

Wave modelling and altimeter wave height data

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Wave Modelling and Altimeter Wave Height Data

by

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Abstract

The relevance of satellite data for the benefit of society is perhaps most pronounced in the use of Altimeter wave height data in the specification of the initial sea state for ocean wave forecasting. Sea state forecasting started more than fifty years ago when there was a need for knowing the wave state during landing operations in the second world war. In particular, the last decade, has seen a rapid development in this field, partly stimulated by the prospect of the availability of wave data from satellites.

Ocean wave forecasting has a significant value for society through use in ship routing, fisheries, coastal protection and oil exploration. Altimeter wave height data have been of great help in improving the quality of the wave forecast by using it to specify the initial sea state and as validation tool of the wave forecast. This presentation briefly describes the present state of modern ocean wave forecasting and the role Altimeter data play in the process of producing an optimal wave forecast. In addition, some of the assumptions that allows one to obtain wave height information from the Radar back scattered pulse are discussed. Comparing ERS-2 Altimeter wave height data with in-situ observations from buoys one may infer the good quality of the Altimeter wave height data, although the ERS-2 wave height is too low by about 10%. A possible cause for this underestimation is discussed and, by using model wave spectra, a method is proposed to remove this problem. The suggested correction may also have implications for the Altimeter-range measurement which is used to determine the mean surface elevation, a relevant quantity in dynamical oceanography.

1. Introduction

The main theme of this paper is that the comparison of Altimeter wave height and windspeed data with the corresponding data from the ECMWF wave forecasting system has resulted in benefits for both satellite observations and modelled wave data.

Sea state forecasting started more than fifty years ago when there was a need for knowing the wave state during landing operations in the second world war. The past 5 decades has seen a development in ocean wave forecasting from simple manual techniques to sophisticated numerical wave models based on physical principles. In particular, in the last decade development in this field was rapid. This was caused by the prospect of the availability of wave data from satellites such as Geosat, ERS-1, Topex-Poseidon, ERS-2. In contrast to the conventional observing systems which could provide only local information on the sea state, Altimeters would be able to provide for the first time wave height information on a global scale.

As a consequence, in the middle of the 1980s a group of mainly European wave modellers, who called themselves the Wave Model (WAM) group, started to develop a wave model from first principles, that is, a model that solves the energy balance equation for surface gravity waves. The source functions in the energy balance included an explicit representation of wind input, nonlinear interactions and dissipation by white capping. As already noted, there was considerable progress to be noted, because there was a clear need of improved wave modelling in rapidly changing circumstances (*SWAMP*, 1985), powerful computer became available and the wave model development coincided with the development of remote sensing techniques

which had the potential of providing information on the sea state on a global scale. A complete account of the first version of this new wave model, called the WAM model may be found in the *WAMDIG paper* (1988).

Early investigations of the quality of the WAM model results were based on comparisons with SEASAT Altimeter wave height data (*Janssen et al, 1989; Bauer et al, 1992*) and with Geosat data (*Romeiser, 1993*). Generally, modelled wave heights, obtained by forcing the WAM model with European Centre for Medium-Range Weather Forecasts (ECMWF) winds showed good agreement with observed waveheights, but there were also considerable differences. *Romeiser (1993)* found considerable regional and seasonal differences between modelled wave height and the Geosat data. During the Southern Hemisphere winter WAM underestimated wave height by about 20% in large parts of the Southern Hemisphere and the tropical regions, while for the rest of the time the agreement was fairly good. These discrepancies could be ascribed to shortcomings in the wave model physics and in the driving ECMWF wind fields, which at the end of the 1980s were too low in the Southern Hemisphere because they were produced by a fairly low resolution (T106) atmospheric model.

The shortcomings in the wave model physics were mainly ascribed to too much dissipation of swell and a not strong enough wind input source function. Corrective action was taken which resulted in WAM cy4 (*Janssen, 1991; Komen et al, 1994*), which became part of the ECMWF's wave prediction system in November 1991. In addition, in September 1991 the resolution of ECMWF's atmospheric model was doubled in the horizontal and nearly doubled in the vertical. Because of the increased horizontal resolution, one would expect a better representation of the surface winds, which could be beneficial for the prediction of ocean waves in the storm tracks, in particular of the Southern Ocean. Therefore, in 1991 there was sufficient confidence in the quality of ECMWF's wind-wave forecasting system that it could be used for the validation of the Altimeter wind and wave products from the ERS-1 satellite which was launched in July 1991. The comparison of Altimeter wind and wave products and the corresponding ECMWF fields helped to identify a number of problems in the retrieval algorithms for wind speed and wave height as discussed by *Hansen and Guenther (1992)* for ERS-1 and by *Janssen et al (1997a)* for ERS-1/2.

Because of the promising validation results ECMWF introduced in August 1993 the assimilation of Altimeter wave heights into the wave forecasting system (*Janssen et al, 1989; Lionello et al, 1993*). Generally, this has led to an improved specification of the wave analysis. This follows from work done by *Bauer and Staabs (1998)* who compared ECMWF wave analyses with TOPEX/POSEIDON Altimeter wave height data and who found an improved correlation between the two after the ERS-1 wave height data assimilation was switched on.

The structure of this paper is as follows. First we briefly discuss the present status of wave modelling at ECMWF where in particular two new developments are mentioned, namely the operational introduction of a coupled atmosphere-ocean wave model and the start of ensemble prediction of ocean waves. We next discuss the present day use of Altimeter wave height and wind speed products, and we describe possible problems we experienced with the interpretation of the satellite data. Finally, a summary of conclusions is presented.

2. Present Status of Wave Modelling

2.1 Brief historical account

Interest in wave prediction started during the Second World War because of the practical need for knowledge of the sea state during landing operations. The first predictions were based on the work of *Sverdrup and Munk (1947)*, who introduced a parametrical description of the sea state and who used empirical wind sea and swell

laws. An important step forwards was the introduction of the concept of a wave spectrum (*Pierson et al*, 1955), but a dynamical equation describing the evolution of the spectrum was not known yet. This step was taken by *Gelci et al* (1956,1957) and others who introduced the concept of the spectral transport equation, which will be introduced shortly. Because of the lack of adequate theories, Gelci et al were forced to use a purely empirical expression of the net source function governing the rate of change of the wave spectrum. Only after the new theories of wave generation by *Phillips* (1957) and *Miles* (1957) had been published and the source function for nonlinear transfer had been derived (*Hasselmann*,1962) was it possible to write down the general expression for the source functions (*Hasselmann*,1962), consisting of three terms representing the input from wind, the nonlinear transfer and the dissipation by white capping. For deep water ocean waves this is the form that is still used today.

Thus, although the principles of wave prediction were already known at the beginning of the 1960s, none of the wave models developed in the 1960s and 1970s computed the wave spectrum from the full energy balance equation. Additional ad hoc assumptions were introduced to ensure that the wave spectrum complies with some preconceived notions of wave development that were in some cases not consistent with the source functions. Reasons for introducing simplifications in the energy balance equation were twofold. On the one hand, the important role of the wave-wave interactions in wave evolution was not recognized. On the other hand, the limited computer power in those days precluded the use of the nonlinear transfer in the energy balance equation.

The relative importance of nonlinear transfer and wind input became more evident after extensive wave growth experiments (*Mitsuyasu*,1968,1969, *Hasselmann et al*,1973) and direct measurements of the wind input to the waves (*Snyder et al*,1981). Eventually, this led in the 1980s, after powerful computers became available, to the development of a wave prediction model that computed the wave spectrum by integration of the energy balance equation without any prior restriction on the spectral shape. Denoting the two dimensional frequency(f)-direction(θ) wave variance spectrum by $F(f,\theta)$, the energy balance equation for deep water gravity waves gives the rate of change in time of the wave spectrum caused by advection and physics. It reads,

$$\frac{\partial}{\partial t}F + v_g \cdot \nabla F = S_{in} + S_{nl} + S_{ds} \quad (1)$$

where v_g is the group velocity, and the source functions on the right hand side of Eq.(1) denote the rate of change of the wave spectrum due to wind input (S_{in}), nonlinear four wave interactions (S_{nl}) and dissipation by white capping (S_{ds}).

In the present version of the WAM model the wind input term is based on a parametrisation of *Miles'* instability (*Miles*,1957) including the feedback of growing waves on the wind profile (*Janssen*,1989,1991). As a result the drag of airflow over the ocean is sea state dependent in agreement with findings by *Donelan* (1982), *Hexos* (1992), *Donelan et al* (1993) and *Johnson et al* (1998). The sea state dependent drag may have consequences for the atmospheric climate as discussed by *Janssen and Viterbo* (1996).

The nonlinear transfer follows from the work of *Phillips* (1960) and *Hasselmann* (1962) on the resonant energy transfer among four surface waves. In case of gravity waves, the dispersion relation does not permit resonant interactions between three waves. These nonlinear interactions were shown to be important in determining the shape of the wave spectrum, whilst they also play a vital role in shifting the spectrum towards lower frequencies (*Jonswap*,1973). Even with present day computing capabilities a wave model based on the exact representation of the nonlinear transfer is not feasible. Therefore some form of parametrisation of S_{nl} is needed. In the WAM model the direct interaction approximation of *Hasselmann et al* (1985) is utilized.

Finally, the least well-known source function is dissipation due to white capping. However, *Hasselmann* (1974) was able to obtain some general constraints on the form of the dissipation source term but a few parameters were still not determined. *Komen et al* (1984) determined these unspecified parameters by insisting that for large times the wave spectrum would evolve towards the *Pierson-Moskowitz* (1964) spectrum. Here, the wind input term was given by the empirical fit of *Snyder et al* (1981) while the nonlinear transfer was based on its exact representation. Later, *Felizardo and Melville* (1995) compared observed overall dissipation rates of waves at sea with rates obtained from the *Komen et al* (1984) expression for S_{ds} and they found good agreement.

Apart from the comparison studies against Altimeter data, mentioned in the introduction, the WAM model has been validated against conventional buoy data by *Zambresky* (1989), while more recent evaluations of the performance of WAM in operational mode are described in *Khandekar and Lalbeharry* (1996), *Wittman et al* (1995) and *Janssen et al* (1997b). Because wave model results are sensitive to the quality of the driving wind field, there have been several studies in the past that use manually analyzed winds which have much lower rms errors. As a consequence, wave results improve dramatically. An example of this may be found in the study of *Cardone et al* (1995) for the SWADE experiment. The sensitivity of wave results to the quality of the winds was recently confirmed by *Janssen* (1998) who verified wind and wave forecasts against analyses over a period of three years, and, using a simple model for error growth, found a very close relation between the forecast error for wave height and wind speed. In practice, wind speed errors dominate the forecast wave height error after day two in the forecast.

Of course, this does not imply that there are no wave model errors, it only means that wave model errors are smaller than the ones associated with the driving wind field. Wave model errors can presumably only be exposed when the error in the wind field is reduced sufficiently. According to *Janssen et al.* (1997b) this requires a reduction in wind speed error of a factor of two from 1.5 m/s to 0.8 m/s. Nevertheless, there have been two studies recently that suggest there are problems with swell energy in the WAM model, but the conclusions of these studies are conflicting. *Sterl et al* (1998) studied results from 15 years of reanalysed ECMWF winds and waves and by comparison with buoy data they found that in cases of swell the WAM model overpredicted swell wave height. On the other hand, *Heimbach et al* (1998) compared swell wave heights from SAR with operational ECMWF results for swell and they concluded that the WAM model had too low swell wave height. There may be several reasons for this apparent conflict. First, these authors have used different measures for the truth, namely buoy data versus SAR data. Second, the reanalyzed results used by *Sterl et al* (1998) did not assimilate Altimeter wave heights, while the operational results studied by *Heimbach et al* (1998) were based on a wave analysis that did use ERS-1 Altimeter data. Third, the reanalysed waves and the operational wave heights were generated with different wind fields.

All the above mentioned points could help explain the discrepancies between the results obtained by *Sterl et al* and *Heimbach et al*. In particular, it is shown in this paper that analysed wave height results are sensitive to the quality of the Altimeter wave height data used in the wave analysis and that compared to buoy data the ERS-1 wave height data are too low. Moreover, even in the Tropics where the sea state is dominated by swell, wave height results depend in a sensitive manner on the quality of the wind field in the Extra Tropics where the swells are generated.

2.2 Operational wave forecasting at ECMWF

Experimental wave forecasting already started in March 1987, giving the opportunity to perform an extensive validation of the model performance. The verification studies of, for example, *Zambresky* (1989) and *Romeiser* (1993), are based on these quasi-operational wave results. Operational sea state forecasting at ECMWF on a global scale started in June 1992. The WAM model at that time had a resolution of 3 deg.

Shortly afterwards a limited area model for the Mediterranean Sea on a 0.5 deg grid was introduced, whilst in August 1993 we started with the assimilation of ERS-1 Altimeter wave height data.

Presently a global version and a limited area version of the WAM model is run at ECMWF. The limited area model, nowadays called the European shelf model, covers the whole North Atlantic, the North Sea, the Baltic Sea, the Mediterranean and the Black Sea. It uses an irregular latitude-longitude grid in which the grid spacing is kept constant at roughly 28 km. The advantage of this choice of grid over the use of spherical coordinates is that near the North Pole the distance in the latitude direction does not decrease so that for feasible time step there are no problems with numerical instabilities of the propagation scheme. The wave spectrum has 25 frequencies and 24 directions. Shallow water effects are included.

The global version of the WAM model also has an irregular latitude- longitude grid with a grid spacing of 55 km. The spectrum has 25 frequencies and 12 directions and shallow water effects are included. The new aspect of the ECMWF wave prediction system is that following the ideas of *Janssen* (1989,1991) the wave model is now part of the ECMWF atmospheric model so that it is straightforward to provide the wave model with surface winds from the atmospheric model more frequently. In addition, from the sea state one can determine the stress induced by the growing ocean waves on the airflow resulting in a sea state dependent drag coefficient. This two-way interaction of wind and waves gives a more consistent momentum budget at the ocean surface and therefore there is a better balance between wind and waves. Presently, the operational atmospheric model has a resolution $T_{l,319}$ (where the subscript l denotes a linear grid), while in the vertical there are 31 layers.

The introduction of a sea state dependent drag coefficient has impact on the evolution of rapidly developing, fast moving lows. An example is given in Fig.1 which shows the day two forecast of a deep low in the North Pacific. Here, the top left panel shows the surface pressure map according to the Control Forecast (i.e. without two-way interaction), the top right panel shows the forecast of surface pressure according to the coupled version of the atmospheric model, the bottom right panel shows the difference between coupled and control, while the bottom left panel gives the verifying analysis. Clearly, considerable differences between coupled and control are to be noted, while the coupled forecast is in better agreement with the analysis.

This example is exceptional because it is a case of a large scale impact. Normally, as expected of physical processes near the surface, the impact of two-way interaction on the atmosphere is relatively small scale. In addition, extreme events are relatively rare, so one should expect only a modest impact of two-way interaction on the forecast scores. Nevertheless, as shown in Fig.2, there is some impact on forecast scores of the Southern Hemisphere. Here, we have verified the forecast of 1000 and 500 mb geopotential height against the verifying analysis by calculating the root mean square error and the anomaly correlation for each day in the forecast. Because of the large variability in the scores the average over 18 ten day forecasts is taken. It is noted that the coupling of wind and waves has impact on the scores at 500 mb geopotential height and even at 200 mb (not shown) This is in agreement with results of *Janssen and Viterbo* (1996) who found from the ensemble mean of 15 seasonal integrations that the sea state dependent roughness affects the whole atmospheric column, supporting the idea that ocean waves modify the momentum budget which results in atmospheric changes of a barotropic nature.

Two-way interaction appears to have a considerable impact on the scores for surface winds and wave height. As an example, we compare in Fig. 3 mean forecast error and standard deviation of forecast error of wave height of coupled and control run. The area is the Tropics and the number of 10 day forecasts is 74. There are clear reductions of the mean forecast error in wave height and this removes a long standing problem of systematic forecast error growth in our wave forecasting system. In order to illustrate this forecasting problem we display in Fig.4 the monthly mean systematic error in the wave height forecast for different forecast ranges

as function of time since August 1994. The wave analysis was regarded as the truth where it is noted that before May 1996 ERS-1 Altimeter data were assimilated while from May 1996 and onwards ERS-2 Altimeter wave height data were being assimilated. In 1994 systematic errors in the medium range amounted to about 20% of the mean wave height, which is not inconsiderable. However, owing to changes in the atmospheric model (e.g. April 1995), and to the change of Altimeter data in our wave analysis (in May 1996) systematic errors have reduced to a lower level of between 5-10% of mean wave height. With the operational introduction of a coupled atmosphere-ocean wave model on June 29 1998 the systematic forecast error of wave height has virtually disappeared. Note that a similar reduction in systematic forecast error is found in the wave scores for the Southern Hemisphere.

The reduction of systematic forecast error of wave height in the Tropics may be understood as follows. The main component of the sea state in the Tropical area consists of swell. These swells are mainly generated by storms in the Extra Tropics. Usually, atmospheric models parametrize the momentum loss at the ocean surface by means of a sea state-independent drag coefficient. For a logarithmic wind profile the drag coefficient at height $z = L$ reads

$$C_d = \left(\frac{\kappa}{\log L/z_0} \right)^2 \quad (2)$$

where the roughness length z_0 is given by the Charnock relation (1955),

$$z_0 = \alpha u_*^2 / g \quad (3)$$

with u_* the friction velocity, g acceleration of gravity and α the Charnock parameter, which in previous versions of the ECMWF atmospheric model has the constant value of 0.0185. With the introduction of the two-way interaction between wind and waves, the Charnock parameter is not a constant but depends on the sea state. Following *Janssen* (1991) one finds

$$\alpha = 0.01 / \sqrt{(1 - \tau_w / \tau)} \quad (4)$$

where $\tau = \rho_a u_*^2$ is the total stress in the air whilst τ_w is the wave-induced part of the stress, which can be determined when the wind input source function S_{in} of the energy balance equation is known. Young windseas, which are ocean waves that are just generated by wind, are usually much steeper than old windseas (*JONSWAP*, 1973). As a result, for young windsea the contribution of the wave-induced stress to the total stress is larger than for the old windsea giving a sea state dependent Charnock parameter and a sea state dependent drag coefficient. More importantly, in practice, for young windsea the Charnock parameter will be larger than the nominal value of 0.0185 used in the ECMWF model and therefore there is an extra slowing down of the surface wind field in these circumstances. The consequence is that, because of the lower wind speeds in case of two way interaction, wave heights in the Extra Tropics will be lower than in the control run with constant Charnock parameter. Therefore, a reduction of the systematic forecast error of wave height is found in the Extra Tropics, which is the main source for the swells in the Tropics. The reduction of mean wave height in the Extra Tropics caused by the two-way interaction was also found in the seasonal integrations of *Janssen and Viterbo* (1996).

This example of positive impact of two-way interaction on the reduction of systematic errors in the Tropics also serves to illustrate that even for areas which are dominated by swell it is not easy to separate wave model errors from errors in the surface winds. From the above example it is clear that the elimination of relatively

small errors in the driving wind fields in the Extra Tropics has resulted in considerable improvements of the swell field in the Tropics. On the other hand it could be argued, as *Sterl et al* (1998) suggested, that there is an error in the dissipation source function so that there is not enough swell dissipation (note that this suggestion is at variance with results of *Heimbach et al* (1998)). However, the quality of the driving wind fields has improved as well since the operational introduction of two-way interaction on June 29 1998. This follows from a comparison of first-guess winds (a 6 hour forecast) with scatterometer winds from ERS-2, as shown in Fig.5. This Figure shows every 6 hours the bias and standard deviation of error of the difference between the scatterometer winds (the so-called UWI product) and the first-guess winds. Note the sudden drop in standard deviation of about 20 cm/s (which means 10% of the total error) around the 29 of June 1998. Although this is not so clearly visible from Fig.5, the bias is reduced by about the same amount. A similar change was noted by comparing analysed winds with Altimeter wind speeds (not shown). The improvement of the wind field persists during the forecast (not shown).

2.3 Future developments

At ECMWF there is a continuous effort to improve analysis and forecast of the weather and ocean waves. Since the end of 1992 NCEP and ECMWF started with weather ensemble forecasting in order to be able to obtain information on the uncertainty of the deterministic forecast. The coupling of wind and waves make ensemble prediction of ocean waves a feasible option and therefore since the 29th of June 1998 also ensemble wave products are generated. The present ensemble prediction system consists of the coupling of the T₁159 version of the ECMWF model with the 1.5 deg version of the WAM model. The ensemble consists of 50 members which are generated by perturbing the deterministic atmospheric analysis by the most unstable singular vectors. Presently, we are studying the information content of our ensemble prediction system for waves by running a ship routing program on a standard North Atlantic route using information on surface wind and waves from each member of the ensemble. In this fashion 50 different ship routes are generated which will give information on the variability around the most likely route.

A first example illustrating the usefulness of the ensemble wave prediction system is given in Fig.6 where we show 10 day forecasts of surface wind, peak period and wave height for the buoy 51001 northwest of Hawaii. The forecast starts at the 23rd of June 1998. The deterministic forecast (T₁319) of the peak period suggests the arrival of low-frequency swell after about 5 days in the forecast, while the majority of the members of the ensemble suggest that no swell arrives. The absence of swell is confirmed by the analysis and also by the deterministic and ensemble forecasts of the following day (not shown). Although this example suggests there is future in probabilistic forecasting of waves, it should be clear that an extensive investigation is required in order to show this in a convincing manner. This is presently under way.

An important aspect of sea state forecasting is the ability to predict extreme events such as deep extra-tropical storms or hurricanes. However, weather models have problems with a good representation of the wind fields in such cases because of lack of resolution, and an inaccurate or perhaps inadequate representation of the physical processes involved in the development of a low. Nevertheless, in recent years we have seen at ECMWF considerable progress on this issue. We illustrate this in Fig. 7 which shows the 36 hour forecast of hurricane Luis in its extra-tropical phase. The start date of the forecast was the 9th of September 1995 at 12 UTC, the experimental forecast started from its own analysis which was generated with the 4DVAR system, while the operational analysis was based on Optimum Interpolation. Here, the left top panel shows the operational forecast at that time, while the right top panel shows the forecast with one of the latest versions of the ECMWF forecasting system (CY18R6) at resolution T₁639/31L. Note that the verifying analysis (not shown) may be regarded as a poor one since the observed minimum pressure according to the National Hurricane centre was about 965 hPa while the analysed one is only 991 hPa. Also, the location of the low is according to the observations slightly more to the North East. Comparing the two forecasts we notice that the

new version of the ECMWF model not only gives a much deeper low but also that its location is in better agreement with the observed one. The consequences for wave prediction are shown in Fig.7 as well and in view of the large difference in maximum wave height of more than 7 m, may be regarded considerable.

An attempt was made to try to trace back reasons for the large differences in the simulation of hurricane Luis. However, it should be clear that given the large amount of changes in the ECMWF forecasting system during the last three years only a partial explanation for the cause of the differences seems feasible. In order to determine whether the variational data assimilation method caused the differences we performed a set of forecast experiments starting from the operational O.I analysis and from the 4DVAR analysis at T₁319 for different resolutions. In Table 1 we show results of maximum wave height for hurricane Luis at the 36 hour forecast time.

	oper_ana(OI;T213	own_ana(4DVAR;T ₁ 319
T213	12.5m	13.0m
T ₁ 319	14.0m	15.7m
T ₁ 639	15.7m	16.7m

Table 1: Maximum forecast wave height for hurricane Luis at 1998091100 for different resolutions and different initial conditions. Observed maximum wave height at buoy 44141 was just below 17 m, while the operational wave forecast had 9.6 m.

From Table 1 it is seen that the change of data assimilation method has resulted in a relatively minor improvement of forecast wave height, while improvements due to the change of forecast model are more dramatic, and, as expected also resolution plays an important role. It is remarked that the relative insensitivity of forecast maximum wave height to the change of assimilation method is not typical. This is probably related to the particular circumstance that the low was small scale and the variational assimilation system only affects relatively large scales, because in the minimisation of the cost function the atmospheric fields are truncated at T63 resolution. Regarding impact of model changes, the most likely candidate for the improved forecast performance of hurricane Luis seems to be an improved representation of convection (*A. Beljaars*, private communication) introduced operationally at the end of 1997, although recent changes in the semi-Lagrangian scheme may have contributed to the improvement as well.

The simulation of hurricane Luis suggests that at the present operational resolution (T₁319) realistic results for extreme cases are already obtained while the future looks even more promising because in a couple of years it is expected that there is a further increase of the horizontal resolution of the operational atmospheric model.

3. Use of Altimeter wave height and wind speed

In this Section we shall describe the benefits Satellite data, in particular from the Altimeter, have for our wave analysis and forecasting system and how on its turn the quality of these observation has been improved by comparison with wind and wave model products and in situ observations. In this discussion we shall restrict ourselves to Altimeter products which are available almost immediately (within 3 hours) after the remote sensed observations have been made. The reason for this strict requirement on availability is that operational weather centres wish to produce a wave analysis as soon as possible. These products are the so-called Fast Delivery products of ESA's ERS-1 and ERS-2 satellite.

In the first instance it should be noted, however, that wind and wave model products have been quite useful in the validation and calibration of the ESA's Altimeter wind and wave product. The necessary calibration and validation of a satellite sensor requires large amounts of ground truth data which should cover the full range of possible events. In particular the number of reliable wave measurements is very limited and, because of financial restrictions, dedicated field experiments are only possible at a few sites. In contrast to that, model data are relatively cheap and provide global data sets for comparison. Thus, the combination of both in-situ observations and model data seems to be an optimal cal/val data set. During the ERS-1 and ERS-2 cal/val campaigns the Altimeter-Model comparisons have been very effective in identifying errors and problems in the Altimeter software and retrieval algorithms. Just after the launch of ERS-1, the global mean Altimeter wave height was about 1 m higher than computed by the model. The investigation of the detected bias led to the discovery of a small offset in the pre-launch instrument characterisation data. When the processing algorithm was updated at all ground stations the performance of the Altimeter wave height was found to be satisfactory as follows from an almost zero wave height bias and a standard deviation of error of 0.5m. During the operational phase of ERS-1 another bug in the processing algorithm was discovered which led to unrealistically shaped wave height distributions. This bug was removed in the beginning of 1994 and resulted in a much improved shape of the histograms at low wave height.

Regarding the Altimeter winds, engineering and geophysical calibration could not be separated, as there is no access to independent data from man-made targets or stable known targets of opportunity. For the initial data calibration the system gain as determined by pre-launch instrument characterisation was used and for the initial geophysical calibration algorithms from previous missions such as SeaSat and Geosat were used. First comparisons with ECMWF winds uncovered several problems in the initial algorithm. The problems were solved in a couple of weeks but differences of 20% remained. This difference corresponds to a small (0.8 dB) bias in antenna gain. After thorough validation of the ECMWF reference data set it was shown that the observed antenna gain was well within the error budget for pre-launch characterisation. The data calibration was updated in early December 1991 and the quality of the ERS-1 altimeter wind product has been quite good since then.

Having learned from the ERS-1 experience, the validation of the ERS-2 Altimeter wind and wave products was relatively straightforward. In addition, when ERS-2 was launched the ERS-1 satellite was still operational enabling an intercomparison between ERS-2 and ERS-1 products, using the corresponding model products as a go-between. The validation of the ERS-2 Altimeter wind speeds was therefore relatively easy compared to the ERS-1 exercise. The main problem was again to determine the antenna gain factor. By comparing the histograms for the radar back scatter σ_0 from the two satellites the mean difference between the two gave the antenna gain bias. A comparison of the retuned Altimeter wind speeds and the analysed ECMWF surface winds was favourable and showed that the tuning procedure was sound. The ERS-2 Altimeter wave heights showed from the first day onwards a remarkable good agreement with the first-guess modelled wave height, except at low wave height where ERS-2 had a higher cut-off value than ERS-1. This higher cut-off value is caused by the somewhat different instrumental specifications of the ERS-2 Altimeter. Because of the tandem mission it was possible to compare ERS-1 and ERS-2 wave heights using the WAM model data as reference standard. As a result it was found that ERS-2 Altimeter wave heights were 8% higher than the ones from ERS-1. This change was regarded as favourable because comparison of ERS-1 wave heights with buoy data had revealed an underestimation of the truth by ERS-1 (see e.g. Janssen et al, 1997a). Since both Altimeters use the same wave height algorithm the improved performance of the ERS-2 Altimeter (which also follows from a comparison of ERS-2 Altimeter wave heights and buoy data, see Janssen et al, 1997a) is probably related to a different processing of the data. Indeed, the on-board processor on ERS-2 uses a more accurate procedure to obtain the wave form, which results in better estimation of wave height (R. Francis, private communication, 1997).

After having discussed the role of wind and wave model products in validating and calibrating the Radar Altimeter wind and wave products let us now describe the role these satellite products have played in the improvement of wave model analysis and forecast. We remark that, as already mentioned, the Altimeter wave heights derive their value already from its use in the process of the production of the initial wave analysis. Note, however, that in the ECMWF wind analysis no Altimeter wind speed data are assimilated so these data may be regarded as independent. In the following we concentrate on the use of the Altimeter products as a validation tool for changes in the ECMWF forecasting system.

3.1 Wind speed data and problems

Although there is a considerable scatter in the relation between the Radar backscatter σ_0 and wind speed, the Altimeter wind speed product has nevertheless proven its value. As an example we mention the validation of a new method of weather analysis namely the variational method. Its static version, called 3DVAR, was introduced operationally by the end of January 1996 and had in general a positive impact on the forecast performance of the ECMWF weather forecast system (*Anderson et al*, 1998). When 3DVAR became operational, the new surface wind observations from the ERS-1 scatterometer were also presented to the data assimilation system. The scatterometer data had a positive impact on the quality of the surface wind speed analysis as follows from a verification of analysed wind speed in the Southern Hemisphere with the Altimeter wind speeds from ERS-1 shown in Fig.8. The Figure gives the verification results for the Optimum Interpolation analysis without scatterometer data as well and it is concluded that with scatterometer data there is a favourable reduction of the standard deviation of error of 10% which is not inconsiderable. This shows at once the considerable value of the Altimeter winds as a validation tool for operational model changes. Since these winds are not used in the analysis system they may be considered as an independent source of information on the quality of the surface analysis, which makes them even more valuable.

Nevertheless from long term monitoring of the performance of the quality of the Altimeter winds we have found that in the Northern Hemisphere there seems to be a clear underestimation of wind speed by the Altimeter during the late spring and early summer, whereas in the remainder of the year there is hardly any bias. This problem is illustrated in Fig.9 which shows over the period of December 1996 until October 1998 a clear seasonal cycle in the bias between ERS-2 Altimeter windspeed and analyzed wind speed for the Northern Hemisphere, whilst no such cycle is seen in the Tropics and in the Southern Hemisphere. Initially, it was thought that there was a problem with the model winds in the Northern Hemisphere, but a validation of Altimeter wind speed against wind speeds from buoys mainly located in the Northern Hemisphere storm tracks gives a similar picture. This is seen in Fig. 10 where a scatter diagram between the two is shown. Note that the underestimation of wind speed by the Altimeter mainly occurs in the wind speed range below 10 m/s.

There may be several reasons for the occurrence of a seasonal cycle in the Northern Hemisphere. First of all it is known that slicks of natural origin play an important role in damping the high frequency ocean waves (*Alpers and Huenerfuss*, 1982). Thus in the presence of slicks the ocean surface appears to be smoother resulting in a larger backscatter than when slicks are absent. Hence, when a retrieval algorithm is used that does not take slicks into account a lower wind speed will be retrieved because the radar backscatter is inversely proportional to the wind speed. Since slicks will be washed away at winds above 7-10 m/s this underestimation of wind should be most pronounced in the low wind speed range which agrees with the findings from the validation of Altimeter against buoy winds (cf. Fig.10), and also against analysed wind speed (not shown). In order to investigate whether slicks are relevant for the underestimation of the wind speed by the Altimeter during the Northern Hemisphere Summer the spatial distribution of slicks is required. Information on this was obtained from monthly mean observations of the ocean colour by SeaWiFs, which are displayed as global maps on the Web. Since ocean colour is a measure for biological activity in the upper layer of the ocean and since natural slicks are produced by plankton it seems plausible to relate ocean colour to the

presence of slicks. In doing so it was found that in the spring and summer of 1998 the main area of biological production was in the North Atlantic where production peaks in May. By contrast, in the North Pacific biological production is only confined to coastal areas (for example, near Japan), thus when the presence of slicks would be relevant for the bias problem of Altimeter winds one would expect that the main problem should occur in the North Atlantic in May 1998. This inference is, however, not supported by the present validation of Altimeter winds against analysed ECMWF winds nor by its validation against buoy winds. According to our verification of Altimeter winds the main bias problem seems to occur in June and July when there is relatively little biological activity compared to May. Also, studying the verification statistics for the North Pacific and North Atlantic separately, it follows that the underestimation of wind speed by the Altimeter is similar in both ocean basins. Therefore, although effects of slicks are expected to be relevant in the retrieval of wind speed from an Altimeter, it appears that an explanation of the bias problem in terms of a single cause such as the presence of slicks is not very likely.

An alternative explanation may be found by noting that the radar backscatter is effected by the tilting of short waves by long waves (*Hwang et al.*, 1998). This effect plays an important role in the theory of scatterometry (see *Plant*, 1990, for a review) and it is expected that this effect is relevant for a nadir looking Radar as well. Concentrating now on the problem of specular reflection it is well-known that the local incidence angle of a short wave riding on a long wave depends on the slope of the long wave. Thus the radar backscatter cross section becomes

$$\sigma_0(\theta_i) = \frac{|R(0)|^2}{s_s^2} \int \frac{d\theta}{\cos^4(\theta + \theta_i)} \exp(-\tan^2(\theta + \theta_i)/s_s^2) p(\theta) \quad (5)$$

where the local incidence angle is $\theta_i + \theta$, θ is the angle of the long waves that give rise to the tilting, $p(\theta)$ is the probability distribution of the tilting waves, σ_0 is the radar cross section and s_s is the root mean square slope of the short waves. Assuming a Gaussian distribution for the slope of the uni-directionally propagating long waves,

$$p(\theta) = \frac{1}{s_l \sqrt{2\pi}} \exp(-\tan^2 \theta / 2s_l^2) / \cos^2(\theta) \quad (6)$$

the radar back scatter for normal incidence becomes

$$\sigma_0(0) = \frac{|R(0)|^2}{s_s^2 s_l \sqrt{2\pi}} \int \frac{d\theta}{\cos^6 \theta} \exp(-\alpha \tan^2(\theta)) \quad (7)$$

where $\alpha = 1/s_s^2 + 0.5/s_l^2$. The integral over θ may be evaluated analytically and the final result for σ_0 becomes

$$\sigma_0(0) = \frac{|R(0)|^2}{s_s^2} \sqrt{\frac{s_s^2}{s_s^2 + 2s_l^2}} (1 + 1/\alpha + 1/\alpha^2) \quad (8)$$

In the absence of tilting waves, $s_l \rightarrow 0$, we recover the usual result that for specular reflection the backscatter is inversely proportional to the mean square slope s_s^2 , while in the presence of long waves in general a

reduced backscatter due to tilts is found. This is illustrated in Fig.11 where we show for a Ku-band Altimeter(13.6 GHz) the effect of tilting long waves on the relation between σ_0 (in dB) and wind speed at 10 m height. Here, we used an empirical relation of *Hwang et al* (1996) to relate mean square slope up to the Ku-band range to the surface wind speed. Furthermore the tilt s_l of the long waves was obtained by integration of the Phillips spectrum from the peak frequency until 3 times the peak frequency. This gives approximately $s_l = \sqrt{\alpha_p}$ where α_p is the Phillips parameter and in the plot we took $\alpha_p=0.01$. Fig.11 shows that there is a considerable impact of long waves on the relation between σ_0 and surface wind speed. Ignoring this dependence on tilts and assuming that the σ_0 to U_{10} relation has been tuned to cases with tilts we see that for low wind speed the true wind may be underestimated by 2 m/s which is comparable in magnitude to the aforementioned bias problem of the Altimeter wind speed.

It is important to note now that in the Northern Hemisphere the magnitude of the tilt shows a seasonal cycle, being larger during the winter because windseas are important, while during the summer tilts are smaller by nearly a factor of two because of the dominance of swells. Again because of the relative dominance of swells during the whole year, the seasonal cycle in the tilt is much smaller in the Tropics and in the Southern oceans. As a consequence, since tilts effect the relation between σ_0 and the surface wind speed, this may explain the seasonal cycle in the Northern Hemisphere bias of Altimeter wind speed while at the same time it makes plausible why there is no seasonal cycle in the other oceans. It should be clear, however, that so far we have only giving a qualitative explanation, so more work is needed to be convincing.

We conclude from this discussion on the quality of Altimeter wind speeds that we have benefited from these data in the validation of modelled surface winds and in the validation of model changes. There is however still significant progress in weather forecasting because of improvements in data assimilation and modelling. For example, rms errors in analyzed wind speed have decreased in the past three years from 2 m/s to 1.5 m/s, which was caused by the introduction of 3DVAR in January 1996 and by the introduction of the coupled wave-atmosphere model in June 1998. This trend suggests that modelled winds are becoming more accurate which implies that in order to be useful Altimeter winds should become more accurate as well. This probably requires a further development of the Altimeter wind speed retrieval algorithm. We have mentioned already the role of slicks and tilting of the short waves by the long waves, but also effects of the sea state on the relation between σ_0 and surface wind should be taken into account. See for this the interesting work of *Elfouhaily et al* (1998) who compared two TOPEX Altimeters operating at C and Ku band. Sea state effects become relevant, for example, under fetch-limited conditions when the waves are steep and they affect the Altimeter wind speed retrieval because the implicit assumption that there is a unique relation between rms slope of the surface waves and wind speed is then no longer valid.

3.2 Wave height data and problems

In wave modelling there is a keen interest in the quality of Altimeter wave height data because these observations are used in the ocean wave analysis (for example, at UKMO, NCEP and ECMWF), and for validation on a global scale. The wave analysis scheme is quite sensitive to changes in the quality of the Altimeter wave height data, as follows from the experience we had when we made the switch from the use of ERS-1 to ERS-2 wave height data. In Fig. 12 a time series is shown of the bias between analysed wave height and observed wave height of a number of Northern Hemisphere buoys over the period of January 1995 until March 1997. It shows a pronounced jump between April and May 1996 when we switched from the use of ERS-1 to ERS-2 data. As already remarked, the ERS-2 data were 8% higher than the ERS-1 wave height data and this is immediately reflected in the verification of analysed wave height against buoy data.

Therefore, with the introduction of ERS-2 wave height data the bias during the Northern Hemisphere summer (characterised by a sea state with swells and low steepness) is removed but, as can be seen from Fig. 12, there

is still an underestimation of wave height by the analysis during the following Winter, when the sea state has a considerable fraction of windseas with large steepness. Of course, it could be argued that the ECMWF wave forecasting system is underestimating wave height in circumstances of wind wave generation by wind but, verification of the Altimeter wave height data against buoy data, shown in Fig. 13 for February 1997, confirms the underestimation by the Altimeter (in this case by about 7%). This led us to suspect that perhaps there are problems with the retrieval of Altimeter wave height in cases of large steepness, because most wave height algorithms assume that the wave height and steepness distribution follows a Gaussian law which for large steepness when nonlinear effects become important may not be valid. It would thus be natural to study the dependence of the Altimeter wave height error with respect to some measure of the truth on the steepness of the ocean waves. However, if one uses buoy observations as truth a few years of collocated data is needed in order to obtain results that are statistical significant. Using model data as truth requires, however, only a month worth of collocated data, since there are typically about 40,000 collocations between Altimeter and modelled wave height during one month. Hence, we used the first-guess wave heights as truth and the result of the study is shown in Fig. 14 which gives the dependence of Altimeter error on the mean slope of the ocean waves. It is seen that for relatively small slopes when swells dominate the sea state the Altimeter wave height error is small whilst for large slope the Altimeter underestimates wave height by as much as 50 cm. We therefore thought it was worthwhile to study the role of nonlinearity in the problem of the retrieval of wave height from the Altimeter wave form.

4. Nonlinear effects and the Altimeter Retrieval Algorithm

Barrick (1972a,b) derived the average radar cross section for back scatter of randomly distributed specular points in the rough surface approximation (for an overview cf. *Barrick and Lipa* (1985)). The radar cross section is a function of time because contributions from the crests of the ocean waves arrive first at the receiver end of the Altimeter. The time-dependent cross section is called the wave form. For a nadir looking Radar the wave form is found to depend on the joint probability distribution (jpd) of surface elevation and surface slope under the condition of zero slope. In order to obtain a practical Altimeter retrieval algorithm, one normally makes the assumption that the probability distribution has a Gaussian shape which is a fair assumption for weakly nonlinear ocean waves (*Longuet-Higgins*, 1963). Although this results already in good estimates for Altimeter wave height it should be remarked that the Gaussian assumption does not allow to obtain an estimate for corrections in the Altimeter Range measurement, known as the ElectroMagnetic Bias (EMB). EMB mainly results from the fact that ocean waves are weakly nonlinear, having a sharp crest and a wide trough. Thus, it emphasises the part of the sea surface that is below mean sea level and therefore the Altimeter estimates a somewhat longer distance between satellite and ocean surface. For linear waves with small steepness the Gaussian probability distribution is valid and therefore EMB vanishes. However, there is another aspect to sea state induced errors, namely waves may distort the Altimeter pulse, producing errors in the Altimeter's determination of the height of the satellite above the sea surface. This latter error is called the instrumental error, while the sum of instrumental error and ElectroMagnetic bias is called the Sea State Bias (SSB).

Incidentally, it is not needed to make any assumptions regarding the jpd of ocean waves as pointed out by *Lipa and Barrick* (1981) because the jpd may be obtained from the slope of the wave form and the knowledge of the shape of the transmitted pulse. This approach was applied to a few Seasat cases showing some deviations from the Gaussian distribution, but in practice this approach has not been pursued, presumably because noise at the foot of the leading edge of the wave form results in noisy tails of the retrieved probability distribution. The consequence is an unreliable estimate of higher moments of the jpd such as the skewness.

Deviations from the Gaussian distribution may occur for a number of reasons, but here we shall explore the consequences of a nonlinear wave surface with sharp crests and wide troughs on the wave height retrieval and the EMB. Work on this subject began around 1980 (for example, *Jackson (1979)*, *Lipa and Barrick (1981)* and *Hayne and Hancock (1982)*) and started from the study of *Longuet-Higgins (1963)* on the probability distribution of the surface elevation and slope of weakly nonlinear gravity waves. *Srokosz (1986)* extended this approach to the problem of finding the jpd of surface elevation and slopes in two dimensions because oceans waves have a pronounced directional distribution. He then applied this jpd to the problem of finding the expression for the leading edge of the Radar return for a pulse limited Radar. Choosing a Radar pulse with Gaussian shape and width v , the following wave form W was obtained:

$$W \sim \frac{1}{2} \left[1 + \operatorname{erf}(\tau) + \frac{e^{-\tau^2/2}}{\sqrt{\pi}} \left\{ \frac{\lambda\sqrt{2}}{3} (1 + \tau^2) - \frac{1}{\sqrt{2}} (\lambda + \delta) \right\} \right] \quad (9)$$

with $\operatorname{erf}(\tau)$ the error function, τ the normalised time t/t_p , $t_p = \frac{2}{c} \sqrt{(v^2 + H_s^2/8)}$, c the speed of light and H_s the significant wave height. Deviations from Gaussianity are measured by the skewness factor λ and the elevation-slope correlation δ . These parameters can be obtained from the nonlinear corrections to the surface elevation as found by *Longuet-Higgins (1963)* and it turns out that they depend in a complicated way on the wave spectrum.

In order to have some idea how the deviations from Gaussianity depend on the wave spectrum we follow *Jackson (1979)*. Using a Phillips spectrum

$$F(k) = \frac{1}{2} \alpha_p k^{-3} \quad (10)$$

one finds

$$\delta = 2\lambda \quad \lambda = 2\sqrt{\alpha_p} \quad (11)$$

where, as already found by *Srokosz (1986)*, we corrected the expression for λ by a factor of two because there was a factor of two missing in the second order term of the surface elevation obtained by *Longuet-Higgins (1963)*. Now, swell has typically a smaller Phillips parameter α_p (by a factor of 4 to 10) than wind sea, therefore the swell state is in practice Gaussian, while wind sea (with α_p of the order of 0.01 or more) may show considerable deviations from Gaussianity.

In Fig.15 we have plotted the wave form W as a function of time τ for a Gaussian pdf and a non-Gaussian one. The origin, $\tau = 0$, corresponds to the mean sea level and the half power point of the wave form of the non-Gaussian case is shifted towards positive time because in the presence of weakly nonlinear waves the Radar overestimates the distance between mean surface and satellite. For a Gaussian pdf only the first two terms in the expression for the wave form remain. The significant wave height H_s is then determined from the slope s of the wave form at the half power point and the result is

$$H_s = 4 \sqrt{\frac{\kappa_1}{s^2} - \kappa_2} \quad (12)$$

where κ_1 and κ_2 depend on the power and width of the transmitted pulse and on the speed of light. Of course, in this approximation the EMB vanishes.

For a non-Gaussian jpd we retain the full expression for the wave form W and it is assumed that the significant wave height is still determined by the slope at the half power point (which does not coincide with $\tau = 0$). Assuming small deviations from Gaussianity, i.e. λ and δ are small, one may then obtain approximate expressions for the observed wave height H_s and the EMB l . For significant wave height, we obtain a similar expression as for the Gaussian case, except that κ_1 should be replaced by κ_3 ,

$$\kappa_3 = \kappa_1 \left(1 + 2 \left(\frac{\lambda}{3} + \delta \right) \left(\frac{5}{24} \lambda + \frac{1}{8} \delta \right) \right) \quad (13)$$

whilst the correction to the Range measurement becomes

$$l = -\frac{1}{8} \left(\frac{\lambda}{3} + \delta \right) H_s \quad (14)$$

For the Phillips spectrum (10) one then finds

$$\kappa_3 = \kappa_1 \left(1 + \frac{154}{18} \alpha_p \right), \quad (15)$$

and

$$\frac{l}{H_s} = -\frac{7}{12} \sqrt{\alpha_p} \quad (16)$$

Therefore, deviations from Gaussianity have on wave height retrieval a modest impact because the correction depends on α_p which is in the extreme conditions of young windsea at most 0.025, thus at most one would expect a correction to wave height of 10%. On the other hand, the EMB already depends on the square root of α_p so deviations from Gaussianity are in this respect quite important. It is noted that depending on the sea state (wind sea or swell) the relative correction to the range may vary by a factor of two to three, being small for swells, as expected, but being large for young wind seas. This expectation agrees with work by *Minster et al* (1992) who analyzed Geosat data and who found that compared to swell cases young wind seas had SSB's which were larger by a factor of 3.

Srokosz attempted to determine λ and δ directly by fitting the theoretical wave form to the observed one. In general results obtained in that fashion were unreliable, presumably because of noise at the foot of the leading edge of the Radar return. We propose an alternative approach, namely to determine λ and δ using WAM model spectra and consequently we have determined the corrections for Altimeter wave height and Altimeter Range. The period was the month of February 1997 which was characterised by a number of extreme events in the North Atlantic and the corrections were applied to the so-called fast delivery ERS-2 data. In Fig.16 we show the comparison between corrected Altimeter wave height and buoy data which should be compared with Fig.13. The correction due to the nonlinear sea state works in the right direction but it is not enough as on average wave heights only increase by about 3%. This is a rather disappointing result but it should be realised that in deriving the correction for wave height we have assumed that wave height was obtained from the slope

of the wave form at the half power point. This is not the actual procedure used by ESA to obtain fast delivery wave heights (R. Francis, private communication, 1997). And it should be emphasized that the impact of the nonlinear sea state on wave height retrieval depends in a sensitive manner on the procedure of how to obtain the slope of the wave form.

Results on the EMB look, however, more interesting as may be inferred from Fig. 17 where the absolute value of the EMB using WAM model spectra is compared with an empirical correction for SSB obtained by *Gaspar and Ogor* (1996). The empirical approach assumes that, because of the practical reason of availability, SSB is a function of Altimeter windspeed and wave height only. The functional form of SSB is fitted to the difference between the Altimeter range measurements at cross-over points, which assumes that the time between the cross-overs is small enough that there is insignificant variation in the mean sea level while the time difference is large enough so that there is a sufficient change in wave height and wind speed. More details on this method, including an application to TOPEX data, may be found in *Gaspar et al*, 1994. The comparison between the present approach and the empirical one, as given in Fig. 17, shows that, despite the fact that in the present approach we have not taken into account the instrumental error, there is a fair agreement but that there are remarkable differences between the two as well. For low corrections the present approach gives too low a correction while for large corrections a slight overestimate compared to the empirical method is found. Realising that to first approximation EMB depends linearly on wave height, it is concluded that in particular for low wave height, i.e. swell conditions, the present approach underestimates SSB, while for windseas our approach seems to give reasonable corrections. We emphasize that according to Eq. (16) one should expect, compared to windseas, lower corrections for swell, because for swell we deal with nearly linear waves. The question therefore is why the empirical approach gives fairly similar corrections for both wind sea and swell, in other words why the empirical correction gives less sea state dependency.

For this reason we have investigated following *Fu and Glazman* (1991) the seastate dependence of the relative EMB. As a measure of wave development the dimensionless wave height gH_s/U_{10}^2 was taken because wave height and wind speed are readily available from the Radar Altimeter. Remark that for old wind sea the dimensionless wave height is about 1/4 while young windsea has smaller values and swell has larger values for this parameter. We took once more the case of February 1997 but we limited ourselves to those cases where buoy data were available (this therefore corresponds with the collocations shown in Fig.13 and 16). The result from our approach and from *Gaspar and Ogor's* empirical fit is shown in Fig.18 and it is seen that the empirical fit gives hardly no sea state dependence while our approach does. On the other hand, we thought it worthwhile to search for empirical evidence based on direct estimates of the mean sea level and EMB, because the method of *Gaspar and Ogor* is indirect and their error estimation involves the instrumental error as well. This evidence was found in the work of *Melville et al* (1991) and these authors found a dependence of EMB on wave height and wind speed which differs from the one based on cross-overs. For ERS-2, *Gaspar and Ogor* (1996) found

$$\frac{SSB}{H_s} = -(0.048 + 0.0026U_{10} - 0.000126U_{10}^2) \quad (17)$$

while *Melville et al* (1996) found

$$\frac{EMB}{H_s} = -(0.0146 + 0.00215U_{10} + 0.00389H_s) \quad (18)$$

and the main difference between these two fits is that the second one gives an increase with wind speed while the first one gives for large wind speed a decrease with wind speed.

Using Altimeter wave height and wind speed from our restricted data set of February 1997 we determined the relative range correction according to *Melville et al* (1991) and plotted the result in Fig.18 as well. It is seen that Melville's et al fit shows a clear sea state dependence and, in addition, there is a fair agreement between our approach and the latter fit. On the other hand, the Melville et al fit is not in agreement with the result of *Gaspar and Ogor* (1996). It should be emphasized that the reason for the latter discrepancy is not understood, unless it could be argued that the neglect of the instrumental error is solely responsible for the disagreement.

To summarise, our discussion on the Altimeter retrieval algorithms for wave height and Range and the relation to nonlinear wave effects we remark that in our approach the assumption was made that deviations from the Gaussian probability distribution were only caused by the presence of bound second harmonics that gives rise to sharper crests and wider troughs. Regarding corrections to Altimeter wave height we assumed that wave height was obtained from the wave form slope at the half power point. Corrections to Altimeter wave height using modelled wave spectra seem realistic, but are fairly small, being of the order of 3%. The reason for the relatively small size of the corrections is understood, because it is linearly dependent on the Phillips parameter which is small.

Regarding the Altimeter Range we once more emphasize that there may be other effects relevant for a correct determination of the sea state bias. For example, it may be important to include the instrumental error, tilting of short waves by the long waves may be relevant and modulation of short waves by the long ones may play a role. Nevertheless, despite all these shortcomings, Range corrections using the approach of Srokosz and WAM spectra are already in fair agreement with the empirical approach of *Gaspar and Ogor* (1996). This gives confidence in the use of WAM spectra in correcting the Altimeter Range measurements for sea state effects, which is, in addition, supported by agreement with the empirical fit of *Melville et al* (1991). Presently, it is not understood why the Gaspar and Ogor fit for relative SSB does not have a sea state dependency, whereas the present method, the Melville et al fit of the EMB and the work of *Minster et al* (1992) show a clear dependence on sea state. There is evidently more research needed to clarify this issue.

5. Conclusions

In this paper we have pointed out the benefits that Altimeter wave height and wind speed data have for wind and wave analysis and validation of wind and wave forecasts. At the same time it is clear that model data have played an important role in the calibration and validation of the Altimeter products. This approach may result in a potential incest problem but validation against an independent set of buoy data has shown that the procedure that was followed is sound. As a consequence we have seen over the past 10-20 years considerable progress in the field of wave modelling and satellite wind and wave products.

On the other hand, the combined use of satellite and wave model products has also revealed problems. For example, there may be a problem with dissipation of swell in the WAM model although presently it is not clear whether there is too much dissipation or not enough. Furthermore, there may be a problem with the Altimeter wave height retrieval in case of young, nonlinear sea states while tilting of short waves by the longer waves may affect the Altimeter wind speed retrieval. Finally, two-dimensional WAM spectra may provide information on the sea state bias of the Range measurements. This suggests that in the near future interesting studies are expected that attempt to resolve some of the issues mentioned in this paper.

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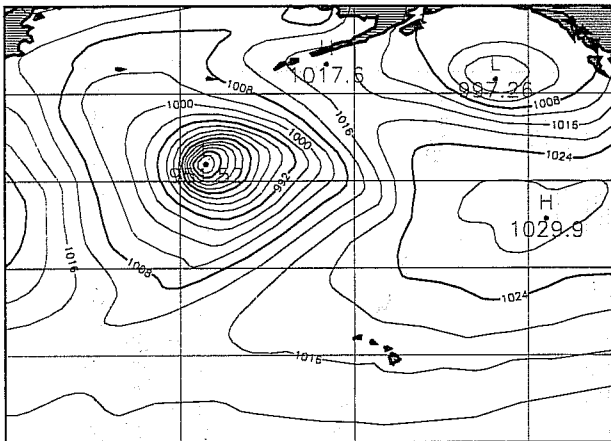
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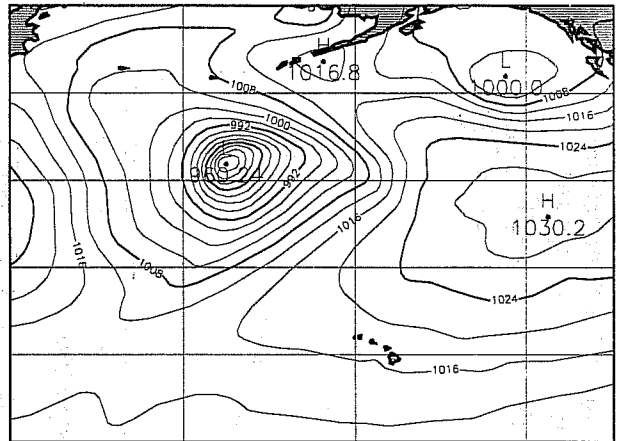
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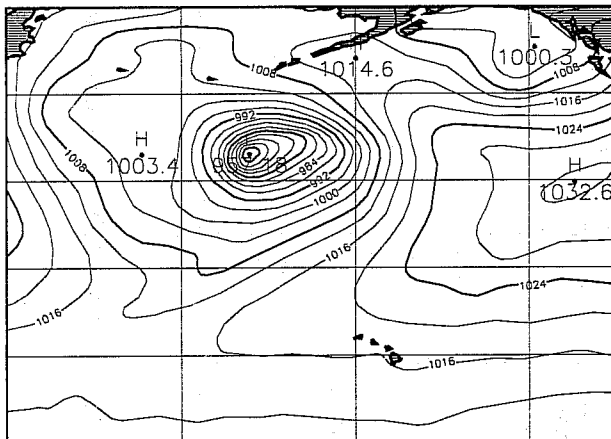
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COUP: MSL Pressure 971224, 48h FC



Oper analysis 971226



Diff COUP-CTRL: MSL Pressure 971224, 48h FC

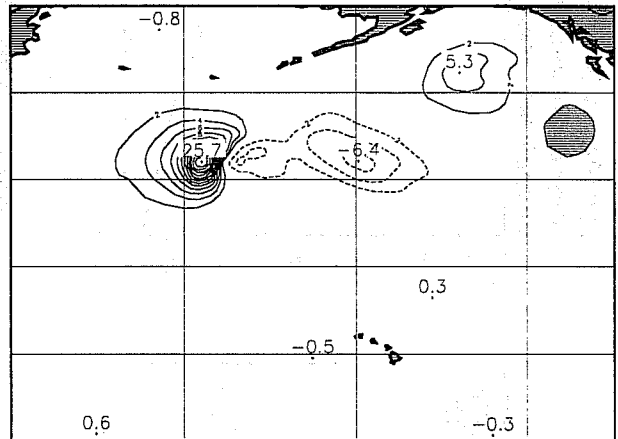


Fig.1 Example of impact of two-way interaction on the day 2 forecast of the mean sea level pressure in the North Pacific. Forecast starts at 1997122412 UT. The top left panel shows the control forecast, the bottom left panel shows the verifying analysis, the top right panel shows the coupled forecast, while the bottom right panel shows the difference between coupled and control.

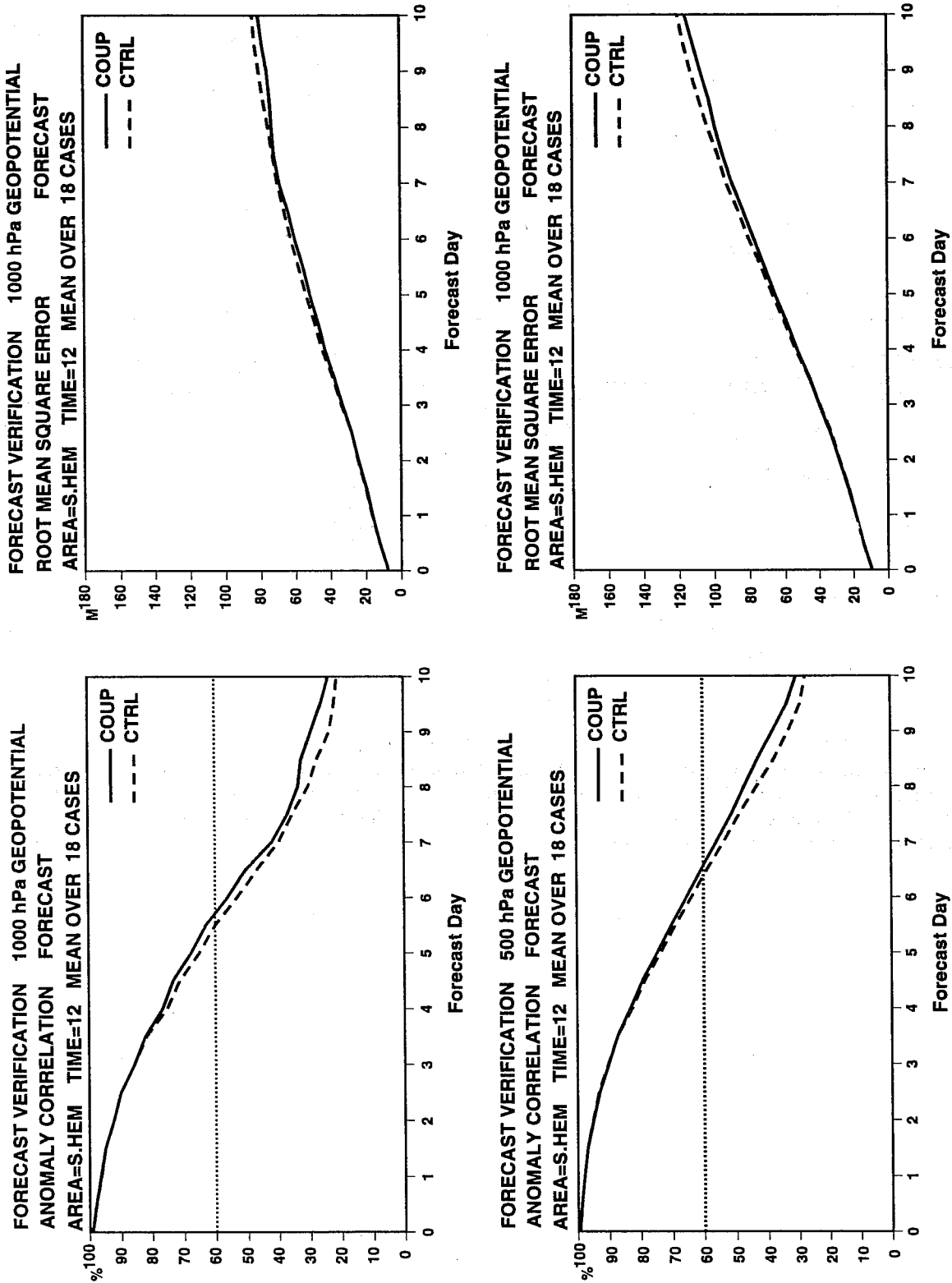


Fig.2 Impact of two-way interaction on the forecast scores of geopotential height in the Southern Hemisphere for an 18 day period in December 1997. Full line: coupled; dashed line: control.

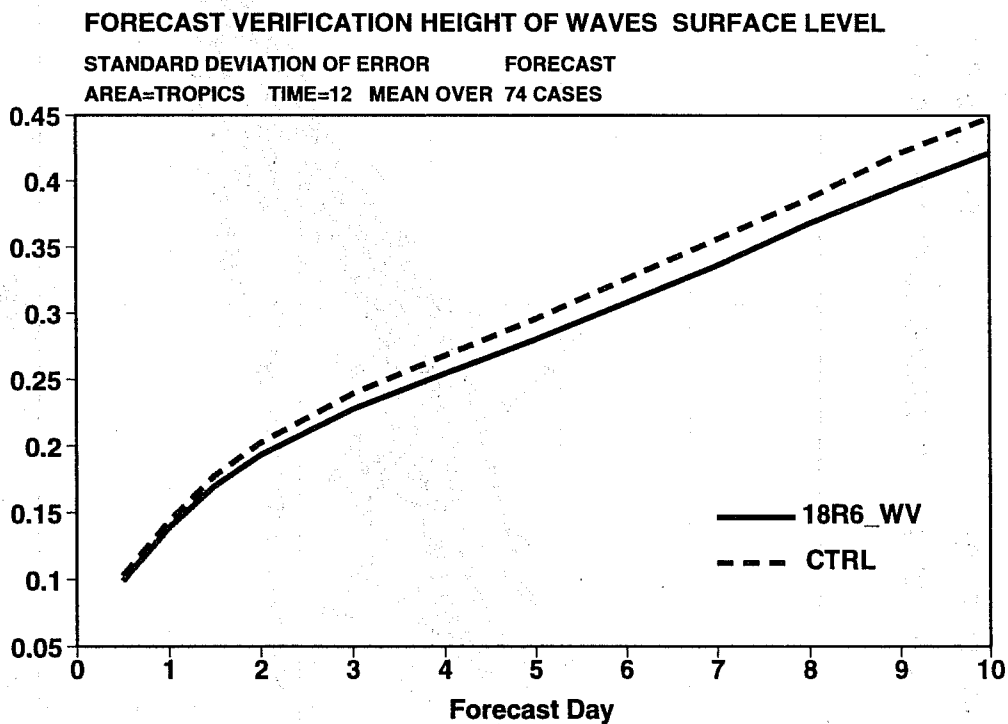
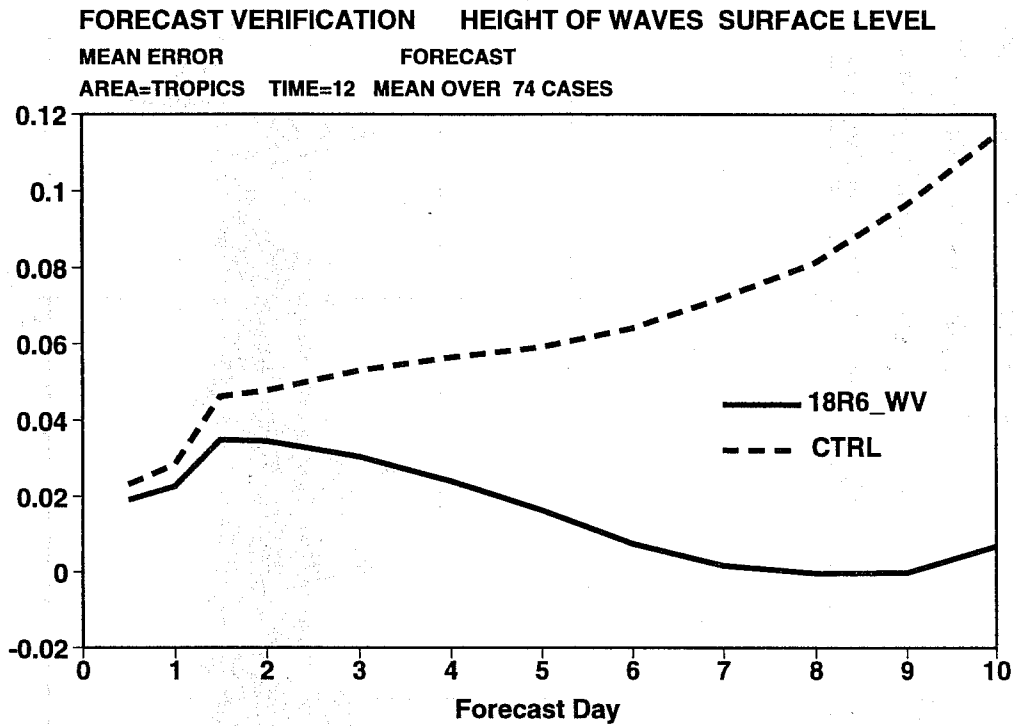


Fig.3 Impact of two way interaction on the forecast scores of wave height in the Tropics for a period of 74 days. Full line coupled; dashed line: control.

WAVE_G FORECAST VERIFICATION 12UTC

HEIGHT OF WAVES SURFACE LEVEL

MEAN ERROR FORECAST

TROPICS LAT -20.000 TO 20.000 LON -180.000 TO 180.000

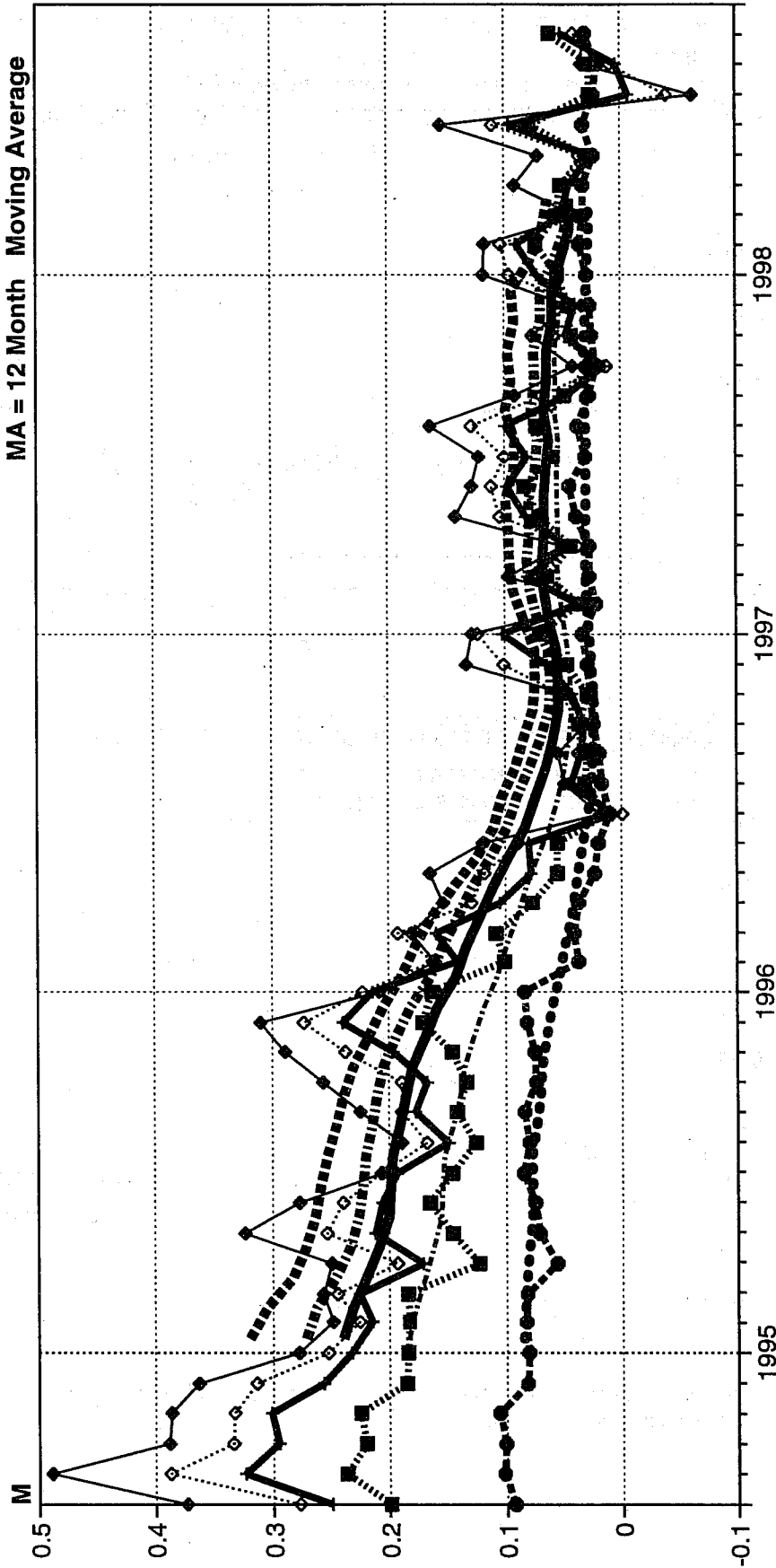


Fig.4 Monthly mean forecast error of wave height for the Tropics since August 1995. Note the marked reductions in mean error in Jan 1996 (introduction of 3DVAR), May 1996 (switch from ERS-1 to ERS-2 wave height assimilation) and July 1998 (introduction of the coupled model).



Monitoring of UWI winds versus First Guess for ERS2

from 19980608 to 19980713

(solid) wind speed bias UWI - First Guess over 6h (m/s)

(dashed) wind speed standard deviation UWI - First Guess over 6h (m/s)

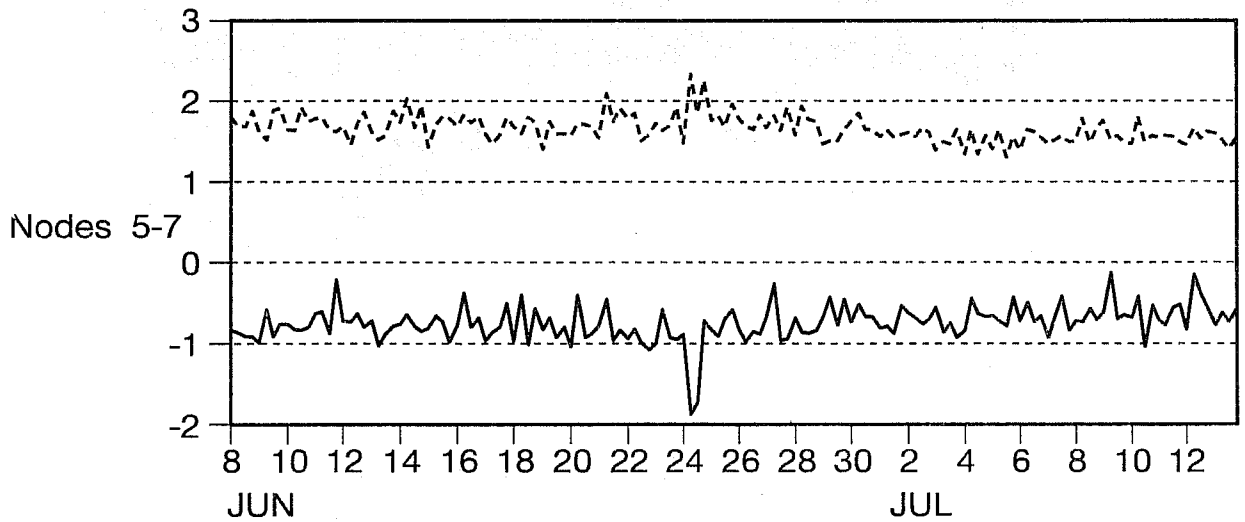


Fig.5 Time series of bias and standard deviation of error of the ERS-2 scatterometer winds as compared to ECMWF's first-guess surface speeds. Note the sudden drop in error when the coupled system was introduced on June 29 1998.

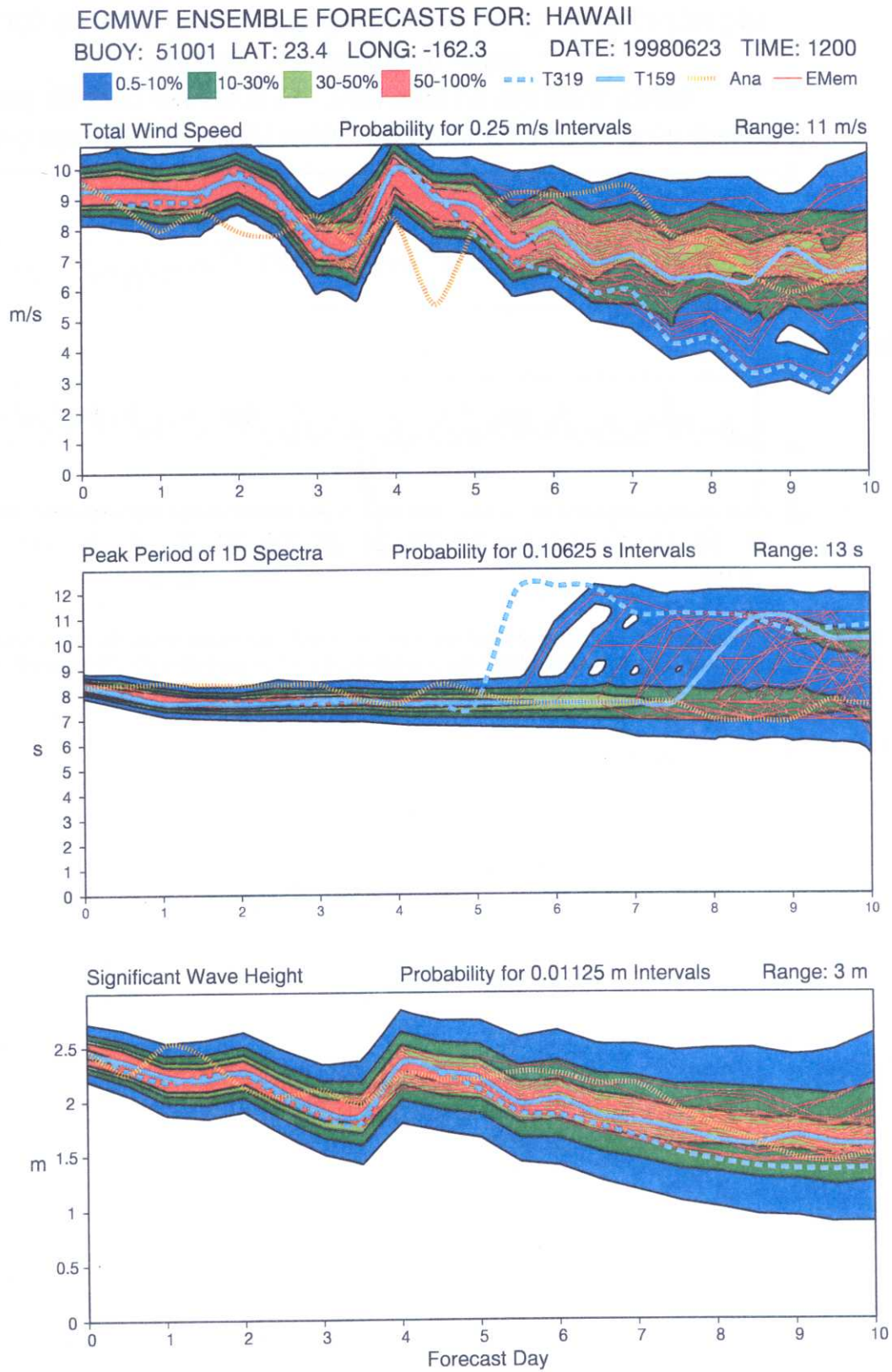
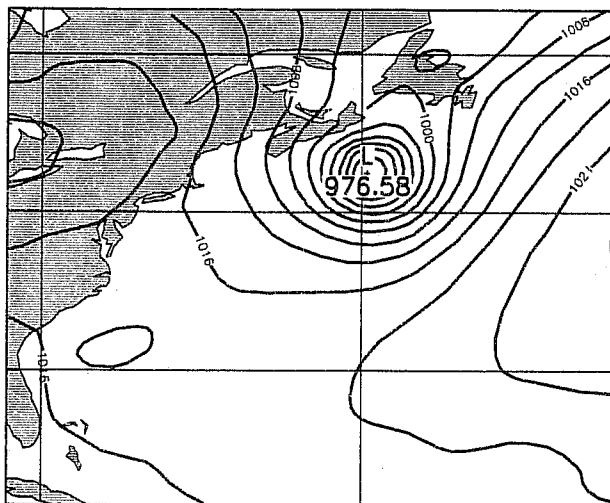


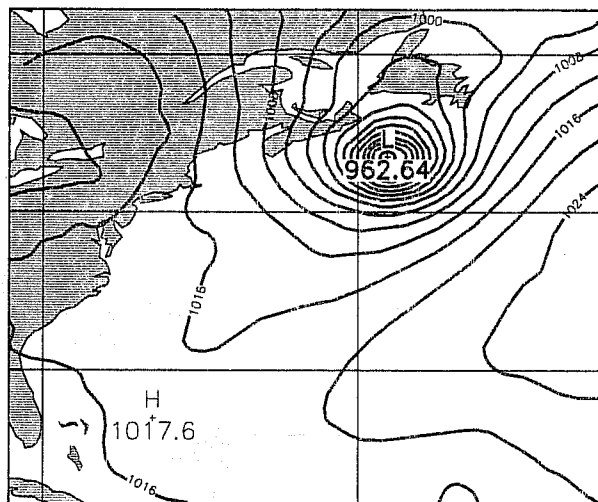
Fig.6 Plume diagrams for wind speed, peak period and wave height for the ensemble forecast of the 23rd of June 1998 for buoy 51001 near Hawaii. Results from the deterministic forecast and the analysis are shown as well.



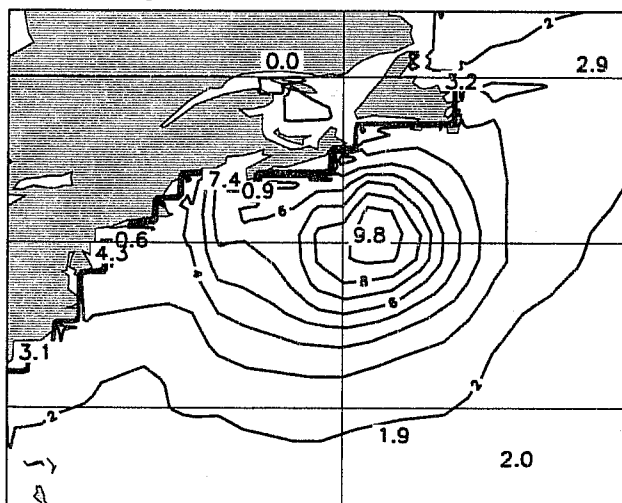
MSL Pressure 36h forecast from 950909



T639 exp: 36h forecast from 95090912



Wave height (1) 36h forecast from 95090912



T639 exp: 36h forecast from 95090912

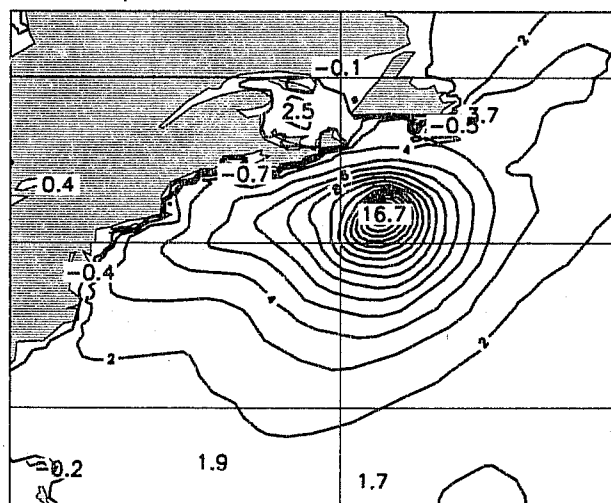


Fig.7 Progress in extreme sea state forecasting as illustrated by the 36 hour forecast of Hurricane Luis. The top left panel shows the operational forecast mean sea level pressure of 1995090912 UT with T213 resolution while the top right panel shows the 36 hr forecast with the present operational system with resolution T₆₃₉. The bottom panels show the 36 hr wave height forecast with the old (left) and new (right) forecasting system.

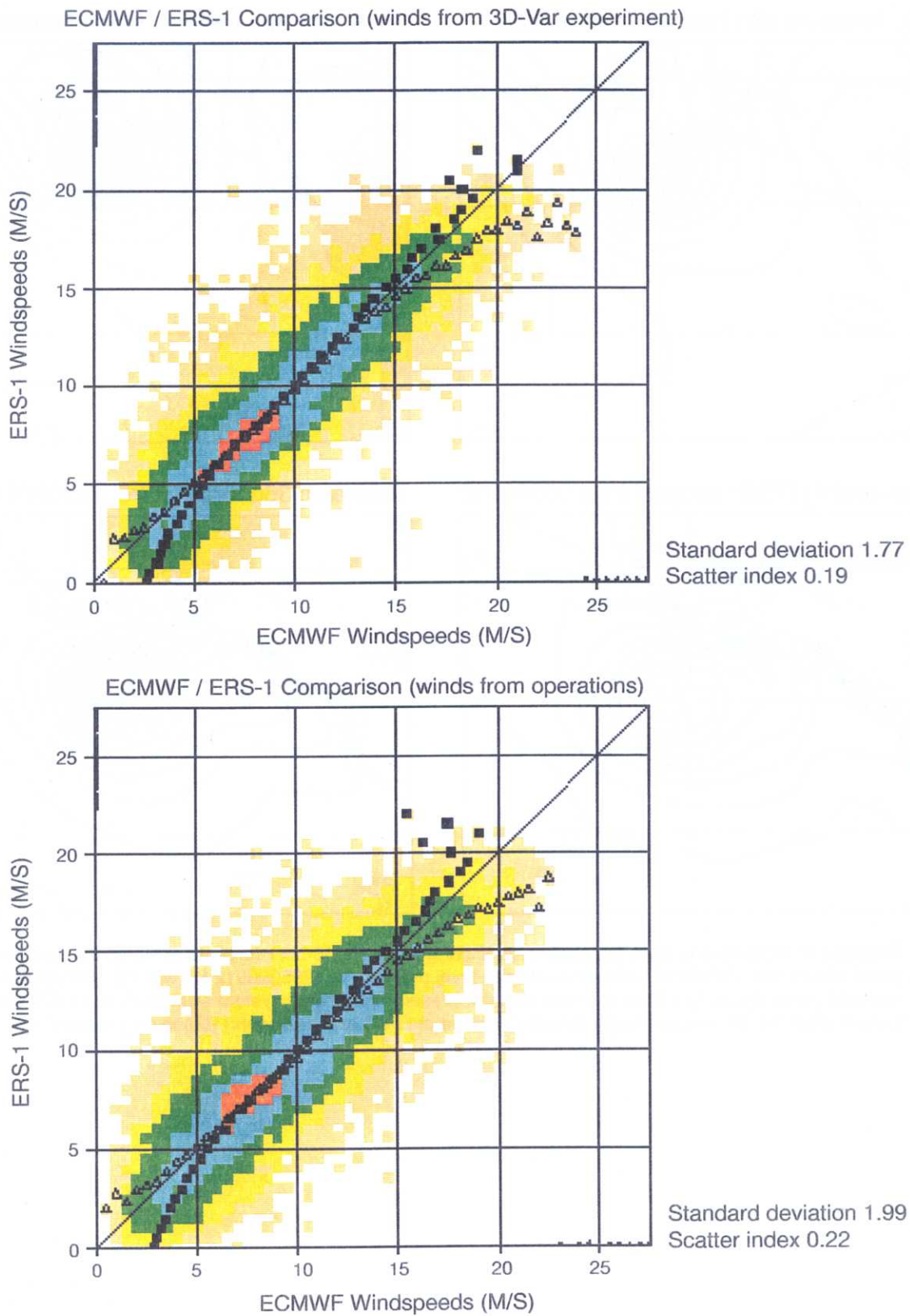


Fig.8 Validating the 3DVAR analysis system, including the use of scatterometer data, by comparing analyzed surface winds with Altimeter wind speeds. Top panel 3DVAR, bottom panel OI.

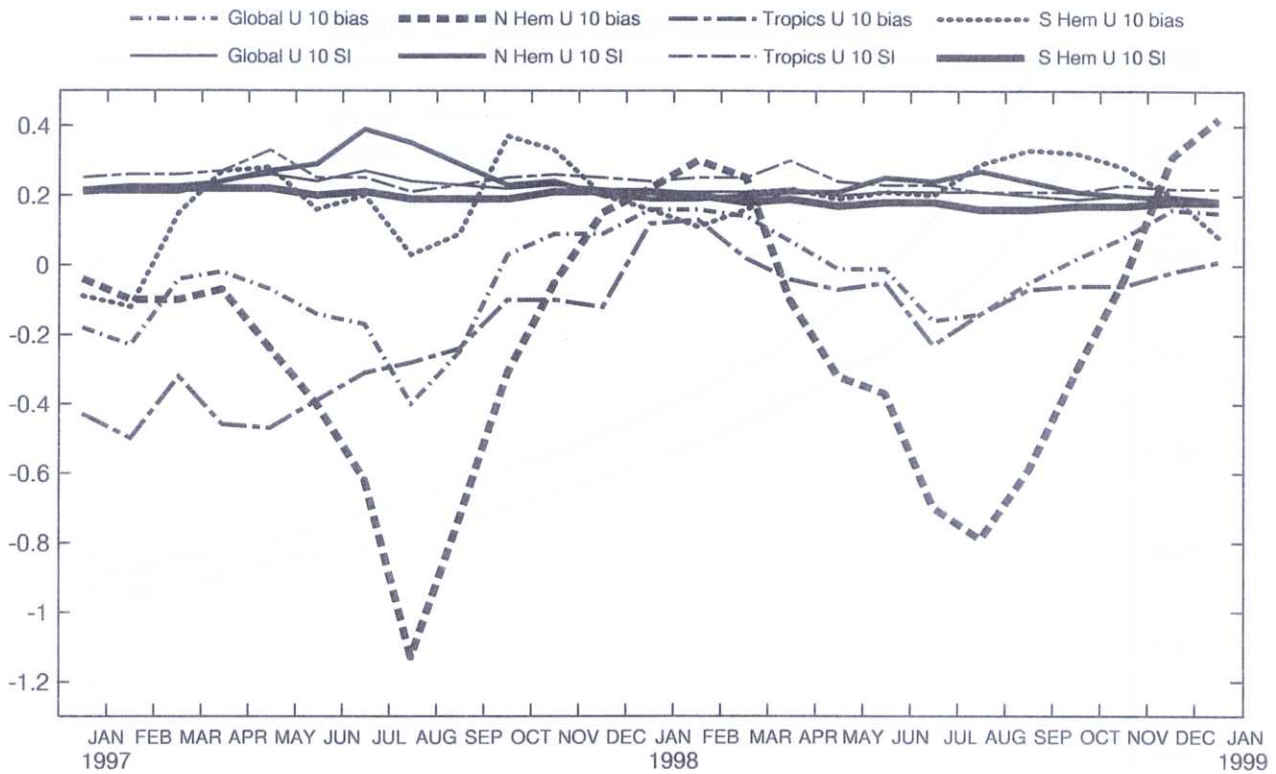


Fig.9 Comparison of Altimeter wind speeds with analyzed model winds for the Globe, the Northern Hemisphere, the Tropics and the Southern Hemisphere. Shown are monthly mean values of bias and scatter index over the period of December 1996 until November 1998. Note the clear seasonal cycle in the bias for the Northern Hemisphere.

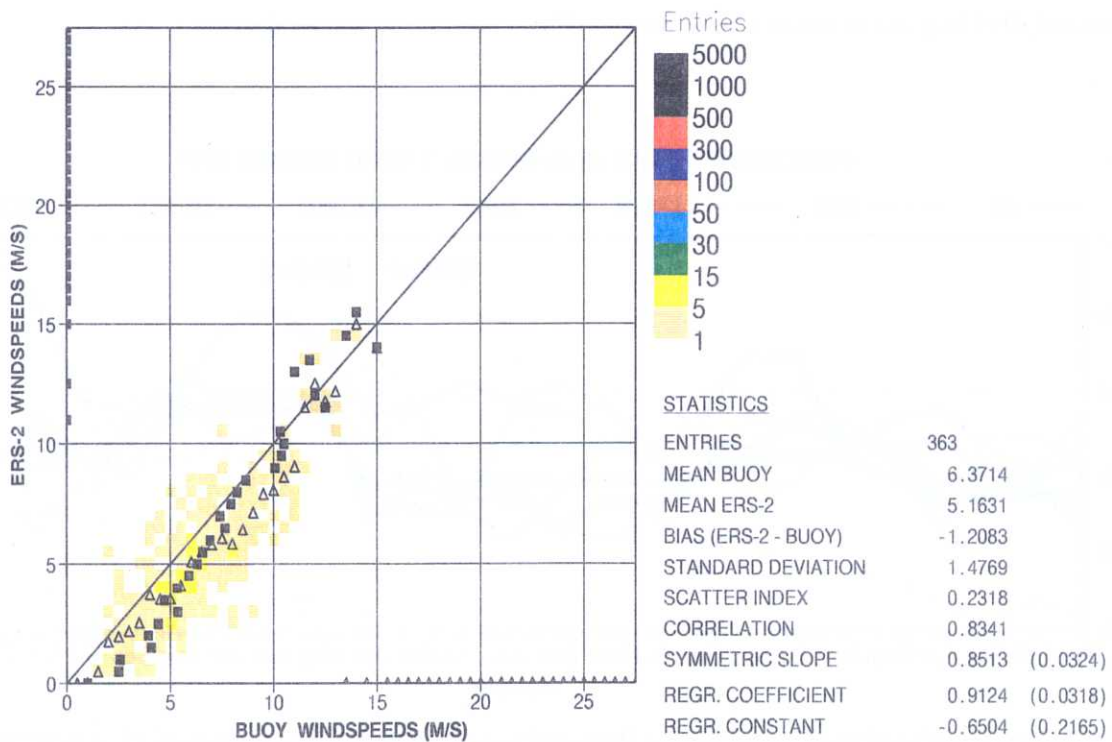


Fig.10 Comparison of Altimeter wind speeds with Northern Hemisphere buoy data for July 1997, confirming the underestimation of winds by the Altimeter in the summertime.

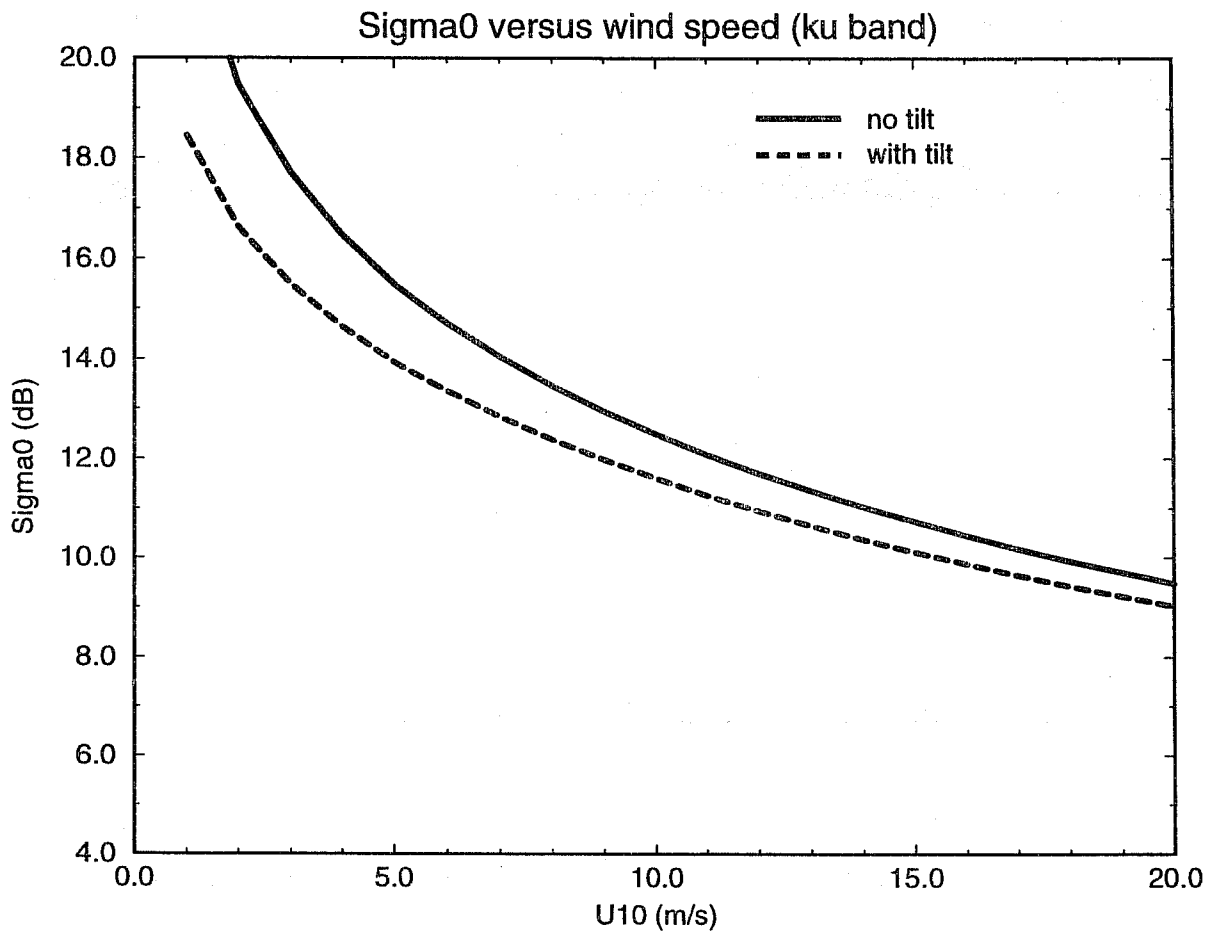


Fig.11 Impact of tilt of long waves on the relation between Radar backscatter and wind speed.

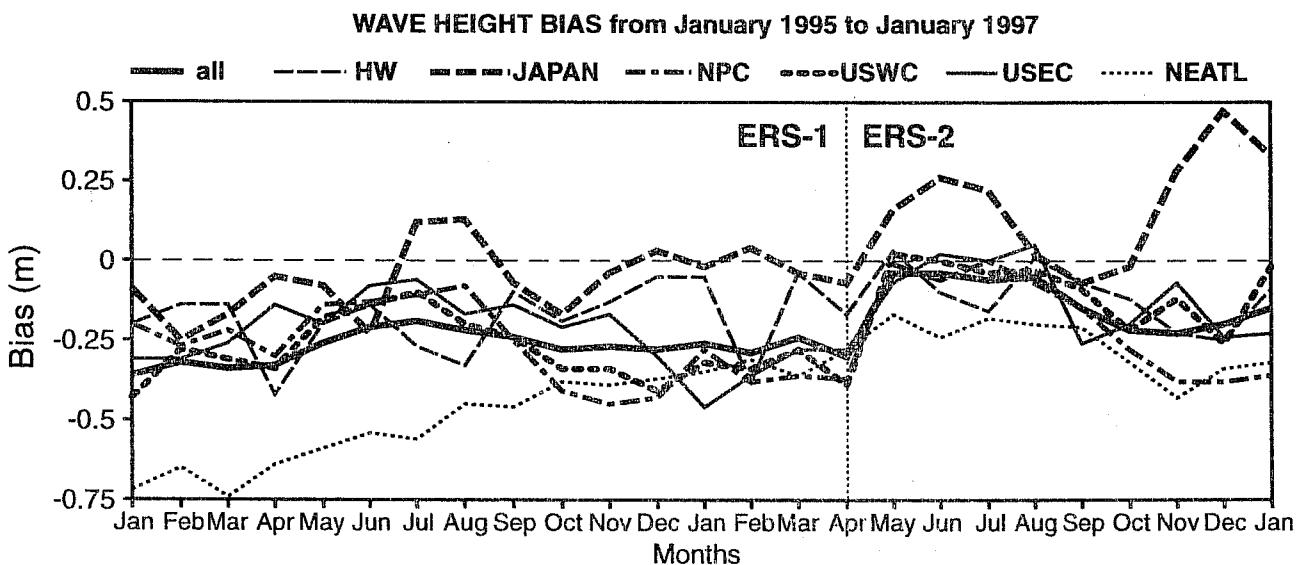


Fig.12 Verification of analyzed wave height against Northern Hemisphere buoys for a number of areas over the period of January 1995 until March 1997. Note the pronounced change in Bias in May 1996, when we switched from ERS-1 to ERS-2 data in our wave analysis.

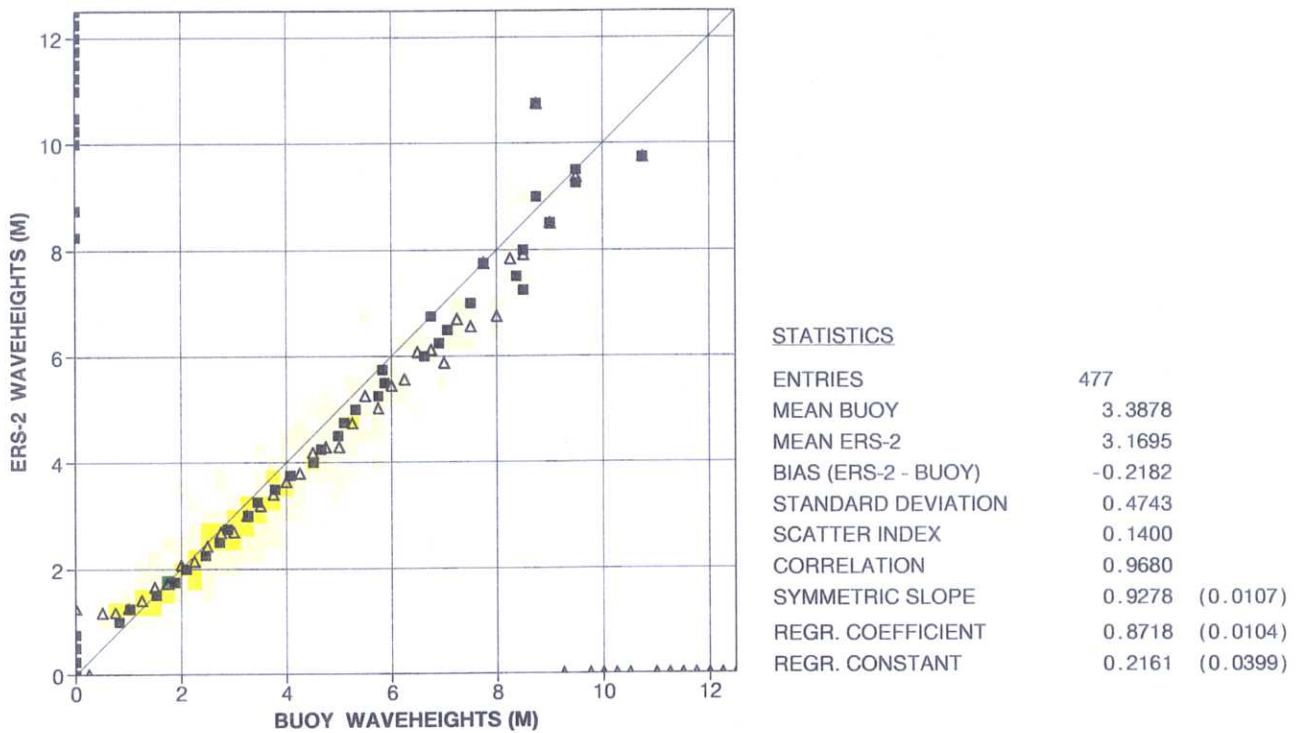


Fig.13 Comparison of ERS-2 wave height against buoy data for February 1997 over a large dynamic range. It shows that the Altimeter underestimates wave height by about 8% in conditions of wind-wave generation.

Sea state dependency of the ERS-2 Radar Altimeter wave height retrieval - February 1997
(bin width = 0.01)

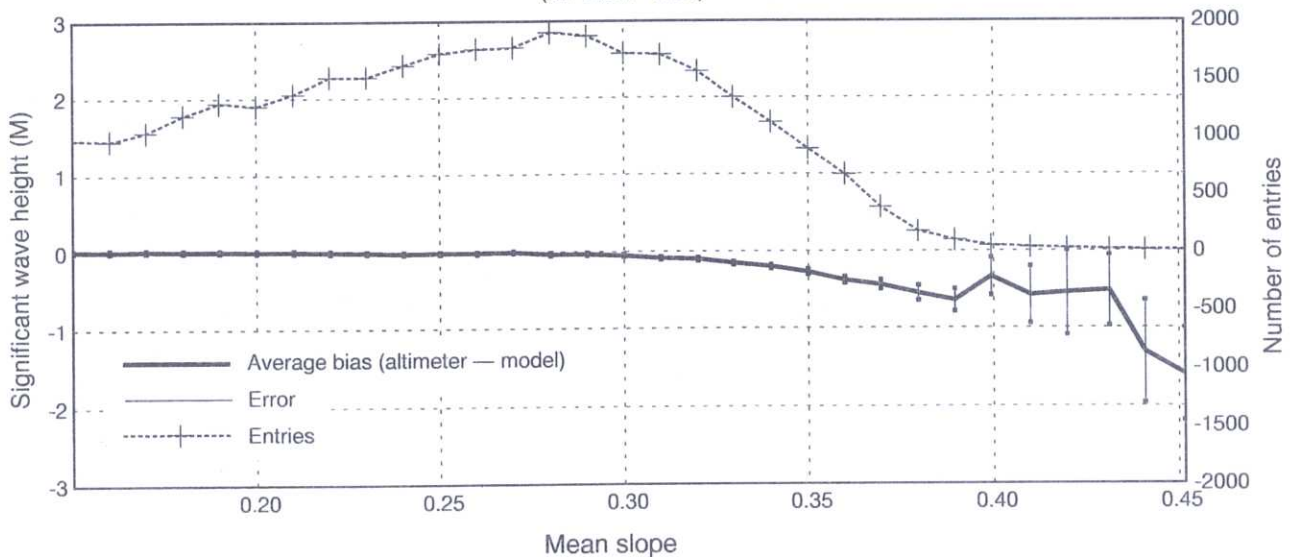


Fig.14 Mean difference of Altimeter wave height and first-guess wave height as function of mean slope, showing under-estimation of wave height by Altimeter for large slopes.

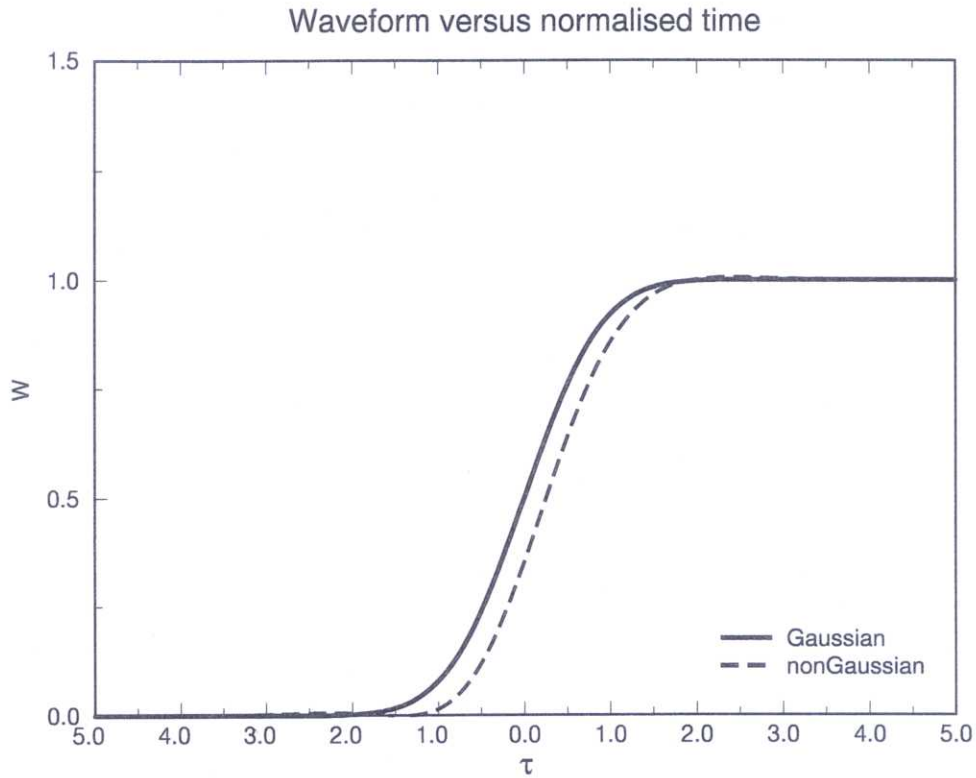


Fig.15 Impact of shape of the joint probability distribution for surface elevation and slope on the form of the return echo as received by the Altimeter.

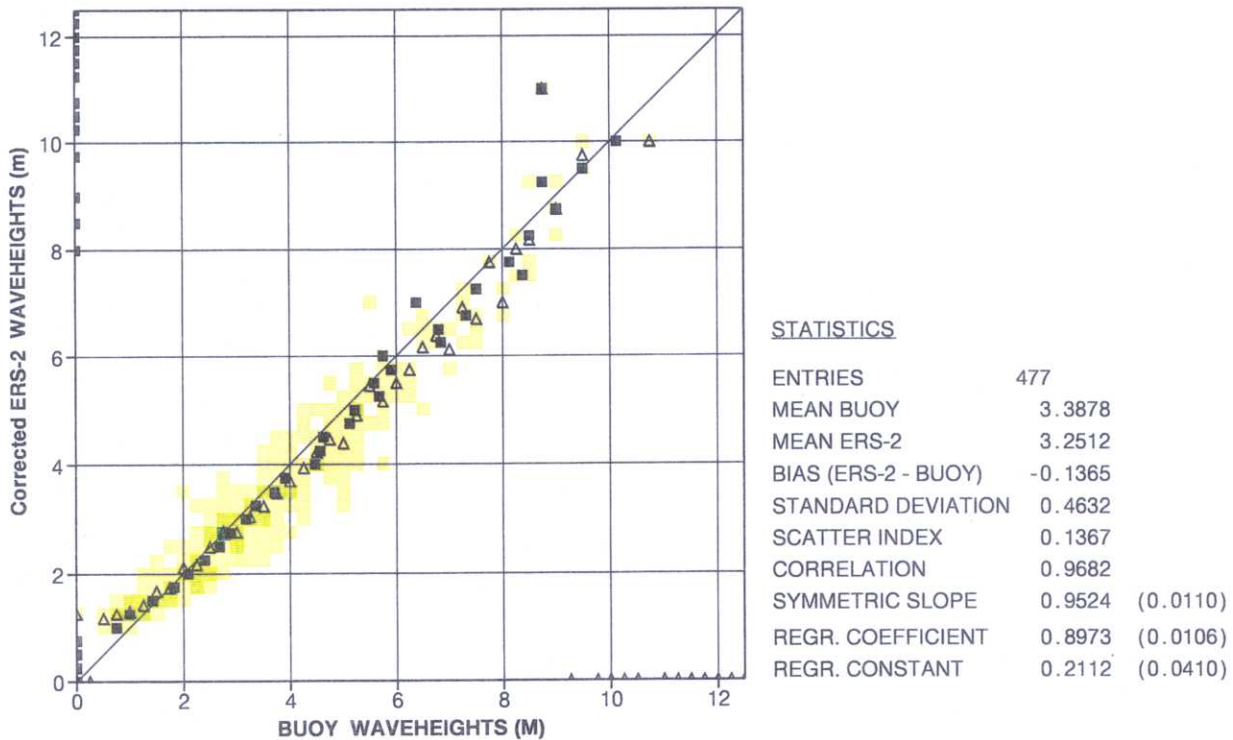


Fig.16 Comparison of corrected Altimeter wave height data with buoy data for February 1997. In order to appreciate the improvements compare with Fig 13.

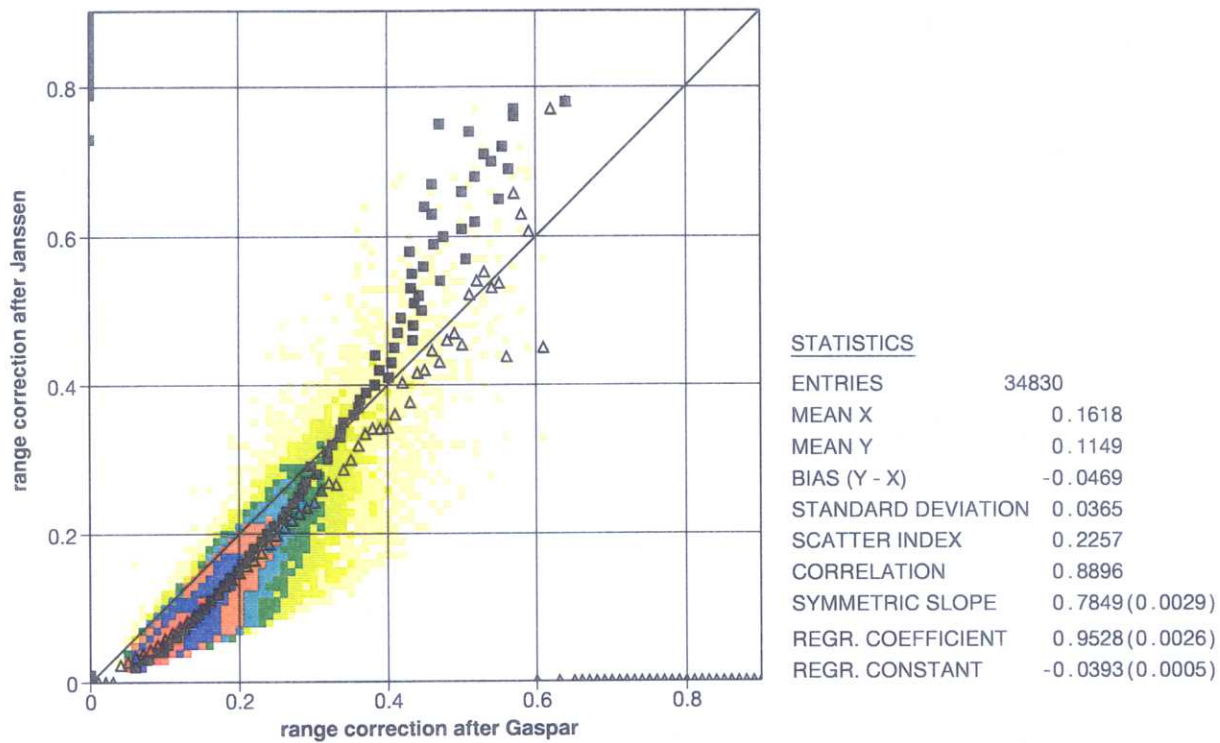


Fig.17 Comparison of electromagnetic bias (EMB) using the approach of Srokosz (1986) and wave model spectra with the empirical correction of Gaspar and Ogor (1996) for the sea state bias (SSB). Area is the globe and period is February 1997.

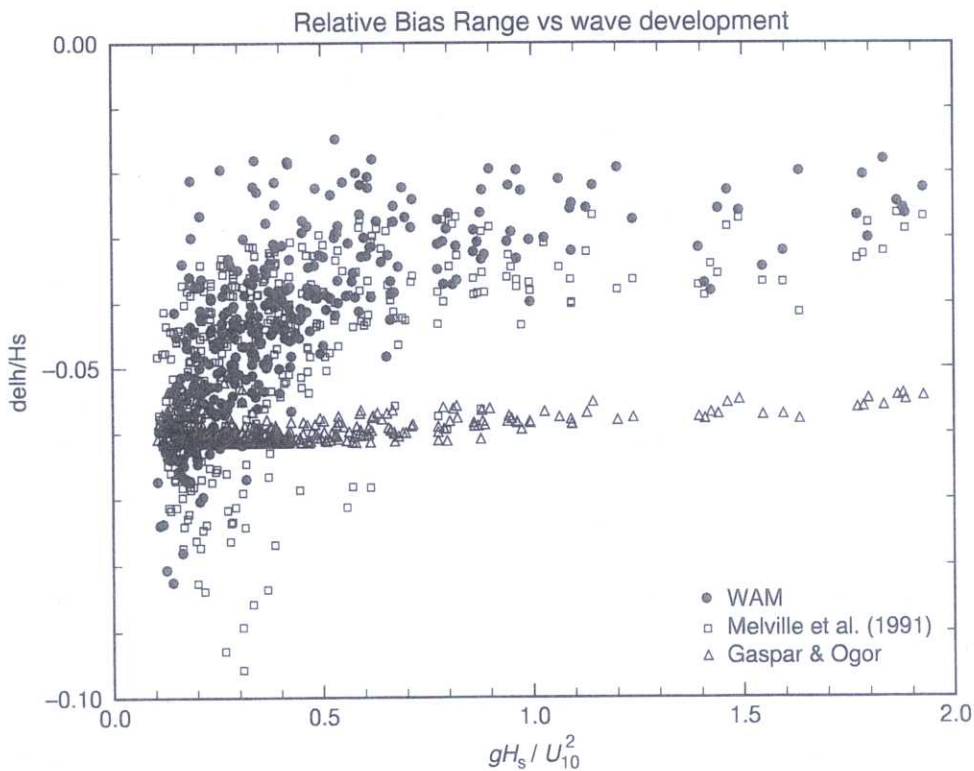


Fig.18 Sea state dependence (as measured by gH_s/U_{10}^2) of the relative Range correction. Dots: present approach; triangles: Gaspar and Ogor (1996); squares: Melville et al (1991).