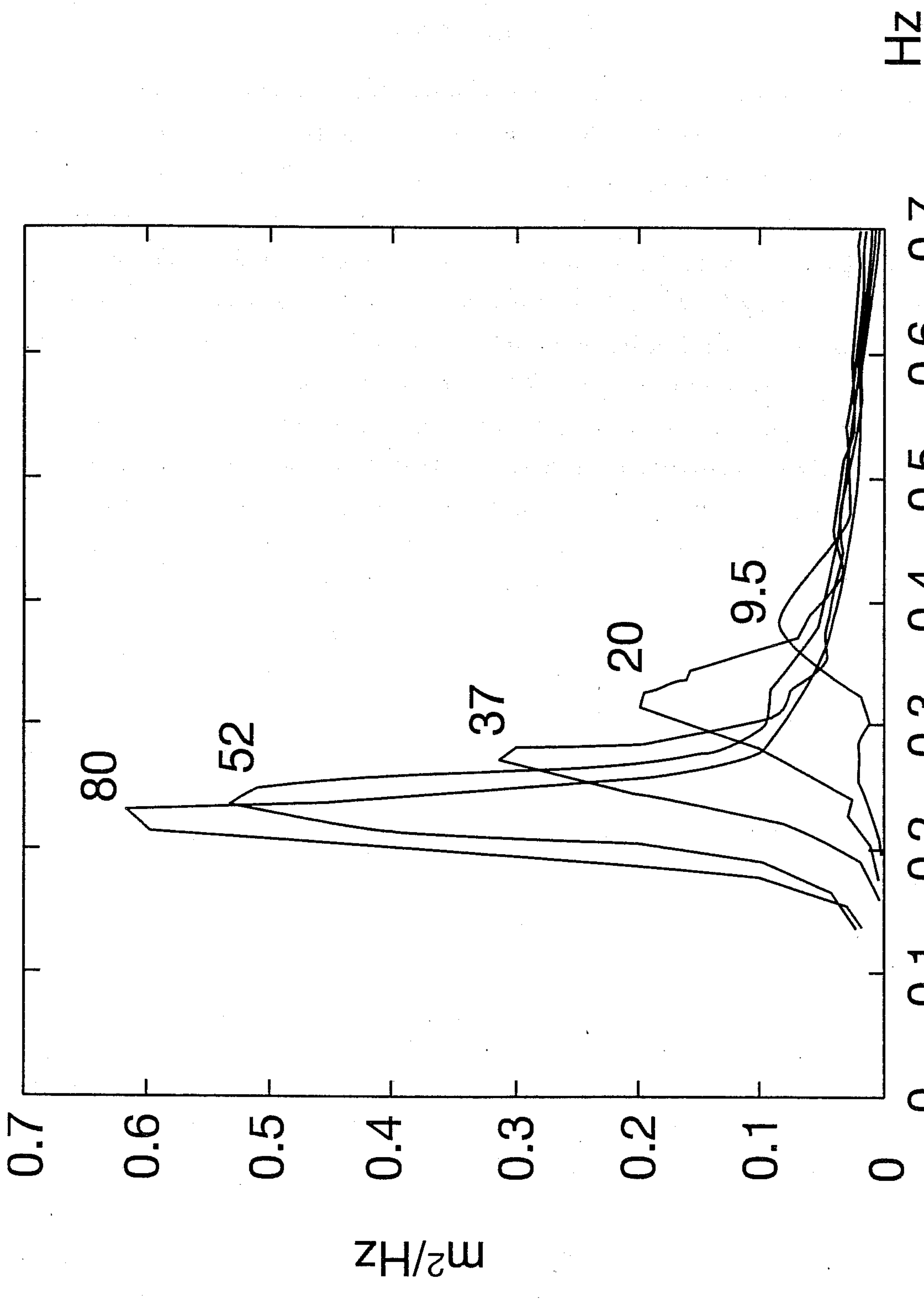

Fig. 2. Spectral evolution as observed during Jonswap (Hasselmann et al, 1973(, under conditions which were close to the idealisation of fetch-limited growth.

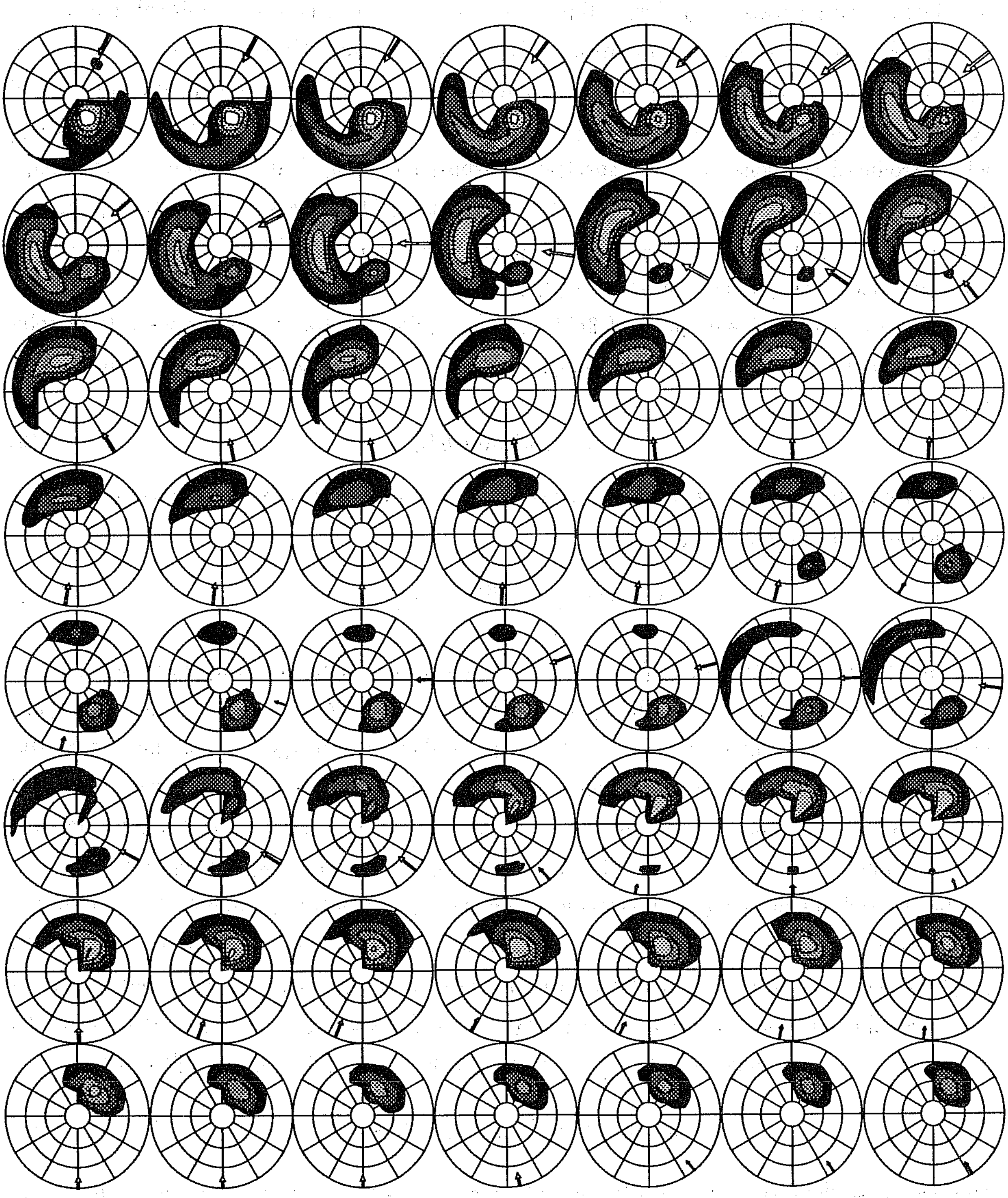

Fig. 3. Time sequence of two-dimensional spectra as simulated by Tom Gerling (1991). The arrows indicate the wind direction. This sequence illustrates the complexity of the spectral shapes under realistic conditions.

3 The WAM model

3.1 The theoretical basis

The first operational wave predictions were based on the work of Sverdrup and Munk (1947), who introduced a parametrical description of the sea state and who used empirical relations to describe wind sea and swell. An important advance was the introduction of the wave spectrum by Pierson et al (1955). This was not yet accompanied, however, by a corresponding dynamical equation describing the evolution of the spectrum. This step was made by Gelci et al (1956) who introduced the concept of the spectral transport equation. This equation is used today as the basis for wave prediction models. The derivation of this spectral transport equation is rather lengthy. Details can be found, for example, in Komen et al (1994). Here we summarise the main steps that are made in its derivation.

First, the linearised equations of fluid dynamics allow for free harmonic wave type solutions. The dynamics determines the the dispersion relation, relating wavenumber k and angular frequency ω:

$$\omega^2 = gk \tanh kD. \tag{6}$$

This relation depends explicitly on the depth D. A general solution takes the form of a superposition of monochromatic waves:

$$\eta(\mathbf{x}, t) = \sum_{\mathbf{k}} a_{\mathbf{k}} e^{i(\mathbf{k}\cdot\mathbf{x}-\omega t)} + \text{c.c.}, \tag{7}$$

Here η is the height of the sea surface, $\mathbf{x}$ is a two-dimensional vector denoting position, t is time, $\mathbf{k}$ is the two-dimensional wavenumber, a denotes the amplitude of the different wave components, and $\omega = 2\pi f$ is the angular frequency. Of course, realistic waves are not free and linear. But one may assume that nonlinearities and interactions with the environment are weak – at least in the mean – and can be accommodated by allowing for slow variations of amplitudes and corresponding wavenumbers. This is known as the geometrical optics approximation.

In a second step one introduces statistics. One does not go for a deterministic description, but merely seeks to make statements about average quantities. So one does not want to predict the height of the sea surface at a particular position and a particular time, but one wants to predict the probability that the waves have a particular height at a particular place and time. Statistics is introduced by assuming that the different components are - and remain - independent:

$$< a_k a_{k'}^* >= |a_k|^2 \delta_{k,k'} = \frac{1}{2} F(\mathbf{k}, \mathbf{x}, t)\delta_{k,k'}. \tag{8}$$

One can now show that the two-dimensional spectrum $F(\mathbf{k})$, (we suppress the subscript 2) satisfies the following equation:

$$\left\{\frac{\partial}{\partial t} + \mathbf{c_g} \cdot \frac{\partial}{\partial \mathbf{x}}\right\} F(\mathbf{k}, \mathbf{x}, t) = \sum_{\ell} S_\ell(F; \mathbf{u}) = S_{in} + S_{nl} + S_{ds} \tag{9}$$

The form given here applies to deep water. A more general version with shallow water effects and wave-current interactions is given in Komen et al (1994). The left-hand side describes propagation (c_g is the group velocity defined by $d\omega/dk$); the source terms on the right describe the interaction of wave components with the atmosphere (S_{in}, the input), among themselves (S_{nl}) and with the ocean (S_{ds}, dissipation). Thanks to major research efforts made over the last 30 years we now have suitable parametrisations for the different source terms. In general, these are based on a mixture of theoretical arguments and empirical observations. All source terms are functionals of the wave spectrum, F. Dissipation and wave-wave interactions are nonlinear in F. The wind input is linear in F, but it depends also on the local wind vector $\mathbf{u}$.

It seems appropriate to repeat the main assumptions that went into the derivation of equation 9 because they presumably specify its validity range. These assumptions were 1. the geometrical optics approximations; 2. weak nonlinearity (in the mean); and 3. (local) statistical homogeneity. These conditions seem to be well met at sea, but they surely break down in situations in which the waves change on a scale which is short with respect to their wavelength or period, such as may happen closely to the shore.

Equation 9 can be solved numerically by forward integration in time. To do this, one needs to specify forcing wind fields at each grid point and for each time, as well as initial and boundary conditions. Wind fields are usually taken from NWP models. Coasts are normally treated as fully absorbing. Open boundaries may be avoided by a suitable choice of the integration area. If this is not possible one may nest a limited area model within a larger scale model.

The initial conditions need some discussion. Physically, the instability of air flow over a flat sea occurs at centimetre wave lengths. At these short length scales surface tension still plays a role. Also the physics of the interaction between the boundary layer and these short waves is rather different from the physics governing the growth of longer waves. Therefore, but also because of computer limitations, the very short wave scales are not resolved in the WAM model. It is then consistent to assume that there is always a minimum spectral energy level, which can be parametrised in terms of the grid size, the time step and the wind speed. When a cold start is made the model is initialised at this level.

3.2 The energy balance

The first applications of the WAM model concerned the idealised cases discussed in the previous section. It is instructive here to consider the case of duration-limited growth. The phenomenology of this type of growth was discussed above. It contained the following elements: a peaked spectrum; a peak that moves to the left when duration increases; overshoot at high frequencies; and an approach to a fully grown stage in which the spectrum does not change anymore. All of these features have been reproduced (Komen, Hasselmann and Hasselmann, 1984). An analysis of these results has greatly improved our understanding of what is going on. An illustration of this is given in figure 4 which shows the different source terms during the stage of active

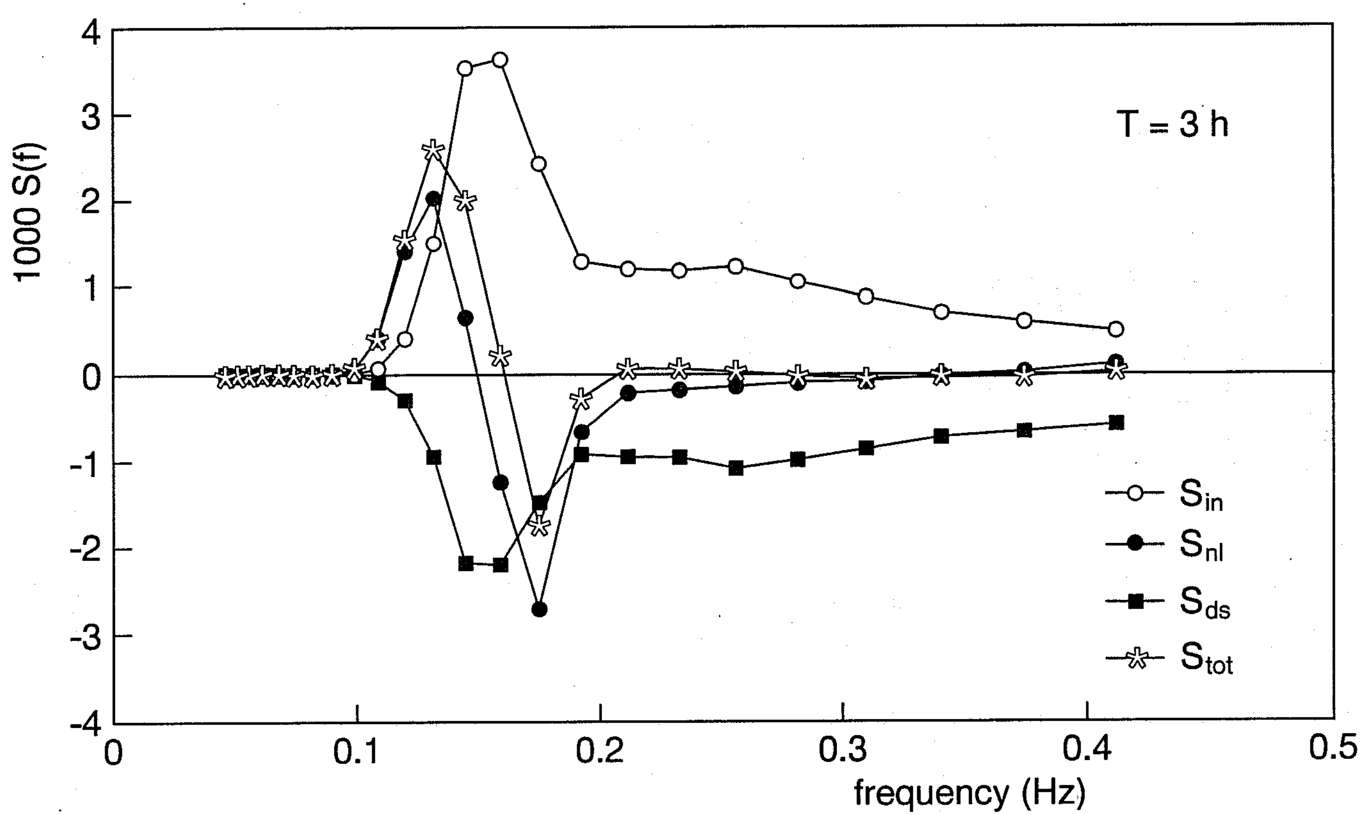

Fig. 4. The energy balance in growing waves (from Komen et al, 1994). The source terms were computed for a spectrum that peaked about 0.17 Hz.

growth. The total source term is characterised by a positive lobe at frequencies below the peak frequency of the spectrum. This is a combined effect of the different source terms. The wind input occurs mainly at relatively high frequencies, but the nonlinear interactions transfer this energy to lower frequencies. A similar analysis can be made of the fully grown stage, when the three source terms, wind input, nonlinear transfer and dissipation balance each other to within several orders of magnitude.

3.3 Model implementation and validation

The WAM model has been implemented operationally in many forecasting centres. ECMWF runs a global version of the model with a half-degree resolution to make 10-day forecasts. KNMI runs a North Sea version of the model. An example of the output of the global model is given in figure 5, which shows isolines of significant wave height, as computed for one particular time. Quasi-synchronous wave height observations by ERS-2 are also plotted. The performance of the model has been tested against buoy and satellite observations (Komen et al, 1994; Janssen, Hanssen and Bidlot, 1996). A typical bias in the operational forecast is of the order of 0.25 m. A measure for the quality of the forecast is the anomaly correlation between the forecast and the verifying analysis. For the northern hemisphere this correlation typically decays from 95 % for day 1, to 80 % for day 3 and 60 % for day 5. In the tropics, where swell is relatively more important, this decay is even slower. Beyond the 5-day horizon the forecasting skill is further reduced.

This clearly illustrates the dependence of the quality of computed wave heights on the quality of the input winds. Figure 6 (based on work by Cardone et al, 1995, and taken from Komen et al, 1996) shows a comparison of observed and modelled wave heights, made during the SWADE experiment. The comparison is made for two buoys located in the western North Atlantic. For this experiment very high-quality winds were available. Shown are wave height predictions obtained with routine ECMWF wind estimates and wave heights obtained with winds from a dedicated regional analysis, labelled OW/AES. In either case the WAM model was used to compute the wave heights. The agreement between modelled and observed wave heights is striking, especially in the very-high-quality wind case.

Sterl et al (1997) concluded from this that wave observations in combination with the WAM model can be used to assess the quality of surface wind estimates, and they applied this to validate the ERA reanalysis. They obtained a good climatological agreement between observed and computed wave climatologies, which confirmed the accuracy of the ERA reanalysis winds. However, they also detected a slight underestimate of high wind speed cases, which they ascribed to the limited spatial resolution of ERA.

Wave *forecasting* is only one application of a wave model. Another important application is the use of a wave model to study the wave climate (Guenther et al, 1997). Interest in wave climate studies has recently increased when it was realised that the wave climate may exhibit a

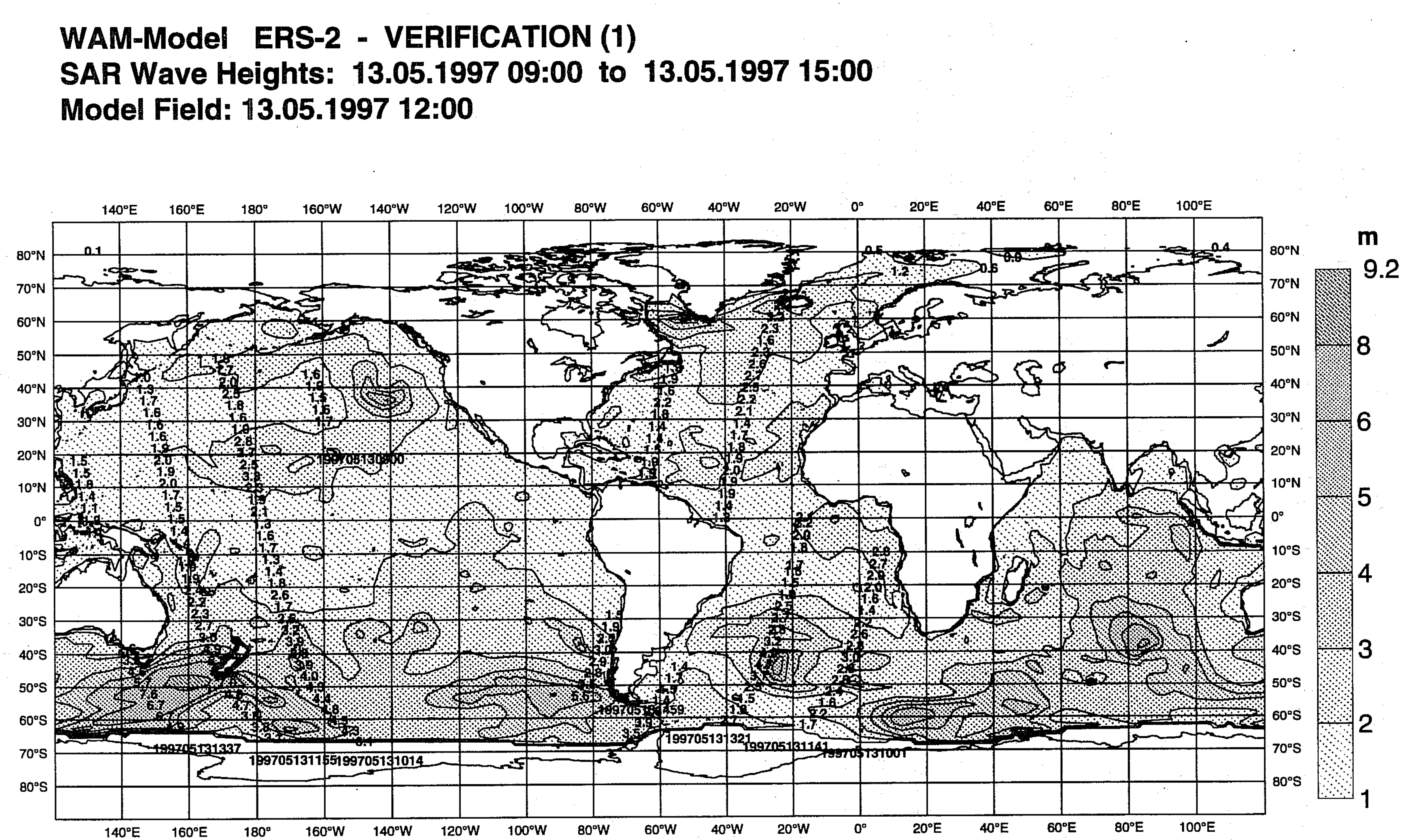

Fig. 5. An example of a global wave analysis for one particular time. Isolines give the significant wave height. Wave height observations by ERS-2 are also plotted.

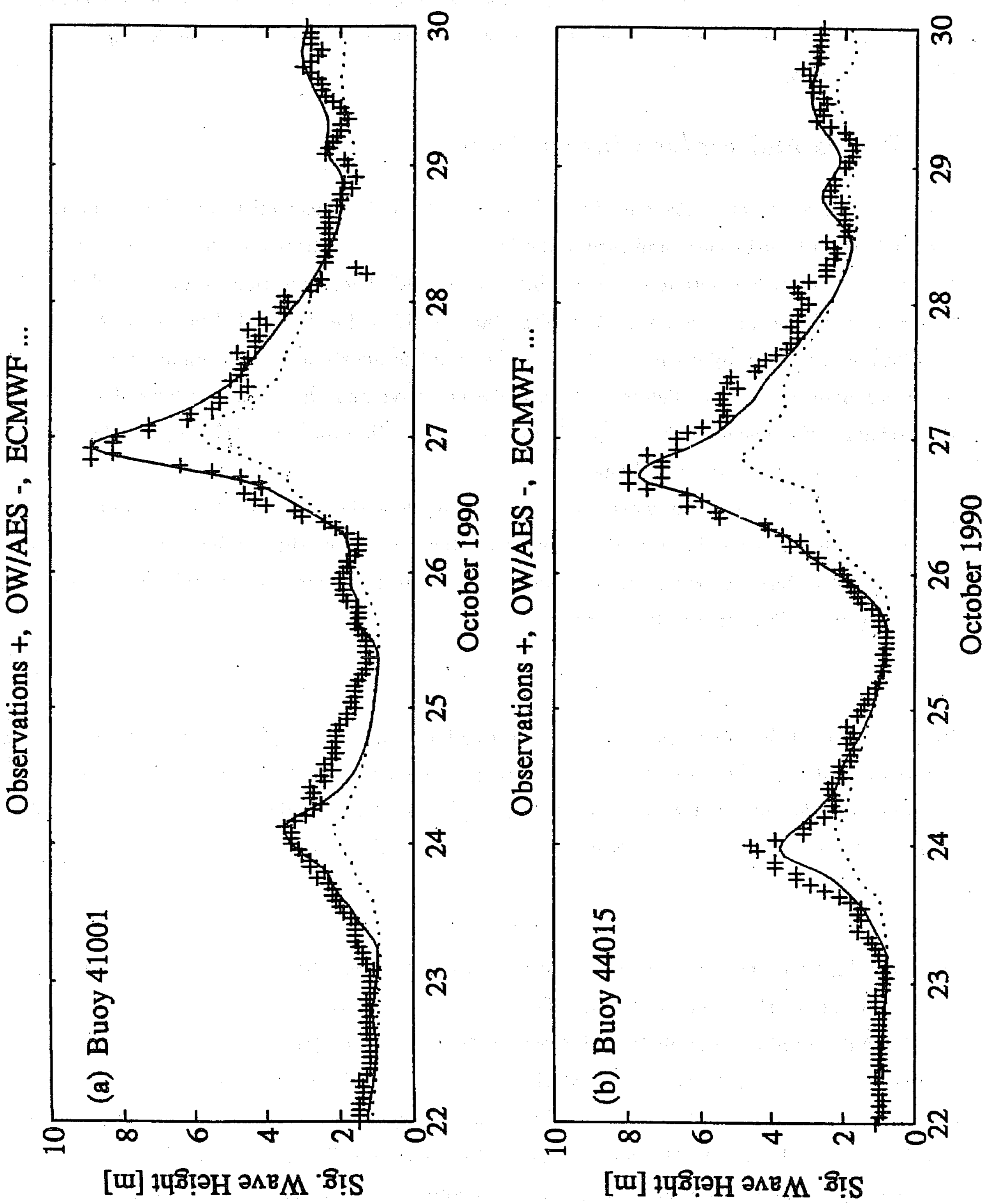

Fig. 6. Comparison between the significant wave height recorded at two positions in the western North Atlantic and the corresponding values obtained by forcing the WAM model with ECMWF and OW/AES winds, respectively.

considerable amount of decadal variability (Carter and Draper, 1988; Bouws et al, 1996; WASA, 1997; Kushnir et al, 1997). The WAM model has also been forced with winds obtained in climate simulations to estimate the effect of CO2 doubling on the wave climate of the North Atlantic (Rider et al, 1996).

4 Waves and air/sea interaction

There remain some very basic questions. What exactly is the mechanism of wave generation? How important is turbulence, and the interaction between wave motion and turbulence in the air? How can one better understand the role of breaking waves? These questions have been addressed by many people (see, for some recent studies: Janssen, 1994; Belcher and Hunt, 1993; Makin et al, 1995), and much progress has been made, but a lot of questions remain unanswered, or the answers remain unverified. The interaction between waves and the boundary layer determines the exchange of momentum, but it also influences evaporation and gas exchange. In this paper we only consider momentum exchange.

There can be no doubt that waves play an essential role in the air/sea exchange of momentum. A question that has not been settled is whether their effect on the surface stress, τ, can be parametrised in terms of wind speed alone. To illustrate this point, we consider the standard neutral logarithmic boundary layer profile:

$$u(z) = \frac{u_*}{\kappa} \ln(\frac{z}{z_0}). \tag{10}$$

Here κ is the von Kármán constant, z_0 is the roughness, and $u_* = \sqrt{\tau}$ was already defined as the friction velocity. From this relation the stress can be determined in terms of the wind speed at observation height provided one knows the roughness. This roughness is determined by the shape of the ocean surface. Charnock (1955) related it to the wind stress:

$$z_0 = z_0^* u_*^2 / g. \tag{11}$$

For a constant Charnock parameter z_0^*, one can use equations (10) and (11) to determine the stress in terms of *the wind speed only*. The result is in fair agreement with observations. In particular it reproduces the observed increase of C_D with wind speed.

However, from a theoretical point of view it is hard to believe that this is the whole story (see also, Komen et al, 1997). This can be illustrated by reconsidering the energy balance in growing waves. This energy balance, as described by equation (9), can also be be interpreted as a momentum balance equation, because the wave momentum can be expressed in terms of the wave spectrum as:

$$M(\mathbf{k}) = \rho \omega \vec{\imath}_{\mathbf{k}} F(\mathbf{k}). \tag{12}$$

Here $M(\mathbf{k})$ is the two-dimensional momentum spectrum, ρ is the density of water, and $\vec{\imath}_{\mathbf{k}}$ is the unit vector in the $\mathbf{k}$ direction. The total momentum in the waves is the integral of M over all

k. The rate of change of M is determined by the sum of three terms: the momentum flux form the atmosphere to the waves, the transfer of momentum between different wave components due to resonant nonlinear interactions, and the loss of momentum due to wave dissipation. Here we are only interested in $\dot{M}|_{in}$, the momentum flux from the atmosphere to the waves. It is straightforward to estimate this flux as

$$\tau_w \simeq \int \dot{M}|_{in} \mathrm{d}\mathbf{k} \simeq \int \omega \vec{\imath}_{\mathbf{k}} S_{in}(F, \mathbf{u}) \mathrm{d}\mathbf{k}. \tag{13}$$

(We left out uninteresting factors ρ). This gives the wave stress in terms of the wind speed u *and* the wave spectrum F.

The total flux from the atmosphere to the ocean is the sum of τ_w and a viscous distribution. However, many studies have suggested that this viscous contribution is relatively small, especially at higher wind speeds, so one has $\tau \simeq \tau_w$.

At first sight, it seems unlikely that (13) can be parametrised in terms of the local, instantaneous wind speed only, because we have seen that the wave spectrum, in general, is a very complex function of the wind field history. However, $S_{in} \simeq \beta\omega F$. The function β depends on the wind speed and increases with ω. This and the extra factor ω in the source term combined with the explicit factor ω in (13) gives extra weight to the short waves at high frequency:

$$\tau \simeq \int \omega^2 \vec{\imath}_{\mathbf{k}} \beta F(\mathbf{k}) \mathrm{d}\mathbf{k}, \tag{14}$$

These short waves have a shorter time scale than the energy containing long waves, so they may be more or less adjusted to the instantaneous local wind. This enhances the importance of the local wind for roughness parametrisations and explains the success of Charnock's relation. Nevertheless, recent research has attempted to demonstrate the existence of an additional dependence on the long wave spectrum.

For fetch- or duration-limited wind sea the spectrum can be determined in terms of the non-dimensional fetch or duration. A parameter that is suitable in either case is the non-dimensional wave age or stage-of-development parameter ξ. This parameter is defined as $\xi = c_p/u_*$, with $c_p = \omega_p/k_p$ i.e. the phase velocity corresponding to the peak frequency of the wave spectrum divided by the friction velocity. (Remember that waves become longer with increasing fetch or time; long and old waves propagate more rapidly, so ξ does increase with fetch and duration.) Early observational and theoretical efforts have attempted to describe the wave dependence of the drag in terms of the wave age (Donelan, 1982; Janssen, 1992, Donelan et al, 1993). Experimental evidence is still quite confusing. In the Humidity Exchange over Sea Experiment (HEXOS, Smith et al, 1992) the following result was obtained for the Charnock parameter z_0^*:

$$z_0^* = 0.48/\xi. \tag{15}$$

In this relation the dimensionless roughness explicitly depends on sea state. Equation (15) is based on observations made from a tower near the Dutch coast, using eddy correlation techniques to determine u_*, and on earlier observations in the open ocean (see, Komen et al, 1997 for a more elaborate discussion of this point). Recently, Yelland and Taylor (1996) carefully performed a large number of observations using the - indirect - inertial dissipation method. They did not find a wave dependence.

So, in summary, one may safely say that experimental evidence is conflicting. Answers to the above questions will not only help improve the quality of wave prediction, they will also provide better forcing for ocean circulation models of the ocean (Mastenbroek et al, 1993; Burgers et al, 1996), better boundary conditions for atmospheric GCM's (Doyle 1994, Janssen and Viterbo, 1995), and improved coupling for coupled atmosphere/ocean models.

5 Outlook

To make useful wave fore- or hindcasts one needs good quality input winds. Fortunately, atmospheric models have considerable skill, which allows us to make useful wave forecasts. A further improvement of wind analysis and forecast over the ocean will be crucial to increase the quality of wave computations.

Wave models can be further improved. There are three broad areas for improvement: numerical resolution (both spatial and spectral), numerics (propagation scheme and the integration of the non-linear transfer integral) and physics (i.e. the representation of wave growth and dissipation in the source terms). One should not forget to test "improvements". After all, what really counts is the quality of the wave predictions, when compared with observations. The WAM model has been tested in numerous hindcasts and in many years of operational application. Repeating such tests is a formidable task. Maybe inverse modelling, using an adjoint (Hersbach, 1997) can be useful here.

Wave forecasts can also improve by the assimilation of wave observations. Present operational methods are still based on optimum interpolation (see, for example, Lionello et al, 1992; or Voorrips et al, 1997). The practical applicability of more advanced methods (adjoint, Kalman filter, etc) still remains to be explored.

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