

15 Years using Altimeter sea state products

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Abstract

The prospect of global observations of surface winds and waves gave a significant stimulus to wave model development in the 1980's, while the need to have reliable wave predictions stimulated the development of operational altimeters that could provide accurate wind and wave products. Over the past fifteen years there has been a continuous interplay between ocean wave forecasting and altimeter sea state products resulting in improvements in both. Altimeter sea state data are presently used in the wave height analysis and wave forecast verification, in the monitoring of the quality of the modelled surface wind and in obtaining a global wave height and wind speed climatology.

1 Introduction

Sea state forecasting started more than sixty years ago when there was an evident need for knowing the wave state during landing operations in the Second World War. The past six decades has seen a development in ocean wave forecasting from simple manual techniques to sophisticated numerical wave models based on physical principles. For a recent review of these developments and the resulting progress in wave forecasting see Janssen (2007). In particular, in the 1980's development was rapid, because of the need to improve wave modeling in rapidly varying circumstances and because of the availability of powerful computers needed to solve the energy balance equation. Furthermore there was the prospect of the availability of wave data from satellites such as Geosat, ERS-1, Topex-Poseidon and ERS-2. In contrast to the conventional observing systems which could only provide 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, a group of mainly European wave modellers started to develop a surface wave model from first principles, i.e. a model that solves the energy balance equation for surface gravity waves. The source functions in the energy balance included an explicit representation of the wind input, nonlinear interactions and dissipation by white capping. 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 into the quality of the WAM model results were based on a comparison 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 ECMWF winds showed good agreement with observed wave heights, 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 Ocean and the Tropical oceans. 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 1980's were too low in the Southern Hemisphere because the atmospheric model had a fairly low resolution (T106, corresponding to a spatial resolution of 190 km).

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. In November 1991 the next version of WAM, named WAM cy4 (Janssen, 1991; Komen *et al.*, 1994) became part of the ECMWF wave prediction system. In addition, in September 1991 the horizontal and vertical resolutions of the ECMWF atmospheric general circulation model were doubled to produce a better representation of surface winds, in particular for the Southern Ocean. Therefore, in late 1991 there was sufficient confidence in the quality of the ECMWF wind-wave forecasting system that it could be used for the validation of ERS-1 altimeter wind and wave products. ESA launched ERS-1 in July 1991 and started dissemination of observed products in near real time soon thereafter. Comparison of ERS altimeter wind and wave products with corresponding ECMWF fields identified problems in the ERS wind speed and wave height retrieval algorithms (Hansen and Günther, 1992; Janssen *et al.*, 1997b), which, because of the strong interaction between the wave model community and ESA, resulted in correction of the operational retrieval algorithms.

Because of the promising validation results altimeter wave height data have been assimilated in the UKMO and ECMWF global wave forecasting systems since mid 1993. The analysis scheme used by ECMWF is based on the Optimum Interpolation Method, details of which are more fully described by Janssen *et al.* (1989) and Lionello *et al.* (1992). In general, the assimilation has led to an improved wave analysis as follows from comparisons with independent observations from buoys. In addition, Bauer and Staabs (1998) compared ECMWF wave analyses with TOPEX/POSEIDON altimeter wave height data and they found an increase of correlation between the two after the ERS-1 wave data assimilation was switched on.

We start this review at the time the ERS-1 satellite was launched. Apart from the good quality observations,

ERS-1 was also the first satellite that provided near real time information on sea state parameters such as significant wave height and wind speed from the altimeter, the wind vector from the scatterometer and 2D wave spectra from SAR. The review will be restricted to the sea state observations from the altimeter. The near real time availability of these sea state parameters is an important factor because, as will be clear from this paper, these data provided a significant stimulus to the improvement of operational ocean wave forecasting, while, on the other hand, the need to have reliable wave predictions stimulated research in the development of altimeter wind and wave products.

The paper will demonstrate this point by discussing some illustrative examples. A brief description of the various satellite instruments is given in section 2, while in section 3 it is pointed out that altimeter wind speed observations are important for monitoring the quality of modelled surface wind, and for wind speed climatology. Nevertheless, the physics behind the retrieval algorithm is not fully understood yet. The wind speed observations follow from an empirical relation between the radar backscatter and wind speed. A well-known example is the 'classical' Chelton-Wentz retrieval algorithm (MCW) of Witter and Chelton (1991) which is based on a few hundred collocations between wind speed observations from buoys and radar backscatter from Geosat. An improved version of this, based on much more collocations, was recently developed by Abdalla (2007). However, the radar backscatter is inversely proportional to the mean square slope and in this paper it is suggested that mean square slope is a physically interesting quantity to observe as it is closely related to air-sea interaction. Alternatively, a number of researchers have made attempts at developing a new retrieval algorithm that includes a measure of the sea state, e.g. the significant wave height (Lefèvre *et al.*, 1994, Gourrion *et al.*, 2002). A comparison of results from the Gourrion algorithm and the Abdalla algorithm will be presented. Finally, the wind speed retrieval algorithms are normally only applied to relatively modest wind speeds up to 20 ms^{-1} , but it is known that the radar backscatter even provides useful information in hurricane conditions (Quilfen *et al.*, 2006).

In section 4 we show that the assimilation of altimeter wave height data is of importance for the quality of the wave analysis and wave forecast, as judged by comparisons with in-situ observations. In addition, it turns out that these data have been of great help in diagnosing problems in the formulation of the physics source functions of the wave model. The wave analysis and forecast are strongly constrained by wind forcing, and to a lesser extent by errors in the wave model. Since over the past 15 years the quality of surface winds has improved quite considerably (cf. Janssen, 2004) and also the wave model has improved (Bidlot *et al.*, 2007) the impact of the assimilation of altimeter wave height data is now less pronounced than in the early days. Furthermore, we show that altimeter wave height data are very useful in obtaining a global wave height climatology. In a first approach one simply averages altimeter wave heights over large enough boxes over the oceans over a large enough period in time. A more promising approach is perhaps from Caires *et al.* (2005) who took ECMWF reanalysis data for wave height and corrected the wave height climatology by means of buoy and TOPEX altimeter wave height data.

In section 5 we will briefly discuss a new method of observing the spectral properties of surface gravity waves, based on the principles of a Real-Aperture Radar (RAR). Compared to the Synthetic Aperture Radar (SAR) the advantage of a RAR is that the spectra are not distorted by velocity bunching effects. Nevertheless, due to speckle noise there is still a limit of detection of the short waves, expected to be around 70 m, which is smaller by a factor of three than the azimuthal cut-off wavelength of a typical SAR. We conclude with a summary of conclusions.

2 Description of the Sensor and Calibration

The radar altimeter is a nadir looking active microwave device instrument. This instrument emits pulses, and by measuring the travel time of the return pulse, after extensive corrections for atmospheric delays, and accurate determination of the satellite orbit, for example, information on the mean sea level may be obtained. To a good approximation the backscattered return, which may be described by specular reflection, is inversely proportional to the mean square slope of the sea surface. As according to Cox and Munk (1954) mean square slope is closely related to the surface wind speed, the radar backscatter is a good measure for wind speed. Finally, the radar altimeter also provides a measure of the significant wave height through the distortion of the mean shape of the return pulse. The earlier return from the wave crests and the retarded return from the wave troughs leads to a deformation of the return pulse which can directly be related to the significant wave height. To determine the mean pulse shape, in the order of one thousand pulses are averaged, yielding one significant wave height measurement about every 7 km along the satellite track. For a Gaussian sea surface, the relation between pulse shape and the mean square slope (rms) sea surface displacement can be determined theoretically (although there are small corrections needed caused by deviations from Normality, cf. Janssen, 2000; Gómez-Enri *et al.*, 2007). This model has been confirmed by numerous comparisons with *in situ* measurements. The typical accuracy of significant wave height of the older generation of radar altimeters is thought to be the maximum of 0.5 m or 10% of wave height in the range of 1 to 20 m, while the windspeed accuracy is between 1.5 and 2 ms^{-1} (for windspeed in the range of 0 to 20 ms^{-1}). As will be discussed in the paper, based on a triple collocation technique, nowadays the accuracy of significant wave height is thought to be 6% of significant wave height, while the error in wind speed is less than 1.25 ms^{-1} .

The radar altimeter is an important component of the payload of a number of satellites such as Seasat, Geosat, ERS-1/2, Topex/Poseidon, GEOSAT Follow On (GFO), ENVISAT and Jason-1. For operational models, the data need to be available in near-real time (i.e. within three hours). ERS-1/2, ENVISAT and Jason-1 provide

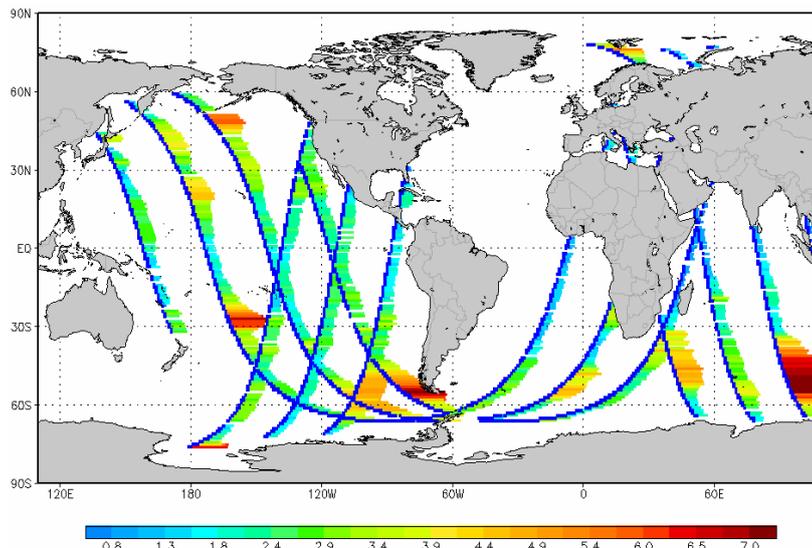


Figure 1: Spatial distribution of significant wave height data from Jason-1 and ENVISAT along tracks during a six hour period. The coloured bars indicate the observed wave height at the location of the tracks.

these fast delivery products. Fig. 1 gives the spatial distribution of significant wave height data from the altimeters on board of Jason and ENVISAT during a typical six hour period, showing that two polar orbiting satellites already give good coverage in such a relatively short time window.

Before we start the discussion on the value of altimeter wind speed and wave height data for numerical wave prediction, it should be noted that, on the other hand, wind and wave model products have been quite useful in the validation and calibration of ESA's altimeter wind and wave product. The necessary calibration and validation of a satellite sensor requires large amounts of ground truth which should cover the full range of possible events. In particular the number of reliable wave in situ measurements is very limited and, because of financial restrictions, dedicated field campaigns are possible only 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 dataset. During the ERS-1/ERS-2 and ENVISAT cal/val campaigns the altimeter-model comparisons have been very effective in identifying errors and problems in the altimeter processing and retrieval algorithms. A few examples are given now.

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 characterization 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 0.5 m when compared with modelled wave height. The second example occurred during the operational phase of ERS-1. A bug was discovered in the processing algorithm which led to unrealistically shaped wave height distributions. This bug was removed at the beginning of 1994 and resulted not only in a much improved shape of the wave height histograms, but also in a reduction of mean wave height of about 30 cm (Bauer and Staabs, 1998).

For altimeter winds a different approach needs to be followed because engineering and geophysical calibration cannot be separated as there is no absolute calibration of the backscatter against independent data from man-made targets or stable known targets of opportunity readily available.¹ For the initial data calibration the system gain as determined by pre-launch instrument characterization 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 set it was shown that the observed antenna gain was well within the error budget for pre-launch calibration. The data calibration was updated in early December 1991 and since that date the quality of the ERS-1 altimeter wind speeds reached an acceptable level.

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 allowing an intercomparison between the products from ERS-1 and ERS-2, 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 backscatter σ_0 from the two satellites the mean difference between the two gave the antenna gain bias. The retuned altimeter wind speeds gave a favourable agreement with the ECMWF surface winds, showing 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 somewhat higher cut-off value than ERS-1. The higher cut-off value is caused by the somewhat different instrumental specifications of the ERS-2 altimeter. Because of

¹Nowadays, there is an ESA effort to calibrate the ENVISAT RA-2 σ_0 using one transponder but it takes a number of years before an accurate estimate of the absolute value of the radar backscatter is found. Once the absolute value is known, all Altimeter missions can be corrected to provide an absolute σ_0 , provided there is sufficient overlap in time between the different missions.

the tandem mission it was possible to compare ERS-1 and ERS-2 wave heights using the wave 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.*, 1997b). 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.*, 1997b) 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 waveform, resulting in a better estimate of the slope at the half-power point and hence of wave height (R. Francis, private communication, 1997).

A similar procedure was adopted for the cal/val of ENVISAT, but now there was a tandem mission with ERS-2. ENVISAT carries a two-frequency altimeter, one frequency is at K-band (as in case of ERS1/2) and the other is at S-band. The Ku-band wave heights are found to be of the highest quality, mainly because of a higher sampling of the waveform. Regarding wind speed, the antenna gain bias was obtained in a similar fashion as in the case of ERS-2, and by comparing with analyzed wind speed the altimeter wind speed based on recalibrated σ_0 was found to be of high quality.

In the next two sections we shall describe the benefits altimeter data have for operational wave analysis and forecasting systems and how in turn the quality of these observations has been improved by research developments and by comparison with wind and wave model products and in-situ observations. We restrict most of the discussion to altimeter products which are available in near-real time (within 3 hours) after the remote sensed observations have been made. The reason for this strict requirement on availability is the need of operational centres to produce a wave analysis as soon as possible. These products are the so-called Fast Delivery Records (FDR) of the ERS-1/2, ENVISAT and Jason-1 satellites. A few weeks after the time of observation final Ocean Products (OPR) are produced by the relevant processing facilities. These OPR products are usually of higher quality. The exception is ENVISAT since the FDR sea state products have already the required accuracy and the OPR wave heights and wind speeds are identical to their FDR counterparts.

3 Altimeter wind speed data

3.1 Benefits of altimeter wind speed observations

While there is a considerable scatter in the relation between the radar backscatter σ_0 and the wind speed U_{10} , reflecting that perhaps the physics behind the altimeter wind speed retrieval is not fully understood, it should be pointed out that the altimeter wind speed product has nevertheless proven its value. Three examples are mentioned. The first example concerns the validation of a new method of weather analysis namely the variational method, which replaced the Optimum Interpolation (OI) scheme in the ECMWF atmospheric forecasting system. 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 forecast system (Andersson *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. This follows from a verification of analysed wind speed in the Southern Hemisphere against altimeter wind speeds from ERS-1 altimeter (for details see Janssen, 1999), where the OI scheme without scatterometer data gave a standard deviation of error in analyzed surface wind of 1.99 ms^{-1} while the combination of 3DVAR and scatterometer gave an error in surface wind of only 1.77 ms^{-1} . The reduction in standard deviation of 10% may be regarded as quite considerable. This shows at once the 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.

Indeed, the altimeter winds have been used in assessing operational model performance and changes and the introduction of new data types such as surface winds from SSMI and from ASCAT on board of Metop. As a second example, in Fig. 2 a timeseries of the standard deviation of error of analyzed wind speed against ERS-2

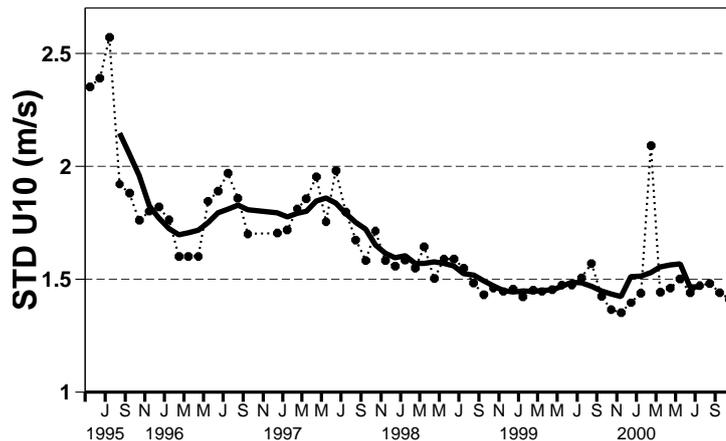
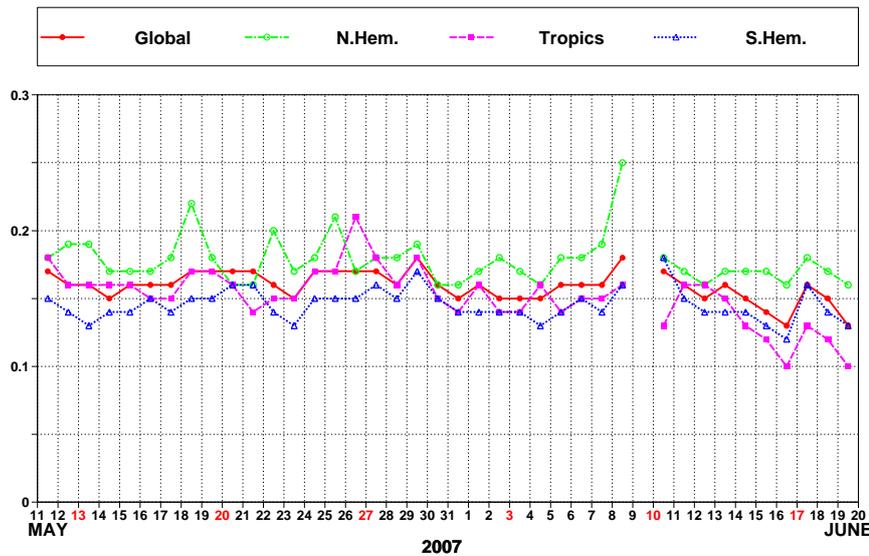


Figure 2: Standard deviation (SD) of analyzed wind speed as obtained from the comparison between analyzed surface wind field and ERS-2 altimeter windspeed over the period May 1995 to October 2000. Area is the whole globe. The filled circles are monthly data while the continuous line is a 6-month running average.

altimeter windspeed over the period May 1995 to October 2000 is presented. The area is the whole globe. It is clear that over this period ECMWF has made considerable improvements in the quality of the surface winds. Inspecting this timeseries in more detail it appears that at several instants fairly sudden jumps in the accuracy of the analyzed surface wind speed may be found. These can be related to changes in the operational weather forecasting system. For example the transition around February 1996 is connected to the introduction of 3DVAR and ERS-2 scatterometer data, the jump in December 1997 is connected to the dynamic version of the variational analysis scheme, called 4DVAR while the transition in July 1998 is connected to the two-way interaction of wind and waves. A more detailed discussion is given in Janssen (2004). Note that at the end of the period the quality of the altimeter wind speed worsened because of problems with the gyros on board of ERS-2; these gyros help to determine the orientation of the satellite and the radar backscatter σ_0 is sensitive to errors in the zenith angle. This is illustrated in Fig. 2 by the significant outlier in February 2000.

As a third example, the operational introduction of ASCAT surface winds is briefly discussed. The ASCAT instrument is of similar design as the AMI scatterometers on board of ERS-1/2 from which data has been operationally assimilated at ECMWF since 30 January 1996. Triplets of radar backscatter from three antennas are combined to estimate surface vector winds over the global oceans. Both AMI and ASCAT operate at C-band (5.3 GHz) and have the same antenna geometry. Two main differences are a different range of incidence angle (optimized for ASCAT to enhance performance in wind direction) and the fact that ASCAT carries two sets of antennas (providing two swaths that double the coverage). ASCAT data has been monitored at ECMWF from the start of dissemination by EUMETSAT via the EUMet-Cast system on the 31th of January 2007. Surface winds are inverted from available (level 1b) backscatter triplets on the basis of a modified version of the geophysical model function CMOD5 (Hersbach *et al.*, 2007). Resulting winds are collocated with operational short-range ECMWF forecast winds. The monitoring confirms that the ASCAT instrument is working well.



ENVISAT Radar Altimeter wind Wind Speeds: Timeseries of scatter Index (SI) (analysis)

Figure 3: Timeseries of scatter index (SI) of the difference between ENVISAT altimeter and analyzed ECMWF wind speed. Note that on the 12th of June 2007 the assimilation of ASCAT data was introduced in operations giving a closer match between altimeter and analyzed wind.

Besides a few short periods of data interruption, data volume is constant, and quality is found to be high and stable. Main concern is the availability of currently only one transponder out of a set of three, which prevents an absolute calibration of the level 1B product. Although the calibration of the ASCAT instrument is still preliminary and non-optimal according to several groups that have worked with the data so far, it appears very stable. Instrument noise is low. Appropriate bias corrections, therefore, allow for the retrieval of a high-quality wind product. Especially the quality of ASCAT wind direction is found to be excellent, outperforming scatterometer data from ERS-2 for example. Assimilation experiments with the ASCAT data confirmed the high quality of the product as there were small but significant improvements in forecast performance, in particular in the Southern Hemisphere scores of 1000 and 500 hPa geopotential. On the 12th of June 2007 ECMWF was the first to present these data to the analysis scheme. Improvements of the analyzed surface winds were already evident after a few days as is shown in Fig. 3. Here, for different areas on the globe, timeseries of normalized standard deviation of error (SI) of the difference between ENVISAT altimeter wind speed and analyzed windspeed are shown for the period of the 11th of May until the 20th of June, 2007. The assimilation of ASCAT data commenced on the 12th of June and the impact of these new observations on the quality of the surface analysis is clearly visible after a few days.

3.2 Altimeter wind speed data and problems

Nevertheless from long term monitoring of the quality of the altimeter winds it was found that in particular in the Northern Hemisphere there is a clear underestimation of wind speed by the altimeter during the late spring and early summer, whereas in the remainder of the year the bias is much smaller (Janssen, 1999). This problem is illustrated in Fig. 4 which shows over the period of December 1996 until January 2000 a pronounced seasonal cycle in the bias between ERS-2 altimeter wind speed and analyzed wind speed for the Northern Hemisphere,

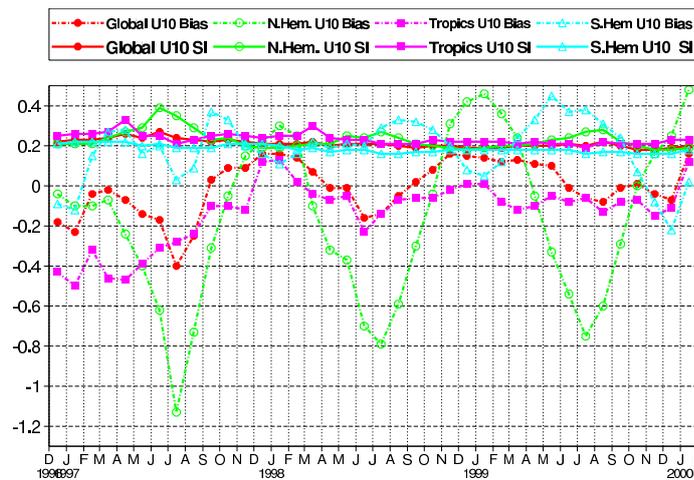


Figure 4: Comparison of ERS-2 Altimeter wind speeds with analyzed model winds for the Globe, the Northern Hemisphere, the Tropics and the Southern Hemisphere. Shown are monthly values of bias (alt-mod) and scatter index over the period of December 1996 until January 2000. Note the clear seasonal cycle in the bias for the Northern Hemisphere.

whilst no such cycle is seen in the Tropics and in the Southern Hemisphere.² One may argue that this is perhaps a problem with the model winds in the Northern Hemisphere, but a validation of Altimeter wind speed against wind speeds from buoys in the Northern Hemisphere gives a similar picture (Janssen, 1999). This is seen in Fig. 5 which shows for July 1997 (which has according to Fig. 4 the largest negative bias) a scatter diagram between the two. From the scatter diagram it is immediately evident that the underestimation of wind speed by the altimeter mainly occurs in the wind speed range below 10 ms^{-1} . With a mean buoy wind of 6.4 ms^{-1} a fairly large negative bias in the altimeter wind speed of -1.3 ms^{-1} is found. This bias is a robust feature in the summertime. On average one finds for June, July and August 1997 a negative bias of -1 ms^{-1} , whereas by contrast in the Winter of 1996-1997 the bias is only -0.06 ms^{-1} . Similar findings were reported by Chen *et al.* (2000) who compared Topex wind speeds with 6 years of buoy data from the Japan Meteorological Agency. These authors pointed out that the seasonal cycle in the altimeter bias depends on the choice of the retrieval algorithm. The MCW algorithm (which is the one used for the ERS-2 wind retrieval) gave a very similar seasonal cycle to the one displayed in Fig. 4, but the Hwang *et al.* (1998) algorithm gave a seasonal cycle which is just the opposite of the seasonal cycle obtained with the MCW scheme.

This prompted Abdalla (2007) to refine the MCW algorithm by doing a tuning exercise based on much more data than used by Witter and Chelton (1991), namely two months of analyzed ECMWF wind fields and the buoy data for the same period. In particular, attention was paid to the low wind speed regime as in that range the largest discrepancies between altimeter winds and buoy data exists (cf. Fig. 5). The resulting algorithm may be regarded an improvement as follows from a validation of the Abdalla algorithm applied to ENVISAT radar backscatter against buoy data over a one year period, giving a better fit in the low wind speed range, a reduction in standard deviation of error of 5% and a reduction in bias from -0.59 ms^{-1} to -0.13 ms^{-1} . A similar

²The wind speed climatologies in different areas of the globe are really different, in particular regarding the seasonal cycle in the mean wind speed. Areas such as the Southern Hemisphere and Tropical Oceans show hardly any seasonal variation around the annual mean wind speed (the amplitude of the seasonal cycle is less than 0.5 ms^{-1}) while in the Northern Hemisphere the amplitude of the seasonal cycle is much larger, about 1.5 ms^{-1} . The largest seasonal variations are found in the North Atlantic with an amplitude in the seasonal cycle of about 3 ms^{-1} . A similar remark applies to the seasonal variations in mean significant wave height. Therefore it should not come as a big surprise that there are only seasonal variations in the Northern Hemisphere.

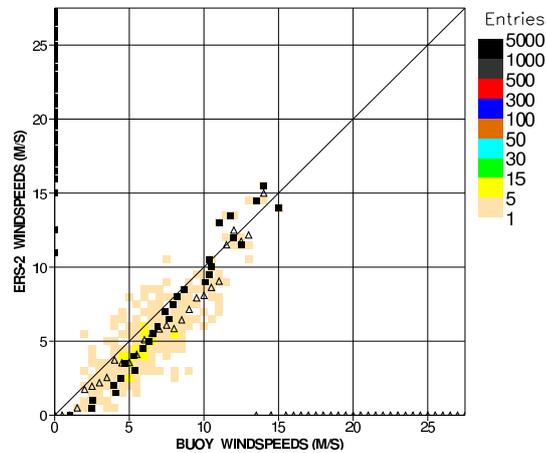


Figure 5: Comparison of Altimeter wind speeds with Northern Hemisphere buoy data for July 1997, confirming the underestimation of winds by the Altimeter in the summertime, in particular in the low wind speed range.

reduction in standard deviation of error was found when comparing the newly obtained altimeter winds against analyzed wind speed over a two year period from April 2003 until April 2005. There was also a reduction in the amplitude of the seasonal cycle in the bias, although the decrease was smaller than hoped for. For ENVISAT, using the MCW algorithm, the amplitude in the seasonal cycle was found to be about 0.5 ms^{-1} and with the Abdalla algorithm an amplitude of 0.4 ms^{-1} was obtained. This new wind retrieval algorithm was introduced for ENVISAT in October 2005.

Before possible reasons for a seasonal cycle in the Northern Hemisphere altimeter bias are discussed, it is worthwhile to point out that other satellite instrument may have a similar problem. An example is the ERS-2 scatterometer for the North Atlantic area over the period January 2004 until June 2006 (Abdalla and Hersbach, 2006) which has a bias which depends on the season. The amplitude of the seasonal cycle is however, somewhat smaller, about 0.3 ms^{-1} .

The following discussion will be in the context of the assumption that the radar backscatter σ_0 is caused by specular reflection from the short ocean waves, i.e. the amount of backscatter is related to the state of the short waves. We will discuss a number of additional effects which, so far, have not been considered in the standard backscatter model for operational wind speed retrieval (an exception is sea state effects for Jason-1), and we indicate how the additional physics may contribute to a better understanding of the seasonal cycle in the altimeter wind speed bias.

Effect of slicks

One plausible reason for the seasonal cycle in the altimeter bias may be related to the presence of slicks, as it is known that slicks of natural origin play an important role in the damping of high frequency ocean waves (Alpers and Hühnerfuss, 1982), and from Janssen *et al.* (1998) it is known that the effect of slicks is important in determining the radar backscatter. 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 $7\text{-}10\text{ ms}^{-1}$ this underestimation of the strength of the wind should be most pronounced in the low windspeed range which agrees with the findings from the validation of altimeter winds against buoy winds (cf. Fig. 5) and also against analyzed 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 as function of the time in the year. Information on this was obtained from monthly mean observations of ocean colour by SeaWiFs, which are displayed as global maps on the Web. Since ocean colour is a measure of biological activity in the upper layer of the ocean and since 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 confined to coastal areas (for example, near Japan), thus when the presence of slicks would be relevant for the bias problem in altimeter winds one should expect that the main problem occurs in the North Atlantic in May 1998. This inference is, however, not supported by the comparison of altimeter winds against analyzed ECMWF winds. According to the present verification of altimeter winds, the main bias problem seems to occur in June, July and August when there is, compared to May, relatively little biological activity. Also, when studying the verification statistics for the North Pacific and the 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.

Neutral versus 10 m Winds

Another interesting deviation from the standard backscatter model is related to the physics of how short waves are generated by wind. Since the contribution by Miles (1957) there is a growing belief that ocean waves are forced by the surface stress and not by the surface wind. As a consequence, instruments such as the altimeter and the scatterometer observe the surface stress. Rather than correlating radar backscatter σ_0 with the wind speed at 10 metre height, U_{10} , it may be more appropriate to correlate σ_0 with the equivalent neutral wind U_{10N} , because this measure of surface wind is more closely connected to the surface stress. This approach has been followed, for example, in the development of the QuikSCAT geophysical model function, and at ECMWF the wave prediction model is forced by neutral winds.

The use of neutral winds, rather than 10 m winds, is in particular relevant for wind speeds up to 10 ms^{-1} and, therefore, may have some impact on the seasonal variations of the Altimeter bias. By inspecting a number of ECMWF surface wind fields in the winter of 2006 and the summer of 2006 one may obtain information such as the fraction of unstable and stable cases and the systematic difference between U_{10} and U_{10N} (For a similar exercise on the winter of 2001 see Abdalla and Hersbach (2006) and Brown *et al.* (2006)). We restrict our attention to the Northern Hemisphere oceans. One then finds that in the winter time the fraction of unstable cases on average is 79 % and, therefore the fraction of stable cases is 21 %, while the average difference between U_{10} and U_{10N} is -0.24 ms^{-1} . In the summertime there are more stable cases as the fraction of stable cases increases to 42 %, and consequently the fraction of unstable cases is on average 58 %, while the difference between 10m wind and neutral wind is almost zero, about -0.02 ms^{-1} . Hence, interpreting the radar backscatter in terms of the neutral wind rather than the 10 m wind speed will result in a reduction of the amplitude of the seasonal cycle of about $(0.24 - 0.02)/2 = 0.11\text{ ms}^{-1}$. Therefore, although theoretically it is expected that it is more appropriate to relate σ_0 with the neutral wind speed, it is not expected that the use of neutral winds will explain a large part of the seasonal cycle in the altimeter bias.

Sea-state effects

Although there is a reasonable correlation between σ_0 and surface wind U_{10} researchers have often wondered whether there are not additional sea state parameters relevant in the relation between backscatter and wind speed. And for good reasons. For a nadir-looking radar the main scattering mechanism is specular reflection. Therefore the backscatter is proportional to the joint probability density of slopes $p(\eta_x, \eta_y)$ of the surface η , where η_x and η_y are the slope components in two orthogonal directions. The radar backscatter $\sigma_0(\theta)$ at an incidence angle θ is then given by the classical result (Barrick 1968; Valenzuela 1978)

$$\sigma_0(\theta) = \frac{\pi |R(0)|^2}{\cos^4 \theta} p(\eta_x, \eta_y)|_{spec}, \tag{1}$$

where $R(0)$ is the Fresnel reflection coefficient for normal incidence and the probability density of slopes $p(\eta_x, \eta_y)$ is evaluated at the specular points. Accordingly, only surface facets normal to the direction of the incident radiation contribute to the backscattering. For small slopes the pdf of the surface slope is given by a Gaussian distribution (Cox and Munk, 1954)

$$p(\eta_x, \eta_y) = \frac{1}{2\pi\sigma_u\sigma_c} \exp \left[-\frac{1}{2} \left(\frac{\eta_x^2}{\sigma_u^2} + \frac{\eta_y^2}{\sigma_c^2} \right) \right], \tag{2}$$

where σ_u^2 and σ_c^2 are the slope variances in the along and cross direction. Combining (2) and (1) one finds that for normal incidence ($\theta = 0$)

$$\sigma_0(0) = \frac{|R(0)|^2}{2\sigma_u\sigma_c}, \tag{3}$$

which for an isotropic surface, with $\sigma_u^2 = \sigma_c^2 = s^2/2$ where s^2 is the total slope variance, becomes

$$\sigma_0(0) = \frac{|R(0)|^2}{s^2}. \tag{4}$$

In order to apply (4) it should be realized that only a portion of the total mean square slopes of the ocean surface is included in s^2 , namely only those ocean waves contribute whose wavenumber is smaller than the wavenumber k_R of the electro-magnetic radiation. With $F(k, \phi)$ the surface wave spectrum, where k is the wavenumber and ϕ the wave propagation direction, the slope variance s^2 therefore becomes

$$s^2 = \int_0^{k_R} \int_0^{2\pi} k dk d\phi k^2 F(k, \phi). \tag{5}$$

Eq. (4) clearly shows how the radar backscatter return depends on the sea state through the mean square slope s . Traditionally, however, the radar backscatter has been interpreted in terms of the surface wind speed, as Cox and Munk (1954) showed that there is a correlation between mean square slope and surface wind. Nevertheless, the relation between these two parameters is certainly not perfect. Several studies (Monaldo and Dobson, 1989; Glazman and Pilorz, 1990; Glazman and Greysukh, 1993; Lefèvre *et al.*, 1994, Hwang *et al.*, 1998 and Gourrion *et al.*, 2002) suggest that in the presence of swell the radar backscatter may depend on both the local windspeed and the sea state. Therefore, a number of researchers have made attempts at developing a retrieval algorithm that includes a measure of the sea state. Most authors choose the significant wave height as measure of sea state, because this parameter is readily available from the radar altimeter. Gourrion *et al.* (2002) have performed the most extensive study in this direction and one of their main results is displayed in Fig. 6, which shows the relation between Topex backscatter σ_0 and surface wind speed U_{10} from the NSCAT scatterometer whilst the significant wave height H_S (as obtained from Topex) is indicated by colour coding. Fig. 6 clearly shows that in

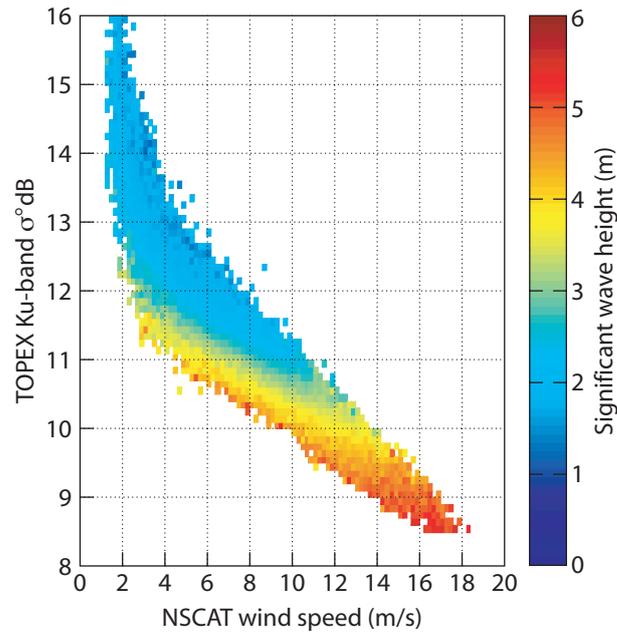


Figure 6: The relation between Radar backscatter σ_0 and surface wind speed U_{10} obtained from collocations between σ_0 from Topex and U_{10} from the NSCAT scatterometer. Colour coding is according to significant wave height observed by Topex. From Gourrion *et al* (2002).

particular for low wind speeds there is a dependence of the radar backscatter on significant wave height, and a two-parameter model of the type

$$U_{10} = f(\sigma_0, H_s) \quad (6)$$

and its inverse was developed using a neural network approach. Gourrion *et al* (2002) presented clear evidence that, compared to the one-parameter models such as the MCW retrieval algorithm, the sea-state dependent algorithm (called from now on the Gourrion algorithm) performed better. Consequently, the Jason-1 altimeter wind speed observations are obtained using this retrieval algorithm. The good performance of the Gourrion algorithm is reflected by the validation efforts at ECMWF where routinely the Jason-1 wind speed are compared with analyzed ECMWF wind speed. An example of this verification is shown in Fig. 7 which shows time series of the wind speed scatter index between Jason-1 and ECMWF analyzed wind speed over the period of November 2003 until May 2005 for the global oceans. The typical value for the scatter index is slightly more than 16% which, with a mean observed wind speed of 8 ms^{-1} , corresponds to a standard deviation of error of 1.25 ms^{-1} .

Nevertheless, one may wonder to what extent the sea state information has resulted in a better wind speed retrieval. For this reason Abdalla (2007) improved the pure wind speed algorithm of Chelton and Wentz by using a much larger data set (consisting of both ECMWF winds and in-situ observations) for fitting the unknown coefficients and this algorithm was introduced for ENVISAT in 2005. Global monitoring the Jason-1 Operational Sensor Data Record products at ECMWF suggests that the Jason Ku-band σ_0 is about 0.4 dB higher than that of ENVISAT. Accordingly, the Jason-1 σ_0 values were reduced by this amount before applying the Abdalla algorithm. Results for the scatter index are displayed in Fig. 7 as well and it is clear that this algorithm is performing somewhat better than the Gourrion algorithm. A reason for this perhaps counter-intuitive finding may be that the comparison of the results of the Gourrion algorithm with alternative algorithms was not quite fair because the older algorithms were trained on a much smaller data set than the sea state dependent algorithm.

Finally, is there a seasonal cycle in the Northern Hemisphere Jason altimeter wind speed bias? Returning to Fig.

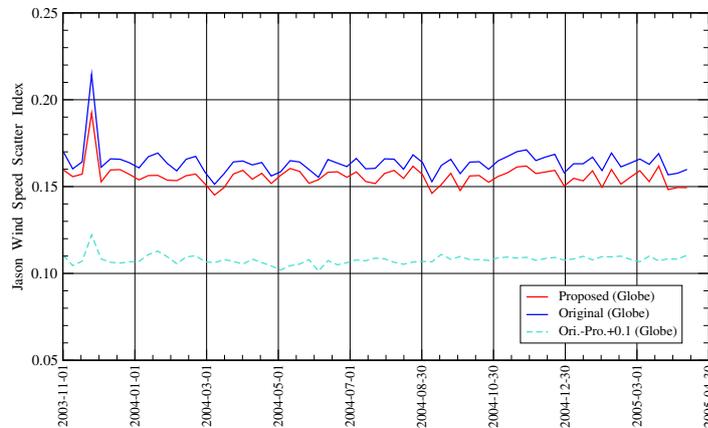


Figure 7: Timeseries of the wind speed scatter index between Jason RA and ECMWF analyzed wind speed from the Gourrion algorithm (blue) and the algorithm proposed by Abdalla (2007) (red) over the period from the 1st of November 2003 until the 9th of April 2005 for the Global oceans.

6 one would expect that a sea state dependent retrieval algorithm would reduce the amplitude in the seasonal cycle. In order to see this consider the low wind speed range from, say 2 to 8 ms^{-1} . For given σ_0 one would have in the Northern Hemisphere summertime a larger retrieved wind speed compared to the wintertime because on average the significant wave height is smaller in the summertime, thus reducing the amplitude in the seasonal cycle. However, from the long-term monitoring of the Jason-1 windspeed against analyzed winds there is clear evidence of a seasonal cycle in the wind speed bias with an amplitude of around 0.35 ms^{-1} which is very similar to the amplitude found with the Abdalla algorithm. This is a disappointing result but it is expected that a more careful tuning of the sea state dependent retrieval algorithm may resolve the problem of the seasonal cycle in the wind speed bias.

3.3 Recent developments

Backscatter versus mean square slope

It is concluded from the previous discussion that the addition of sea state information to the retrieval of wind speed from the radar backscatter gives only a very limited benefit. There have also been suggestions and attempts (cf. Hwang *et al.*, 1998; Gommenginger *et al.*, 2003) to relate radar backscatter and mean wave period. However, this approach might only work for windseas as explained already by Hwang *et al.* (1998). This was confirmed by the validation work of Caires *et al.* (2005) who found that only in 35% of the cases a meaningful retrieval of mean wave period could be obtained. The selection of these windsea cases was based on a criterion obtained from a wave prediction model. Nevertheless, following a neural network approach, Quilfen *et al.* (2005) generated an empirical algorithm that gave a much better agreement with buoy data than the Gommenginger algorithm. This new algorithm should perform better for swell cases.

On the other hand, it may perhaps be a good idea to follow the suggestion of the theoretical approach of the past (which is briefly summarized in Eqns. (1-5)). Clearly, this approach suggests that the radar backscatter σ_0

is connected to the mean square slope s^2 of the sea surface. The mean square slope is in itself a useful quantity to monitor as it measures the roughness of the sea surface which is closely connected to the air-sea momentum transfer. Therefore, let us assume for the moment that the radar backscatter simply provides the mean square slope, an assumption which is supported by many studies in the past, see for example Valenzuela (1978), Apel

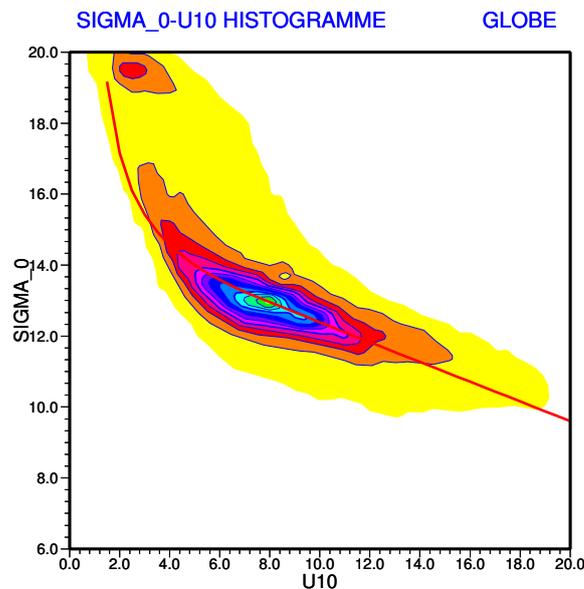


Figure 8: Two dimensional histogram showing the relation between radar backscatter σ_0 and surface wind speed U_{10} obtained from a model of radar backscatter based on specular reflection. Here the slope is obtained from the WAM-VIERS model. As a reference, the σ_0 - U_{10} relation according to Abdalla is shown as the red line.

(1994) and Freilich and Vanhoff (2003).

A problem with the mean square slope is, however, that this quantity is very difficult to model. This follows immediately from the definition of the slope as given in Eq. (5) and the empirical knowledge that the short gravity waves follow the Phillips' wavenumber spectrum which is proportional to k^{-4} . As a consequence, the integral in Eq. (5) has a logarithmic divergence for large wavenumbers, which implies that the value of the mean square slope depends on detailed knowledge of the shape of the gravity-capillary wave spectrum. Fortunately, in the past 15 years we have seen considerable progress in our knowledge of the gravity-capillary wave spectrum, starting with the experimental investigations of Jähne and Riemer (1990) and Klinke and Jähne (1992), followed by semi-empirical model efforts of Apel (1994), Elfouhaily *et al.* (1997) and Liu *et al.* (2000). There is even a model based on the solution of a parametrization of the energy balance equation for short gravity-capillary waves, called the Viers 1 model (cf. Janssen *et al.*, 1998), which was applied with some success to retrieve wind vectors from the scatterometer.

In Fig. 8 is shown results of a simulation of the nadir radar backscatter as function of surface wind speed based on the isotropic model (4-5) for the cross section at normal incidence. The low-wavenumber part of the slope variance s^2 is obtained from two-dimensional operational wave spectra produced by the WAM model at ECMWF while the high-wavenumber part of the slope variance is obtained from the Viers-1 gravity-capillary model. This model basically describes the balance between wind input (including the effects of feedback

of waves on the wind), three-wave interactions, dissipation by white capping and molecular viscosity. The boundary condition for the energy flux at low wavenumbers is determined by the energy flux provided by the resolved waves in the WAM model. In order to check the realism of the short wave model, the mean square slope calculation was applied to the case of visible light and a good agreement was found with the mean square slope observations of Cox and Munk (1954).

Note that as already discussed in the section 2 on the sensor description, an absolute calibration of σ_0 is not readily available, but in a comparison with results from a simulation of the radar backscatter it is highly desirable to have an absolute observed estimate. A dedicated field campaign is under way for the ENVISAT RA-2 altimeter using a ground transponder developed at the European Space Research and Technology Center (ESTEC). The absolute radar cross-section accuracy is expected to be in the region of ± 0.2 dB. The results of this campaign are not yet fully available, but initial results (Pierre Féménias, private communication March 2005) seem to indicate that the *actual* ENVISAT RA-2 altimeter backscatter has a negative bias of 1 ± 0.2 dB.³ Correcting for this bias the global average radar backscatter from ENVISAT is 13.34 dB. Accordingly, in the Abdalla algorithm shown in Fig. 8 appropriate corrections have been applied to agree with this global average of the absolute backscatter.

The plot in Fig. 8 is a histogram as function of σ_0 and surface wind U_{10} . In order to obtain this histogram mean square slopes for the Ku-band case ($k_R = 285$ rad/m) are obtained from the 2D-spectra at a given synoptic time (28 February 2006 at 00Z) with a spatial resolution of 0.5 degrees (giving a total number of cases of around 120,000), and the radar backscatter then follows from Eq. (4). Here, by trial and error it was found that in order to obtain optimal results the nadir reflection coefficient $R(0)$ must have the value $R(0) = 0.82$ which is about 5% higher than obtained from recent estimates of the complex reflectivity of the ocean at 14 GHz using realistic salinity and sea surface temperature (Apel, 1994). By counting the number of cases in a bin of size (0.5 ms^{-1} , 0.5 dB) around a particular U_{10} , σ_0 pair, the histogram is obtained after normalizing by the total number of cases.

As can be seen from Fig. 8, using only wave model information to simulate σ_0 there is a fair agreement with the results of Abdalla's algorithm for wind speeds above 5 ms^{-1} . However, in the low windspeed range the wave model seems to overestimate the backscatter. The reason for this is related to the discretization of the two-dimensional wave spectrum, as the upper limit in the frequency spectrum is only about 0.5 Hz. Hence, for low wind speed the WAM model does not properly represent the local wind sea. Nevertheless, it can be concluded that a wave model provides a reasonable estimate of the nadir radar backscatter, and therefore, if observations of radar back scatter are available they could be assimilated by a variational analysis approach as is quite common nowadays at operational weather centres. Because of the high-frequency nature of the observed information it is expected that this analysis will result in an improved specification of air-sea interaction.

Extreme winds

Ku-band altimeter wind speeds are usually only regarded reliable in the wind speed range below 20 ms^{-1} . One of the reasons for this is that the slope $\partial\sigma_0/\partial U_{10}$ seems to decrease for increasing surface wind speed, hence σ_0 becomes less sensitive to changes in wind speed and thus an increased accuracy in the determination of the radar backscatter is required. Another reason is that the Ku-band backscatter is seriously affected by rain contamination, which is often present in extreme events such as Tropical cyclones and hurricanes. Furthermore, extreme events are rare and therefore it is difficult to develop proper empirical algorithms.

Nevertheless, Young (1993) addressed the special case of altimeter wind inversion at high winds by examining

³ Note that in the σ_0 values reported in the ENVISAT GDR products, a bias of -3.24 dB has already been applied to make the data consistent with ERS-2 ones. Thus, the *actual* ENVISAT backscatter is 3.24 dB higher than the ones in the GDR product.

the rare events when the GEOSAT altimeter swath crossed over strong cyclones. Ground truth was obtained from a model for cyclone winds for six extreme events within the period of 1987-1989. Input of the cyclone model were parameters such as central pressure and the radius to maximum wind. Here, central pressure was obtained from analyzed weather maps, while the radius of maximum winds was obtained from the distance between minimum and maximum σ_0 as Young made sure that only those satellite tracks were considered that went right through the eye of the storm. The resulting model function, valid for winds above 20 ms^{-1} relates GEOSAT Ku-band backscatter to wind speed using a simple linear model:

$$U_{10} = -6.4\sigma_0 + 72 \quad (7)$$

The truth as generated by the cyclone model is of questionable accuracy as analyzed central pressures at the end of the 1980's in the Southern Hemisphere were based on a manual analysis and therefore their quality may be fairly poor. Nevertheless, Gourrion *et al.* (2002) plotted Topex σ_0 as function of QuikSCAT wind speed in clear sky conditions (as inferred from collocated TOPEX radiometer data) and the parametrization in Eq. (7) was found to be a surprisingly good fit for wind speeds above 20 ms^{-1} . However, how accurate are Quikscat winds for extreme cases? Although the winds from SeaWinds using the QSCAT-1 model function may be regarded an improvement over past scatterometer models (e.g. Donnelly *et al.* 1999), a comparison of QuikSCAT winds against one year of buoy data involving 114 buoys suggests that for winds larger than 15 ms^{-1} Quikscat winds overestimate the truth. At a wind speed of 20 ms^{-1} the overestimation already amounts to 1 ms^{-1} . This is illustrated by a scatter diagram in Fig. 9. Therefore, in order to get a reliable estimation of the windspeed by

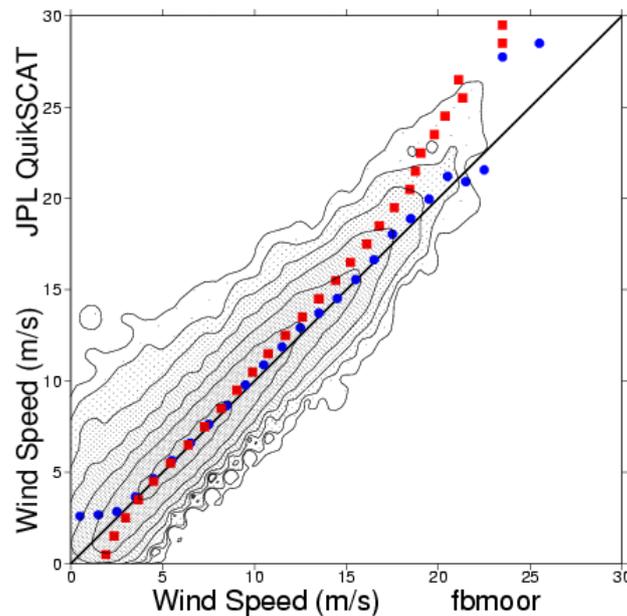


Figure 9: Comparison of QuikSCAT neutral winds with non-neutral wind speed observations from 114 buoys as received via GTS over the year 2004. The number of collocations is 106,497. An average positive bias of 0.24 ms^{-1} of QuikSCAT is consistent with the average increase of 0.2 ms^{-1} due to stability effects. For winds above $10\text{-}15 \text{ ms}^{-1}$, however, QuikSCAT winds show an overestimation of wind speed. Contour intervals in density are 10 dB . Circles (squares) represent binned averages for given buoy (QuikSCAT) wind speed.

the altimeter in extreme circumstances more work is evidently required.

Finally, measurements at Ku band are strongly affected by rain (see e.g. Quartly, 1997, 1998). Information from a radiometer is required to ensure that only observations are flagged as valid that are not affected by rain. However, in extreme circumstances such as those occurring near hurricanes a considerable number of high wind speeds occur during rainfall, which are missed when observed with a Ku band radar with serious consequences as the location of the low may be completely missed. A way to circumvent this omission is by combining observations from a dual-frequency altimeter at C band and Ku band (Quilfen *et al.* 2006) as is present on board of Jason. As the C band altimeter is almost insensitive to the effects of rain while the Ku band altimeter shows a considerable sensitivity to rain, there is the potential to simultaneously estimate wave height, wind speed and rainrate during hurricane conditions. Quilfen *et al.* (2006) carefully selected Jason altimeter tracks intersecting

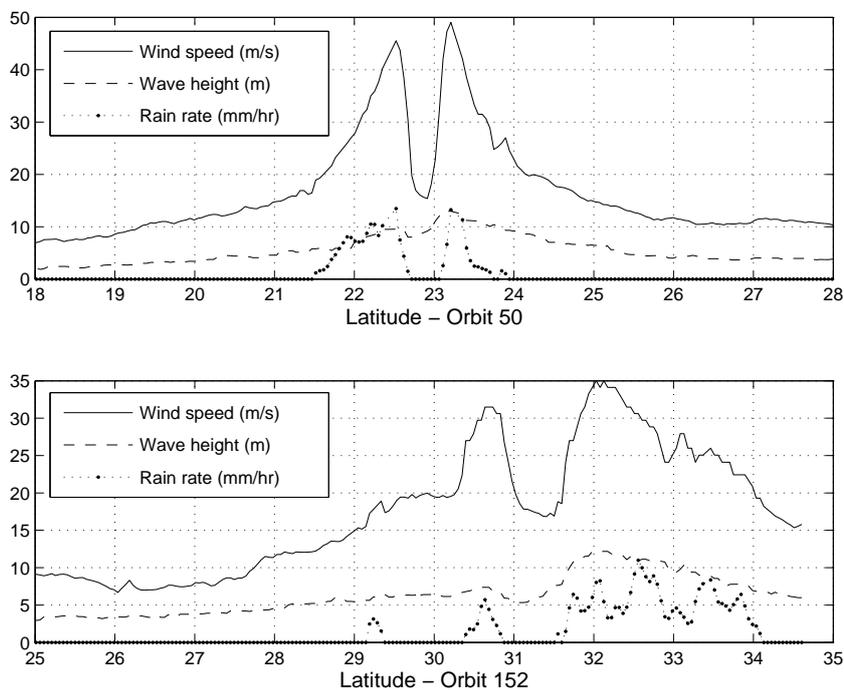


Figure 10: Wind speed (ms^{-1}), significant wave height (m), and rain rate ($mm\ hr^{-1}$) for Jason orbit 50 (top) and 152 (bottom) collocated with Tropical cyclone Isabel in mature and decaying phase, respectively. From Quilfen *et al.* (2006)

Tropical Cyclone Isabel using a systematic screening through the NOAA’s Hurricane Research Division (HRD) fields. A second selection was made to ensure that the Jason track was located in the vicinity of the eye of the hurricane. For Ku band, winds were retrieved using the Gourrion algorithm which was smoothly extended into the large wind speed regime ($U_{10} > 20ms^{-1}$) by means of Young’s (1993) algorithm (Eq. (7)). For C band, a modification to the algorithm was needed in order to take account of a different sensitivity to foam near the ocean surface. Rainrates were derived using the classical Marshall and Palmer (1948) relation for the frequency-dependent attenuation coefficient of radiation by rain. Finally, significant wave height was derived from an analysis of the C band waveform, as in rainy conditions significant wave height from Ku band may be seriously affected. Two examples of the retrieval of wave height, wind speed and rain rate near Tropical Cyclone

Isabel are shown in Fig. 10. The altimeter first sampled the rear left quadrant of the cyclone, followed by the front right one. Strong asymmetries, in particular for significant wave height, are noted, in agreement with the fact that extreme high sea states and winds are known to occur to the right of the direction of the movement of hurricanes. Therefore, dual-frequency altimeters have the potential to provide valuable information for extreme events, but an independent validation of the measured information is highly desirable. Quilfen *et al.* (2006) have made an attempt by comparing the results in Fig. 10 with the HRD wind analysis. This wind analysis is probably the best available wind field as it includes most existing observations. However, a direct comparison with in-situ observations of wave height, wind speed and rain rate is clearly desirable.

4 Significant wave height data

4.1 Data assimilation and diagnostics

Today, several operational numerical weather prediction (NWP) centres around the world run wave forecasting models that routinely assimilate altimeter wave height data, for example ECMWF (Lionello *et al.*, 1992; Abdalla *et al.*, 2004), Meteo-France (Skandrani *et al.*, 2004) and BoM (Greenslade *et al.*, 2001). Wave data assimilation is not as advanced compared to what has been done in the area of weather forecasting. There are two main reasons for this. First, in contrast to weather forecasting, the wave forecast is strongly constrained by wind forcing. The evolution of the atmospheric conditions is mainly controlled by the atmospheric initial state, whereas the initial wave field loses its influence after a relatively short time ranging from a few hours to a number of days depending mainly on the basin size, the sea state conditions, the wind strength and on the atmospheric dynamical time scale. In theory, a perfect wave model driven by perfect winds would produce perfect wave fields after a certain time, whatever the initial state might have been. However, this is not the case for the atmospheric model for which chaotic behaviour makes it very sensitive to its initial conditions. The second reason for the late introduction of wave data assimilation is that before the advent of satellites, only in-situ wave data were available. In particular, the observations from wave buoys are of high quality, but the data coverage is limited to coastal areas in the Northern Hemisphere. Therefore, these data are of little use in global wave data assimilation. In practice, these in-situ data are used for the control and monitoring of NWP and wave models.

The prospect of the advent of satellite data encouraged NWP centres to study the possibility of including wave data assimilation schemes in their operational wave forecast suites. For wave analysis, the wind fields are provided from the analysis of the atmospheric model. Satellite wave data are assimilated to improve the initial sea-state used for the wave forecast. Verified against independent in-situ wave measurements, the benefit of altimeter wave height data assimilation on wave height analysis can be clearly seen from Fig. 11, where for the years between 1999 and 2007 we show time series of the scatter index (standard deviation of error normalised with mean observed value) of wave height as function of time for the ECMWF operational analysis (with data assimilation) and for a hindcast run with the wave model that did not assimilate satellite data. In the summers up to 2004 there were considerable differences between wave height simulations with and without data assimilation, suggesting that satellite data assimilation played an important role in the quality of the wave height analysis. However, in the spring of 2004 and in the spring of 2005 wave model improvements were introduced into operations which had a considerable positive impact on wave forecasting performance. As a consequence, because of these model improvements, the impact of the assimilation of altimeter and SAR data on the accuracy of the wave height analysis has diminished from then onwards (see for more details the end of this section).

Data assimilation also has a positive impact on the wave forecast as can be seen from Fig. 12, but the size of

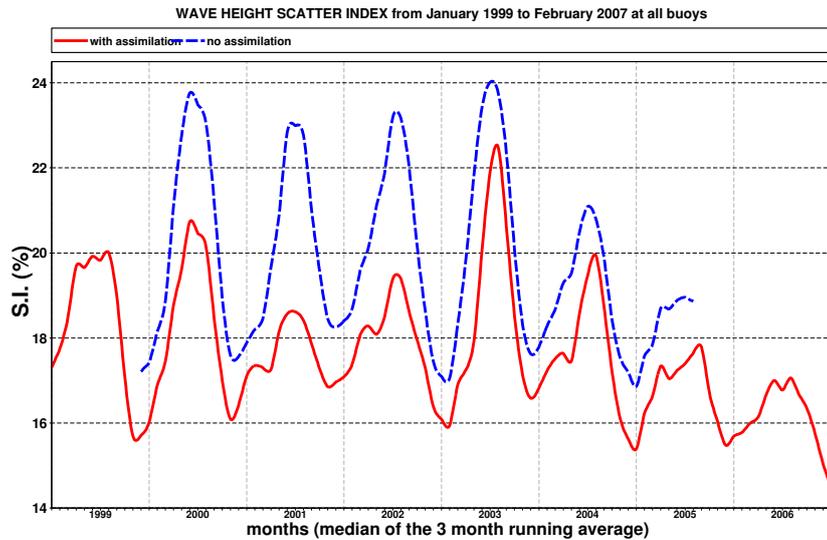


Figure 11: Impact of ERS-2 altimeter wave height data assimilation on ECMWF wave height analysis as compared to independent in-situ wave height observations between 1999 and 2007. Shown is the three monthly mean scatter index (S.I.).

the impact depends on the area. The left panels of Fig. 12 show the impact over the whole globe, and judging by the scores on the scatter index and the absolute correlation there is a modest impact up to 2 1/2 days in the forecast. The bias in wave height is on a global scale a minor problem. However, there are areas which have significant systematic errors. One such area is the eastern tropical Pacific. The corresponding forecast scores are shown in the right panels of Fig. 12. Clearly, forecast impact of data is much larger and longer lasting in areas where swell systems (which give a long memory to the forecast system because their lifetime is large) dominate and where there are significant systematic errors.

The degree of improvement in the wave height forecast achieved by assimilating wave data depends on a number of factors. We briefly discussed already the accuracy of the forecast surface winds and the quality of the wave model itself realizing that a good wave forecasting system will show less impact of wave data on forecast skill. A second important factor is the quality of the wave data that are assimilated in the model fields. This issue will be discussed in the next section while here we briefly discuss the third important factor, namely the assimilation procedure.

When the prospect of global wave height data emerged, the first assimilation methods that were developed were obviously the simplest and the least expensive in terms of computer resources. Several approaches are conceivable and they can be classified into two categories: sequential methods and multi-time level methods. The assimilation techniques most commonly used for operational applications are based on instantaneous sequential methods like Optimum Interpolation (OI) (e.g. Lionello *et al.*, 1992) and successive corrections (e.g. Thomas, 1988). Such methods are very attractive due to their low computational cost. However, the corrections are done at a local scale and at one time level. As from experience it is known that the main error source is the driving windfield, it would make sense to use the winds as control variable in the analysis scheme, i.e. to modify the winds in such a way that an optimal agreement with the observations for wave height is obtained. For windsea, updates to the wind field may be obtained in the context of a single time level approach. However,

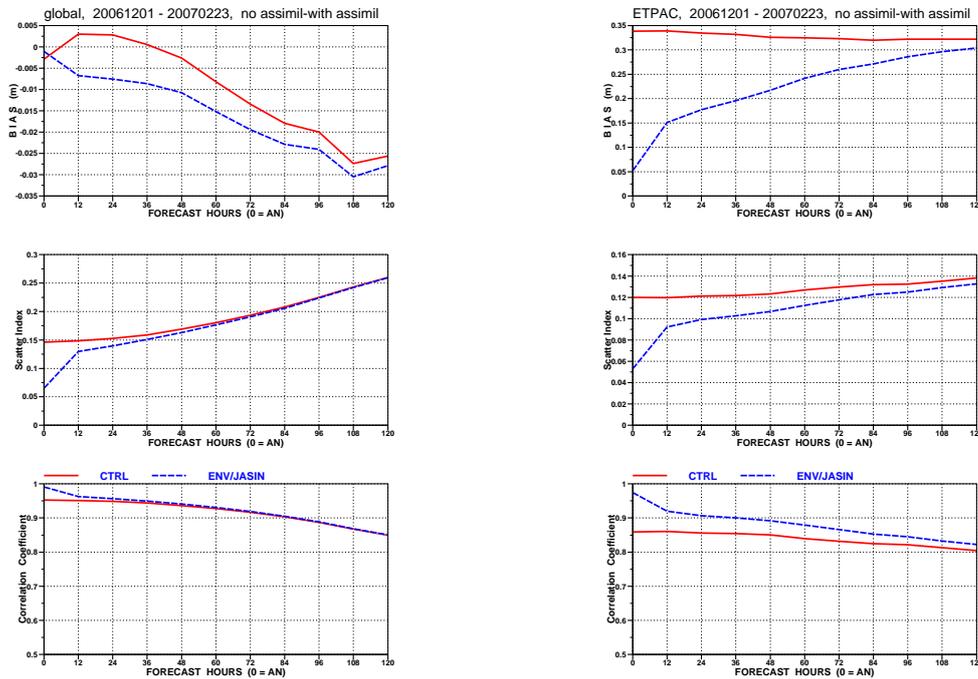


Figure 12: Impact of ERS-2 altimeter wave height data assimilation on ECMWF wave height forecast obtained by comparing wave height forecast against ENVISAT wave height observations. Period is about three months in the winter of 2006-2007. Left panels for the global area and right panels for the eastern Pacific.

in case of swell this approach does not work because swells have been generated by remote storms a number of days ago. The assimilation of altimeter wave height data represents an additional challenge because it only provides information on the integral over frequency and direction of the wave spectrum, whereas modern wave models are based on a spectral description. Applying the assimilation method results in a wave height increment which must be translated to a corresponding change in the local wave spectrum. For windsea, this is fairly straightforward to do by using the evolution laws for wind generated waves, as obtained, for example, during the JONSWAP (1973) field campaign. However, for swell it is assumed that during the transformation the mean square slope is invariant, which may be plausible, but in practice this assumption is hard to justify (Lionello *et al.*, 1992; Greenslade, 2001).

A number of multi-time level methods have been developed and tested in the past. One such a method is based on the Kalman Filter (KF). The KF has the additional advantage that it provides error statistics on the model variables. The KF propagates a forecast error covariance matrix (FECM) which gives further information on the model state. The problem of implementing these techniques arises from the dimension of FECM, which then has implications on the required number of model integrations. Some simplifications are required to reduce the cost of such methods (Voorrips, 1998).

Another multi-time level method is based on the variational approach. Such methods are based on the minimization of a cost function and often use the adjoint technique in order to compute the gradient of the cost function. Multi-level time variational techniques take into account the history of the observations under the constraint of the wave model dynamics. A promising first step was reported by de las Heras *et al.* (1994)

and Hersbach (1998) who managed to create the 'true' wind forcing based on wave height observations alone. These studies were using the adjoint of the WAM model which was based either on coding the analytical adjoint or by means of an automatic procedure. However, the high cost of these methods has slowed down their introduction in the field of wave forecasting, although the variational approach is by now widely used in operational weather forecasting. Simplifications are always required for operational implementations. For example, simplifications in the tangent linear model as well as a reduction of its resolution would allow a significant reduction of its cost as it is necessary to carry out between 10 and 100 integrations of the adjoint of the tangent linear model to converge towards the optimum trajectory. Such an approach works well in the atmospheric context because the main interest is in an analysis of the large scales. However, for wave modelling the main interest is in wave development during storms which have smaller scales, say of the order of 500 km, and therefore a simplification such as a reduction in resolution is probably counter-productive. Using an approximate tangent-linear model, Voorrips and de Valk (1997) compared results of the variational approach with an Optimum Interpolation method. No advantage was found for the variational method, however, presumably because of a not optimally calibrated tangent-linear model.

Finally, there is a problem common to all assimilation methods which has only recently been addressed at Meteo-France and by Greenslade and Young (2004). A key element in any data assimilation scheme is that knowledge is required of the spatial correlation of the model error. In most schemes this is modelled by a simple exponential distribution involving the ratio of the distance between the two points of interest and a correlation lengthscale, where the correlation lengthscale is regarded to be independent of location and is adjusted in such a way that in some sense optimal results are achieved. This is an ad-hoc approach and Greenslade and Young (2004) show, using the Hollingsworth-Lönnerberg (1986) method, how in a rational way the correlations in model error may be obtained by a comparison with superobbed altimeter wave height data (superobbing is required in order to avoid correlations that do exist between individual altimeter observations, see Janssen *et al.*, 2007). It then turns out that the correlation length scale is a function of location on the globe, with large spatial scales, of the order of 700 km, in the Tropics (where swells prevail), while in the extra-Tropics correlation scales are considerable smaller, of the order of 400 km, because windseas, which have smaller scales, are an important component of the sea state in the storm tracks. Greenslade and Young (2005) have used these inhomogeneous model error correlations in the Australian wave analysis system and an improved forecast skill was found when compared to the operational analysis system with a fixed correlation length scale of 300 km. Preliminary tests at ECMWF, however, have only shown a modest impact of this promising change in the model correlation matrix, hence more work is evidently needed.

Despite the above mentioned shortcomings, even today, most weather centres with wave modelling capabilities are using Optimum Interpolation schemes or related schemes to assimilate significant wave height from a number of altimeters. Admittedly, these schemes involve a number of strong assumptions to relate wave heights increments to changes in the wave spectrum. Therefore, instruments that would provide spectral information such as the Synthetic Aperture Radar (SAR) are ideally suited for wave assimilation because no assumption in the mapping from the observed information to spectral change is required. The introduction of SAR data assimilation has been tested in a number of studies (e.g. Breivik *et al.*, 1998; Dunlap *et al.*, 1998; Aouf *et al.*, 2006b) and has been introduced operationally at ECMWF in February 2003 (Abdalla *et al.*, 2004). Forecast impact of the introduction of SAR data was similar to that of altimeter wave height data, which suggests that, although the wave height assimilation schemes involve a number of strong assumptions, the analysis results seem nevertheless realistic.

In closing we note that in the course of the last fifteen years we have seen, coinciding with considerable improvements in the quality of the surface winds and the wave model, a gradual reduction of the size of the forecast impact by assimilating altimeter wave height data. Nevertheless, weather forecasting centres need to provide the best possible wind and wave forecast and need to deliver a forecast product with a consistent quality

Mean wave height increments for July 2001

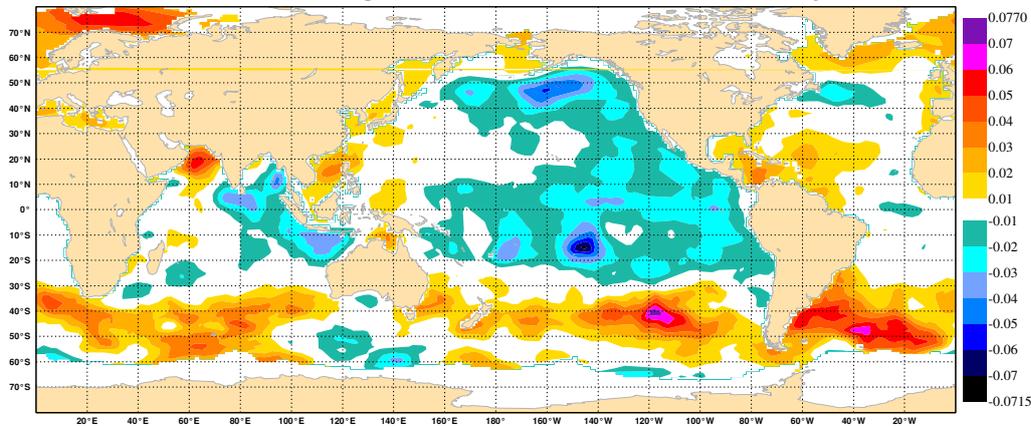


Figure 13: Spatial distribution of the analysis increments (analysis - first-guess) for July 2001, showing the at that time usual underestimation of wave height in the extra-tropical storm tracks. However, in particular in the Tropical Pacific a considerable overestimation of wave height by the model is noted.

in time. A way to deliver a consistent product is to use observations to constrain the analysis in order to keep it on the right track and for this reason assimilation of altimeter wave height data will continue. In this context there is a continuous confrontation between altimeter wave height data and their model counter parts, which can only benefit both. As an example we mention some problems with the ECMWF wave prediction system which were discovered in the summer of 2003, when for a number of weeks ERS-2 could not deliver observations of the sea state because the on-board tape drives had failed, as it turned out indefinitely. This loss of data had serious consequences for the quality of the ECMWF wave analysis as is plainly clear from Fig. 11, where the scatter index of the wave height analysis shows a maximum in the summer of 2003. It is of interest to study the spatial distribution of the error and therefore analysis increments (analysis-first-guess) for July 2001 are plotted in Fig. 13, revealing some large-scale systematic errors in the wave model. First of all, it shows the at that time usual underestimation of wave height in the Extra-Tropics (because the wind forcing was too small), while, in particular in the Tropical Pacific a considerable overestimation of model wave height is noted.

It turns out that these large positive biases are related to swell events generated by the storms in the Southern Hemisphere winter time. It is tempting to speculate on the causes of the overestimate by the wave model. An obvious candidate would be the dissipation source function, because this source term is the least well-understood. Another candidate is the representation of unresolved islands and atolls. Evidently, the results of Fig. 13 show that the main problem occurs in the Pacific ocean and not in the Atlantic. An important difference between the Pacific and the Atlantic ocean is that in the equatorial region of the Pacific there are a vast number of small islands and atolls which are not resolved by the present operational resolution of the wave model. Although these islands are small, they nevertheless block considerable amounts of low-frequency wave energy. Therefore, using a high resolution global topography of 2 minutes, a wavenumber dependent blocking factor was determined. This change was introduced operationally in March 2004 and, as can be seen from Fig. 11, gave a substantial reduction of the analysis scatter index in the Summer of 2004. Also, considerable improvements in the wave height forecast skill in the Tropics were found. Nevertheless, the problem did not

disappear completely.

In April 2005 a new version of the dissipation source function was introduced, which used an alternative definition of the integral parameters, such as mean frequency, in the expression for the dissipation. The new definition is given in Bidlot *et al.* (2007), see also Janssen (2007). As can be seen in Fig. 11, this change had a further beneficial impact on the accuracy of the wave height analysis in the N.H. summer of 2005 and subsequent summers.

4.2 Quality of Altimeter wave height data

It is clear that it is of the utmost importance to assess the quality and accuracy of the altimeter wave height observations. The typical engineering requirements of the older generation of satellites are that the accuracy of the significant wave height observations should be better than the maximum of 0.5 m and 10% of wave height in the range of 1 to 20 m. The newer instruments, such as the RA-2 altimeter on board of ENVISAT, have to satisfy the even more stringent requirement on accuracy of the maximum of 0.25 m and 5% of wave height.

How can one test whether these requirements have been satisfied? A first indication of the quality of the altimeter wave height product is to compare them with model wave height fields or with buoy observations. However, this will only give an upper limit to the errors of the altimeter wave height products as model and buoy data have errors as well. Initial work on the calibration of altimeter significant wave height used comparisons with buoys alone. Carter *et al.* (1992) used regression based methods to compare Geosat to the US buoy network and found that Geosat gave an estimate that was 13% low compared to the buoys. Cotton and Carter (1994, 1995) used an alternative method. Instead of using individual crossings of the buoy positions by the altimeter they looked at monthly means over $2^\circ \times 2^\circ$ squares from the satellites compared with monthly means from buoys within those squares. The results obtained were comparable with a point by point calibration establishing that a significant wave height climatology based on $2^\circ \times 2^\circ$ squares gives a reasonable approximation to the true wave climate. The regressions used in these papers assumed that the buoys had negligible error compared to the altimeter observations. These early comparisons showed that this was not the case and that better methods were needed that would take this into account. Cotton *et al.* (1997) and Challenor and Cotton (1997) used orthogonal distance regression (Boggs *et al.* 1987, 1989) to overcome this problem. This method assumes that the ratio of the error variances is known and in altimeter calibration work it has always been set to 1. However if data from more than two sources are included in the calibration it is possible to calculate the ratio of the variances directly.

The need for estimates of errors of different data sources was realized by Stoffelen (1998). He proposed to use a triple collocation method to calibrate observations of winds from a Scatterometer using winds from buoys, a model analysis and the ERS-1 Scatterometer. It is straightforward to show that with three data sets which have uncorrelated errors, the random error of each data type can be estimated from the variances and covariances of the data sets (Tokmakian and Challenor, 1999). However, unless additional assumptions are being made, it is not possible to perform an absolute calibration among the data sets, simply because there are not enough equations. A possible way out of this dilemma is to use a minimization procedure. Assume that the random errors are not correlated and that the errors of the three data sets are estimated using the triple collocation method. Given these estimated errors, calibration is then performed using the neutral regression approach of Deming (Mandel, 1964, see also Marsden, 1999), which is based on the minimization of the error in both variates. For an extensive discussion of this approach and a number of applications see Janssen *et al.* (2007).

Using the triple collocation method it was possible to estimate the random errors of the ENVISAT and the ERS-2 altimeter wave height. In this case it is emphasized that Fast delivery products are used which are averaged over a length scale which is compatible with the resolution of the ECMWF wave model (55 km).

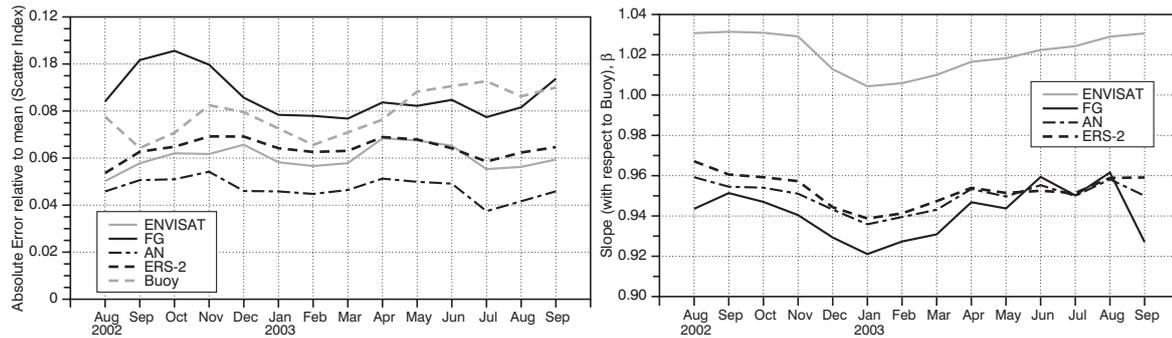


Figure 14: Left panel: Monthly Relative Error of ENVISAT, first-Guess, analyzed, ERS-2 Altimeter and Buoy wave height over the period August 2002 until September 2003. The right panel shows the corresponding monthly calibration constants β , where the reference was the buoy wave height. From Janssen *et al.* (2007)

ENVISAT was launched on March 1 2002 and was maneuvered in almost the same polar orbit as ERS-2, in such a way that the two satellites were only measuring 20 minutes apart. As a consequence, there are ample close collocations between the altimeter observations giving 5 collocated data sets, namely from the ENVISAT and ERS-2 Altimeters, from buoys, from model first-guess and analysis. Note that during the initial phase of the ENVISAT mission ENVISAT data were not assimilated, because the quality of the new data needed to be monitored first, while ERS-2 wave height data were assimilated. Therefore, the ENVISAT and the first-guess model errors can be regarded as independent. However, it was found that there are correlations between ENVISAT and ERS-2 Altimeter wave height errors because these instruments share the same measurement principle. When correlated errors are taken into account it is found that the relative errors (i.e. random error divided by the mean wave height) in wave height are respectively 6%, 6.5%, 8%, 9% and 5% for ENVISAT, ERS-2, buoys, first-guess and wave analysis.⁴ In the left Panel of Fig. 14 we have displayed the corresponding monthly mean relative error over the period from August 2002 until September 2003 (Note that in October 2003 ECMWF started with the operational assimilation of the ENVISAT wave height data and thus the assumption of independence of errors was from then onwards not satisfied anymore, cf. Janssen *et al.* (2007)). For calibration the buoy data were taken as the reference because they are the ground truth. The calibration constant β is then defined as the symmetric slope that is obtained when a fit is made against the buoy wave heights. The right panel of Fig. 14 displays the time series for the calibration constants β . It is striking to see that first-guess, analyzed and ERS-2 data underestimate wave height by about 5%, while ENVISAT overestimates wave height by about 2% on average.

From the triple collocation study so far only bulk statistics have been produced. With a mean wave height of 2.5 m one finds that the standard deviation of error is of the order of 15-16 cm, hence it is concluded that concerning wave height the ERS-2 and ENVISAT missions have comfortably satisfied the engineering requirements of 50 and 25 cm, respectively. On the other hand, there are biases, about -4% of wave height for ERS-2 and $+2\%$ for ENVISAT. These biases are, however, relatively small in magnitude realizing that the buoy data, although of high quality, are not perfect either. Nevertheless, there may be problems with the method of retrieving significant wave height from the altimeter waveform giving rise to systematic errors. For example, as pointed out in Section 2 it is assumed that the pdf of the sea surface elevation is Gaussian, but due to the fact that surface gravity waves have sharper crests and wider troughs there is a positive skewness which gives small deviations from Normality. These deviations lead to the well-known sea state bias in the altimeter range measurement (cf., e.g. Jackson, 1979; Lipa and Barrick, 1981; Hayne and Hancock, 1982 and Srokosz, 1986), but they also

⁴ An alternative approach was followed by Greenslade and Young (2004). They applied the Hollingsworth-Lönberg method to obtain the random model and ERS-2 error from the structure of the spatial correlation function. Assuming uncorrelated observations errors, they found a 12% ERS-2 error, while the Australian wave model error was 24%.

give rise to small corrections to the significant wave height (Janssen, 2000; Gómez-Enri *et al.*, 2007). The sign of the correction depends, however, on the manner the significant wave height is obtained from the waveform. For ERS-2, significant wave height was obtained from the slope of the waveform at the half power point. In that event, Janssen (2000) has shown that finite skewness gives rise to an increase in significant wave height of at most 10% in very steep wave conditions, but on average the correction is about 4%. On the other hand, for ENVISAT, wave height is obtained as a result of a fitting procedure of the theoretical waveform to the waveform data. In that case Gómez-Enri *et al.* (2007) find that a finite skewness will give a reduction in wave height of about 10 cm. Therefore, skewness effects could help to explain the biases seen in the ERS-2 and ENVISAT altimeter data.

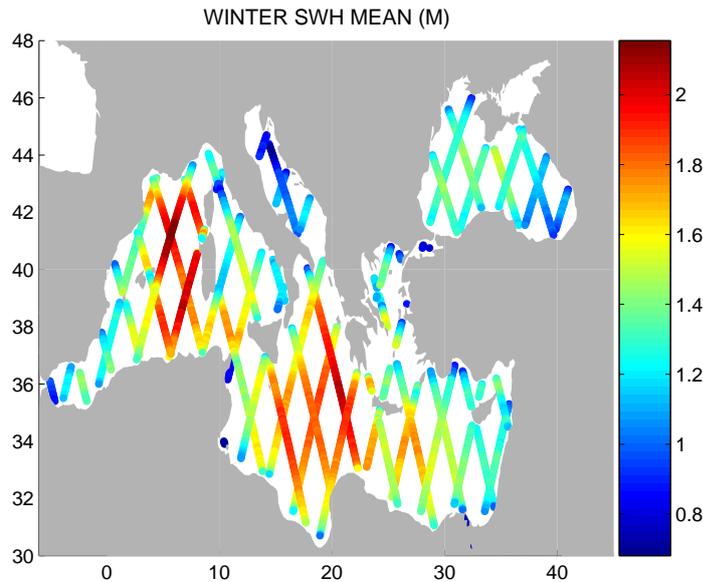


Figure 15: Mean wave height in the Mediterranean and the Black Sea obtained from 13 years of Topex-Poseidon altimeter wave height data. From Queffelec and Bentamy (2007)

Note finally that Gómez-Enri *et al.* (2007) also provide from the fitting procedure estimates of the skewness of the ocean surface as measured from space. However, remarkably there are large areas (in particular in the Tropics) where the skewness is found to be negative. This is odd as a straightforward estimate by Jackson (1979) shows that skewness should be positive. Clearly, more work is needed to understand this surprising result.

4.3 Sea-state Climatology

The design of ship, offshore and coastal structures requires a good knowledge of the most severe wind and wave conditions that they need to withstand during their lifetime. In the past and certainly in the pre-satellite era this knowledge was often difficult to obtain as the amount of observations of the sea state is fairly small because it is mostly confined to the shipping lanes in the Northern Hemisphere. Presently, there are four major sources of wind wave observations, namely, time series of buoys, long-term model hindcasts, satellite altimeter measurements and voluntary observing ship (VOS) data. Each of these sources can be used for the estimation of the wave climate and of climate variability. A well-known example is the 1% secular growth of significant wave height in the northeast Atlantic from the late 1960's to the early 1980s recorded at the Seven Stones Light Vessel and Ocean Weather Station L (Carter and Draper, 1988; Bacon and Carter, 1991,1993). Also, VOS wave

data, reported by marine officers worldwide since 1856, are an important source of wave data for wave height climatology. In the past, a large effort was devoted to correct these observations of many biases and to minimize observational errors inherent to visual observations. In addition, these observations have been validated against in-situ buoy data, altimeter data and wave model hindcasts. The strong point of the VOS data set is that it covers an extensive period. Gulev and Grigorieva (2004) developed a 120-yr-long homogenized time series of significant wave height for the well-sampled locations of the major ship routes. They studied trends in the annual mean significant wave height in the North Pacific and the North Atlantic. A major weak point of the VOS observations is that they are confined to the shipping routes only which limits the spatial distribution of the significant wave height field.

Satellite observations from altimeters cover the globe with quite homogeneous sampling and in the future are expected to become superior with respect to the VOS observations. However, at present these data cover a too short period to be of value for climate variability studies. Also, when using different satellites intercalibration of the different instruments is required, but, as discussed in the previous section intercalibration procedures using triple collocation methods are available nowadays. Nevertheless, altimeter data already give valuable information on the climate as shown in Fig. 15 by the example of wave climatology in the Mediterranean in the winter time obtained from 13 years of Topex-Poseidon data (Queffeuilou and Bentamy, 2007). Clearly, due to the special configuration of the Topex-Poseidon satellite tracks there are many spatial gaps in the wave height

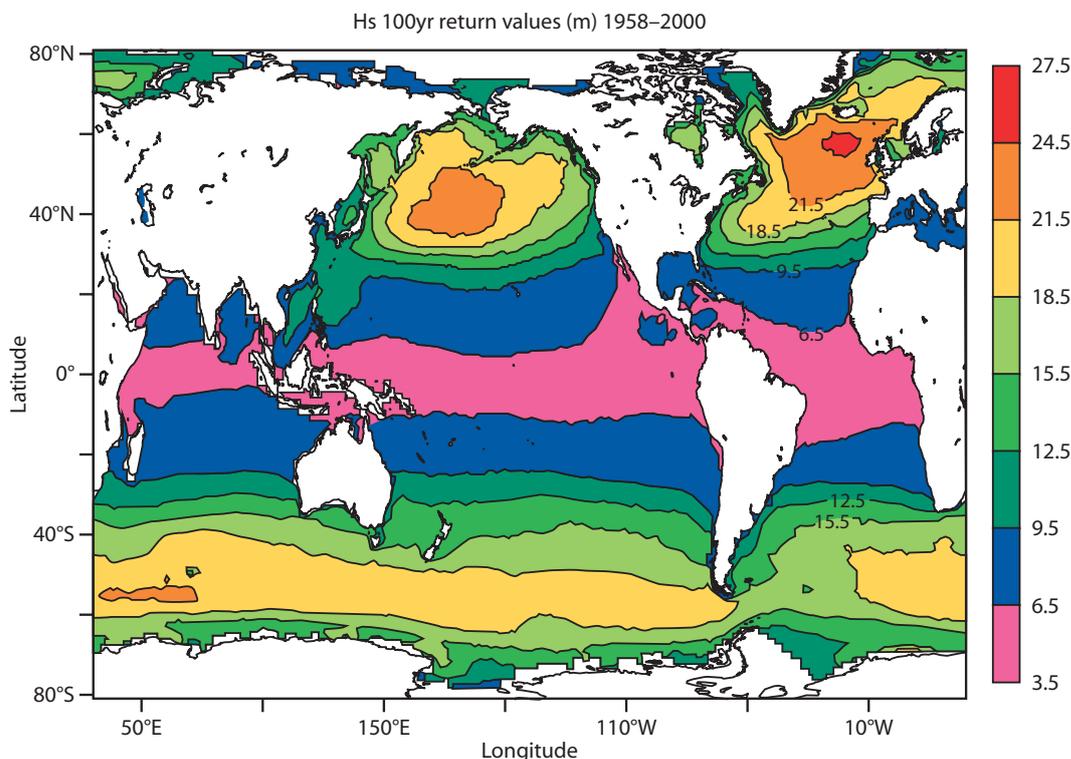


Figure 16: Corrected 100-year return values of H_S based on ERA-40 data for the whole ERA-40 period (From Caires and Sterl, 2005)

climatology derived from the altimeter wave height data. Although the period covered by satellite altimeter data is relatively short some work has been done looking at interannual variability and relating this to climate indicators. The first attempts to look at variation in the wave climate concentrated on the seasonal time scale (Challenor *et al.*, 1990; Carter *et al.* 1991). As the record increased in length interannual scales could be considered. Woolf *et al* (2002, 2003) looked at the interannual changes in wave climate in the North Atlantic

referred to above. They showed that these changes were related to the North Atlantic oscillation (NAO). By comparing the patterns of variability from year to year in the altimeter data, they produced a simple relationship between winter wave height in the North-East Atlantic and the NAO index. Hindcasts produced from this relationship compared well with both Geosat altimetry and with in situ data from weather ships. Wimmer *et al.* (2006) looked at the estimation of extreme waves from altimeter data alone.

For studying changes in wave climate wave hindcasts simulated by advanced wave models driven by reanalyzed winds are attractive. In particular when this is combined with an assimilation of appropriately calibrated altimeter wave height data, an optimal, continuous product for wave height climatology is to be expected. This is the approach followed by Caires and Sterl (2005). Their starting point was wave fields obtained from ERA-40 (Uppala *et al.*, 2005), which is a global reanalysis of meteorological and ocean-wave variables from 1957 to 2002. The data consist of 6-hourly fields on a $1.5^\circ \times 1.5^\circ$ latitude/ longitude grid covering the whole globe. Initial validation of the data revealed a generally good description of variability and trends. However, because of the coarse resolution, underestimation of high wave height and wind speed events was noted as well. Therefore, 100-year return values of significant wave height obtained from the ERA-40 reanalysis need to be corrected for this underestimation. Caires and Sterl (2005) used the peaks-over-threshold method to estimate 100-year return values from buoy data and model data and the model data were corrected using a linear regression based on a comparison with the extreme statistics from the buoy data. Buoy data, although of high quality, are limited to a restricted number of locations. In order to consolidate this linear calibration, a comparison between modelled and global, independent Topex-Poseidon extreme statistics was performed as well. The resulting 100-year return significant wave height map is given in Fig. 16. This plot shows that the most extreme wave conditions are in the storm track regions and that the highest return values occur in the North Atlantic. In agreement with operational experience the highest average conditions are found in the Southern Hemisphere storm track, but the most extreme conditions occur in the North Atlantic. Caires and Sterl (2005) also studied certain aspects of climate variability, but this aspect of the reanalysis is somewhat uncertain. Inhomogeneities in reanalysis winds could influence the patterns of climate variability in the wave hindcasts. These inhomogeneities are primarily present in the Southern Hemisphere (and to a lesser extent in the Northern Hemisphere) where due to the introduction of satellite data there has been a massive increase in the number of data used in the atmospheric analysis. Although the ERA-40 team took great care in intercalibration of the most important satellite data sources and in removing systematic differences between the model and the satellite data, there is of course no guarantee that the introduction of new data types might not lead to artificial trends. Despite this caveat, it seems that the Caires-Sterl approach is the most promising way forward to obtain a reliable wave height climatology. Clearly, altimeter wave height data play a vital role in the establishment of such a climatology, but these data should become available over a much longer period.

5 A new development and concluding remarks

From the discussion given so far it is evident that altimeter missions have proven their usefulness by measuring significant wave height and the surface windspeed over the oceans. A shortcoming of the altimeter missions is that they cannot provide information on the spectral distribution of wave energy with respect to frequency and propagation direction of the waves. The Synthetic aperture radar (SAR) is currently the only technology for estimating the two-dimensional wave spectrum from a spaceborne sensor. But SAR has a major drawback as it is well-known that the wavelike patterns visible in SAR images of the ocean surface may be considerably different from the actual ocean wave field. The motion of the ocean surface leads to Doppler misregistrations in the azimuth (flight direction) leading to a distorted image spectrum and even a cutoff in the azimuth direction. This so-called velocity bunching effect is proportional to the range-to-velocity ratio of the platform. For present and future missions this ratio is high, giving a considerable limitation of the usefulness of SAR spectra over

the oceans. Nevertheless, using model spectra as a first guess in the nonlinear inversion scheme of Hasselmann and Hasselmann (1991), assimilation of SAR spectra commenced at ECMWF in February 2003 (Abdalla *et al.*, 2004) with some positive impact on forecast skill.

However, owing to the velocity bunching effect there is only useful information from SAR for the long waves as the azimuthal cutoff wave length is typically larger than 200 m. It is therefore highly desirable to explore new capabilities to avoid the limitations of a SAR. Such a capability is already known for some time as a Real Aperture Radar (RAR) does not suffer from the velocity bunching effect, but would usually require a very large antenna to achieve the required resolution, except when an intelligent method of scanning is applied. Such a possibility is offered with the satellite project CFOSAT (Chinese French Oceanic Satellite), which will embark on a Ku-band radar whose technology is derived from altimeter techniques, but will use a geometry and data processing appropriate for the estimation of directional spectra of long waves. The concept is derived from a development in the 1970s on board airplanes (Jackson *et al.*, 1985 a-b; Hauser *et al.*, 1992). The initial spaceborne concept proposed under the name SWIMSAT with one beam pointing at 10° incidence (Hauser *et al.*, 2001) has now evolved into a six beam system, one pointing nadir and five pointing off-nadir between 2° and 10° . The five off-nadir beams, each having a footprint of approximately 18 km in diameter, will scan over 360° in azimuth. The principle is to measure the modulation of the radar signal due to the tilt of the long waves in each look direction. The Fourier analyzed modulation will provide the radial wave spectrum, while the azimuth scan will be used to determine the propagation direction. Estimates of the 2D wave spectrum will be possible at a scale of 70×70 km, for waves ranging from about 70 m to more than 400 m. The feasibility of such a concept was demonstrated in Hauser *et al.* (2001) and the impact of assimilating spectral data in wave forecast models was assessed in various recent studies using either simulations (Aouf *et al.*, 2006a) or SAR observations (Aouf *et al.*, 2006b). Results show that the assimilation of spectral information gives forecast skills, which are improved with respect to the case of assimilation of significant wave height only and that this improvement is the most significant for mean wave direction and mean frequency. Also, the role of the cutoff wavelength was studied. For SWIMSAT, with a cutoff wavelength of 70 m, a larger impact on forecast skill was found than for a SAR (with a typical azimuthal cutoff wavelength of 200 m). But, these were results obtained with ECMWF windfields from October 2000. In the course of time the quality of the windfields and the wave model have increased considerably, thus most likely reducing the impact of data assimilation as discussed already in section 4.1.

Let us now summarize the conclusions of this paper. It is indeed true that in the past 15 year or so we have seen massive improvements in our ability to forecast ocean waves. The reasons for these improvements have been discussed to some extent in this paper: vast improvements in atmospheric modelling resulting in accurate surface winds, a large increase in observations of the sea state and the atmosphere, notably provided by satellites, and the more realistic modelling of the dynamics of ocean waves, which is discussed in Janssen (2007).

In particular, altimeter sea state data have played an important role in the progress in ocean wave forecasting. On the one hand, altimeter sea state data can be used very effectively as a tool to diagnose problems in the atmospheric and ocean-wave model. On the other hand, despite the improvements in wave forecasting, altimeter wave height data are important in improving the wave height analysis. Furthermore, the wave height and wind speed data play a key role in the construction of a global wave height climatology, but altimeter observations over a much longer period are needed.

The retrieval of significant wave height from the altimeter waveform seems to be relatively well-understood, while triple collocation studies do suggest that altimeter wave height data are more accurate than expected from the engineering requirements. Although there is still discussion in the literature about how to obtain optimal benefits from the measurement of the radar backscatter σ_0 , the wind speed product derived from it has nevertheless proven its value. The altimeter wind speed has about the same accuracy as the wind speed from scatterometers. In addition, even for high wind speed this product is useful. In particular, when considering

dual-frequency altimeters there is the potential to obtain windspeed, wave height and rain rate at the same time. Finally, there is the promise of obtaining two-dimensional wave spectra from SWIMSAT in a relatively straightforward manner.

In conclusion, sea state parameters from altimeters have proven to be of tremendous value for ocean wave research and operational wave forecasting. And there is the promise for more valuable products to come.

Acknowledgement This paper is dedicated to the memory of Anthony Hollingsworth, who saw at an early stage the benefits described in this paper of comparing and validating satellite data and model counter parts.

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