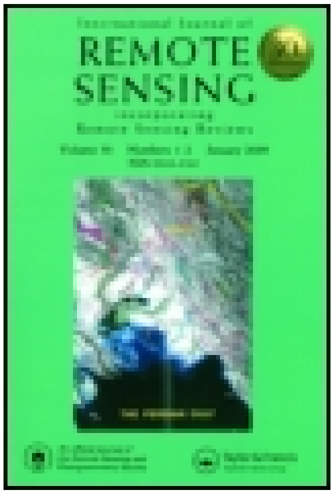

International Journal of Remote Sensing
Publication details, including instructions for authors and subscription information:
http://www.tandfonline.com/loi/tres20

Wave studies with the radar altimeter

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Published online: 27 Apr 2007.

To cite this article: P. G. CHALLENOR & M. A. SROKOSZ (1991) Wave studies with the radar altimeter, International Journal of Remote Sensing, 12:8, 1671-1686, DOI: 10.1080/01431169108955200

To link to this article: http://dx.doi.org/10.1080/01431169108955200

INT. J. REMOTE SENSING, 1991, VOL. 12, NO. 8, 1671–1686

Wave studies with the radar altimeter

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Abstract. In this paper we review the physical understanding and the algorithms necessary to extract wave information from the radar altimeter return from the sea surface. This allows us to assess the progress made since the launch of Seasat, to identify areas where further work is necessary in order to develop our understanding of the basic mechanisms of radar interaction with the sea surface, and to suggest how new wave parameters might be obtained from the radar return. Following on from this we review the wave studies that have been carried out using Seasat and Geosat altimeter data and present some new results, on the joint probability distribution of significant waveheight and wind speed, which we have obtained recently from Seasat data.

1. Introduction

The aim of this paper is to give a flavour of what radar altimeter data can tell us about waves on the sea surface. To this end, we will review the physical understanding and algorithms necessary to estimate wave parameters from the radar return from the ocean and then go on to suggest how some new wave parameters might be obtained from the return. Additionally, we will review the wave studies that have been carried out since the launch of Seasat and present some new results recently obtained from Seasat data, on the joint probability distribution of significant waveheight and wind speed. The contents of this paper reflect the particular interests of the authors and we lay no claim to being either comprehensive or exhaustive in our look at wave studies using the radar altimeter. However, we hope that the potential of radar altimeter wave data, for both scientific studies and practical applications, will be conveyed to the reader.

To set the scene, it is worth noting that in 1978 Seasat provided the first global measurements of the significant waveheight and, despite its short lifetime, showed the potential of satellite-borne radar altimeters of obtaining such measurements. Earlier altimeters, flown on Skylab during 1973 and GEOS-3 between 1975 and 1978, had given wave data, but unlike Seasat, they were limited by not having the capability of continuous operation. Subsequently, the Geosat mission which was started in 1985 and is still in operation has fulfilled some of the promise shown by Seasat by providing continuous radar altimeter data over a period of almost three years.

Essentially, the radar altimeter is a relatively simple instrument when compared to, say, a wind scatterometer or a SAR. It works by transmitting a short pulse (about three nano-seconds) of radar energy which is reflected back from the sea surface. The

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travel time of the pulse gives the height of the satellite above the surface and is used to measure sea surface topography (a topic discussed in detail elsewhere in this volume). In addition, the distortion of the pulse shape and the backscattered power in the return give information about the waves and wind at the sea surface. The estimation of wave and wind parameters from the shape of the return and the backscattered power are the subject of the following section.

2. Estimation of wave parameters

Wave parameters are estimated from the radar altimeter data by fitting a model of the return from the sea surface to the measured return and adjusting the parameters to obtain the 'best' fit (in some sense). It is well known that the altimeter works by specular reflection of the radar pulse from the sea surface and that the return P_r can be expressed mathematically as the convolution of three terms (Brown 1977); thus

$$P_r = P_{FS} \otimes P_{PT} \otimes P_{spec} \tag{1}$$

where P_{FS} is the flat surface response, P_{PT} is the point target response, and P_{spec} is the statistical distribution of specular reflectors on the surface of the waves.

P_{spec} is dependent on the statistics of the waves on the sea surface and therefore the model of the return is dependent on the assumptions that are made about these statistics. Two types of assumptions have been made; these are, first, that the waves are linear and therefore their statistics are Gaussian and, second, that the waves are weakly nonlinear and their statistics are non-Gaussian. Strictly speaking, neither of these assumptions is correct as they fail to take account of strongly nonlinear wave phenomena, such as wave breaking, nevertheless, both have been found to yield useful descriptions of wave statistics for many applications. As will become apparent later they also prove useful in understanding radar altimeter return from the sea surface.

2.1. *The Gaussian case*

If we assume that the statistics of the waves are Gaussian then the statistics of the specular reflectors are also Gaussian, with the same variance as the sea surface elevation (Barrick and Lipa 1985, Srokosz 1986 b). This means that the return may be expressed as

$$P_r = P_r(\sigma_0, H_s, t_0) \tag{2}$$

where σ_0 is the backscattered power, H_s is the significant waveheight ($=4\sqrt{m_0}$, where m_0 is the variance of the sea surface elevation), and t_0 is the travel time of the pulse, which gives the height of the satellite above the sea surface. Thus we can see that the primary wave parameter influencing the radar return is H_s and this is the fundamental wave parameter that can be derived from the return.

It can also be shown that the backscattered power σ_0 is related to the statistics of the waves by

$$\sigma_0 \propto 1/s_{max} s_{min} \tag{3}$$

where $s^2_{max,min}$ are the maximum and minimum variances of the wave slopes, respectively measured in two orthogonal directions (Barrick and Lipa 1985). Clearly, therefore, σ_0 can give further information about the statistics of the waves. To our knowledge only Brown (1979), using GEOS-3 data, has used the relationship between

σ_0 and s^2 to look at the slopes of the waves. In most analyses of radar altimeter data investigators have preferred to relate σ_0 empirically to the sea surface wind speed, either directly, or indirectly through s^2, which in turn was related to the wind speed through other measurements (such as those of Cox and Munk 1954). As we are concentrating on wave measurements in this paper we will not discuss the technicalities of altimeter measurement of wind speed here. We refer the reader to the review of Chelton and McCabe (1985) for more information on this topic.

An alternative to using σ_0 to estimate the wave slope variances is to use the linear dispersion relationship for water waves, connecting the wavenumber of the waves and their frequency (Phillips 1977), to obtain the following result (by transforming the wavenumber spectrum of the waves into a directional frequency spectrum) (Challenor and Srokosz 1984)

$$s_{\max} s_{\min} \propto m_4 \tag{4}$$

where m_4 is the variance of the vertical acceleration of the sea surface (m_4 is also the fourth moment of the frequency spectrum of the waves, see next section). The above relationship holds under the assumption that the directional dependence of the wave spectrum can be decoupled from its frequency dependence (Challenor and Srokosz 1984) and this includes the special case of an isotropic sea surface, for which $s_{\max} = s_{\min}$.

Snyder and Kennedy (1983) and Srokosz (1986 a) have proposed two somewhat different models to describe the statistics of wave breaking, but common to both models is their dependence on the parameter m_4. Snyder and Kennedy's model assumes that breaking occurs when the downward acceleration of the sea surface exceeds αg, while Srokosz's model assumes that a wave is breaking if the downward acceleration at the crest exceeds αg (here g is the acceleration due to gravity) (see Srokosz 1986 a), for a discussion about the choice of the parameter α). Snyder and Kennedy (1983) estimate the fraction β of the sea surface covered by breaking waves as

$$\beta = 1 - \Phi(\alpha g / \sqrt{m_4}), \tag{5}$$

while Srokosz (1986 a) estimates the percentage B of breaking waves as

$$B = \exp\{-\alpha^2 g^2 / 2m_4\}. \tag{6}$$

Here Φ is the cumulative normal distribution. Given that m_4 may be estimated from the altimeter measurements of σ_0, these formulae could be used to estimate and map the global distribution of breaking waves.

2.2. *A new period parameter*

We have seen above that it is possible to estimate the significant waveheight H_s from the radar return. For many applications, for example in offshore engineering, some knowledge of the wave period is also desirable. Here we will define a new period parameter T_A as

$$\begin{aligned} T_A &= 2\pi (m_0/m_4)^{1/4} \\ &\propto H_s^{1/2} \sigma_0^{1/4} \end{aligned} \tag{7}$$

which can be estimated from the altimeter measurements of H_s and σ_0.

Commonly used estimates of wave period are based on the frequency spectrum

$S(\omega)$ of the waves (here ω is the radian frequency) and is moments m_n, given by

$$m_n = \int_0^\infty \omega^n S(\omega)\, d\omega. \tag{8}$$

The usual period parameters are the peak period T_p, based on the frequency at the spectral peak ω_p, given by

$$T_p = 2\pi/\omega_p, \tag{9}$$

the 'mean' period T_1, actually based on the inverse of the mean frequency, given by

$$T_1 = 2\pi(m_0/m_1), \tag{10}$$

the mean zero-upcrossing period T_z, given by

$$T_z = 2\pi(m_0/m_2)^{1/2} \tag{11}$$

and the mean crest period T_c, given by

$$T_c = 2\pi(m_2/m_4)^{1/2}. \tag{12}$$

From this it can be seen that T_A differs from all the commonly used period parameters, but that it is the geometric mean of T_z and T_c. Thus $T_A = \sqrt{(T_z T_c)}$.

Using Seasat altimeter data, from passes over the area of the JASIN experiment, we have compared altimeter estimates of T_A to measurements of T_1 and T_z from Waverider buoys, a pitch-roll buoy and a shipborne wave recorder, which were in operation at the time of the Seasat overflights. The results are presented in figures 1 and 2. From these no clear trend of T_A with either T_1 or T_z can be detected, although if only the Waverider buoy data are considered more of a trend is evident (this may reflect some of the problems with *in situ* wave measurements). This comparison is based on too few data points to draw any final conclusions about the usefulness, or otherwise, of estimating the parameter T_A from altimeter data. However, even if it cannot be used in a quantitative way, it may provide a qualitative means of looking at changes in the wave period.

2.3. *The non-Gaussian case*

Jackson (1979) for the one-dimensional case, i.e., for a uni-directional wavefield, and Barrick and Lipa (1985) and Srokosz (1986 b) both for the two-dimensional case, i.e. for an omni-directional wavefield, have shown that, for weakly nonlinear waves with non-Gaussian statistics, the radar return may be expressed as

$$P_r = P_r(\sigma_0, H_s, \lambda, \delta, t_0) \tag{13}$$

where λ is the skewness of the sea surface elevation, a measure of the nonlinearity of the waves as exhibited by their peakier crests and flatter troughs, and δ is a 'cross-knewness' parameter, related to the correlation between the elevation and slope-squared (see Srokosz 1986 b).

By including nonlinear wave effects and by fitting the model return to the actual return it is possible to gain more information about the surface waves. In particular, the parameter λ gives a measure of the nonlinearity of the waves.

The use of a non-Gaussian statistical model is of interest not only becuause it allows extra wave parameters to be estimated from the return, but also because it enables the problem of sea-state bias in the altimetric height measurements to be understood and a correction made. Sea-state bias in the height measurement (note

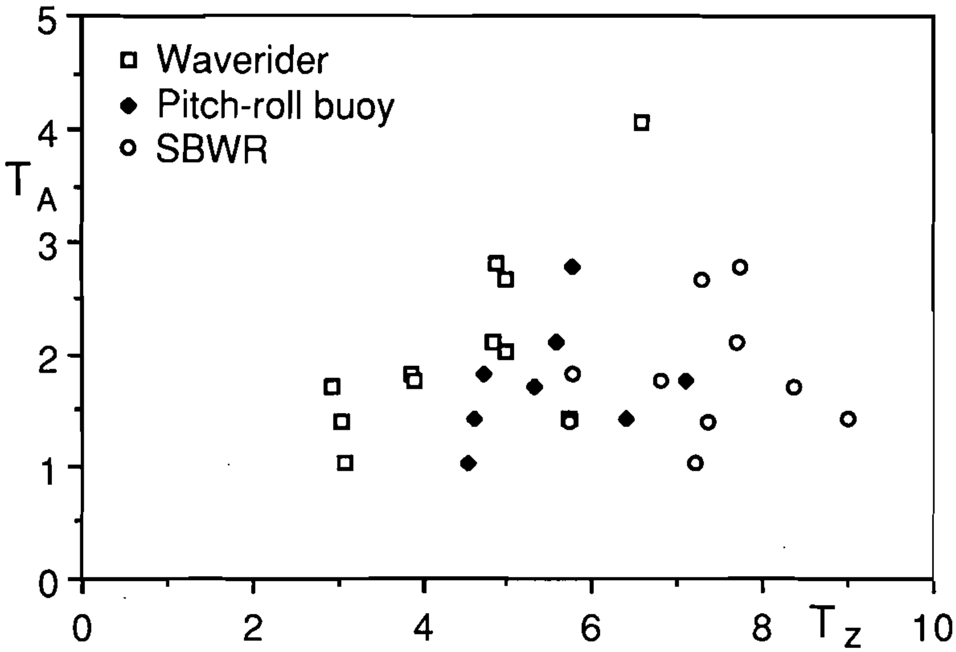

Figure 1. The new altimeter-derived period parameter T_A compared to Waverider, pitch-roll buoy and shipborne wave recorder estimates of T_z. Altimeter data from Seasat *in situ* data from the JASIN experiment. (Period in seconds.)

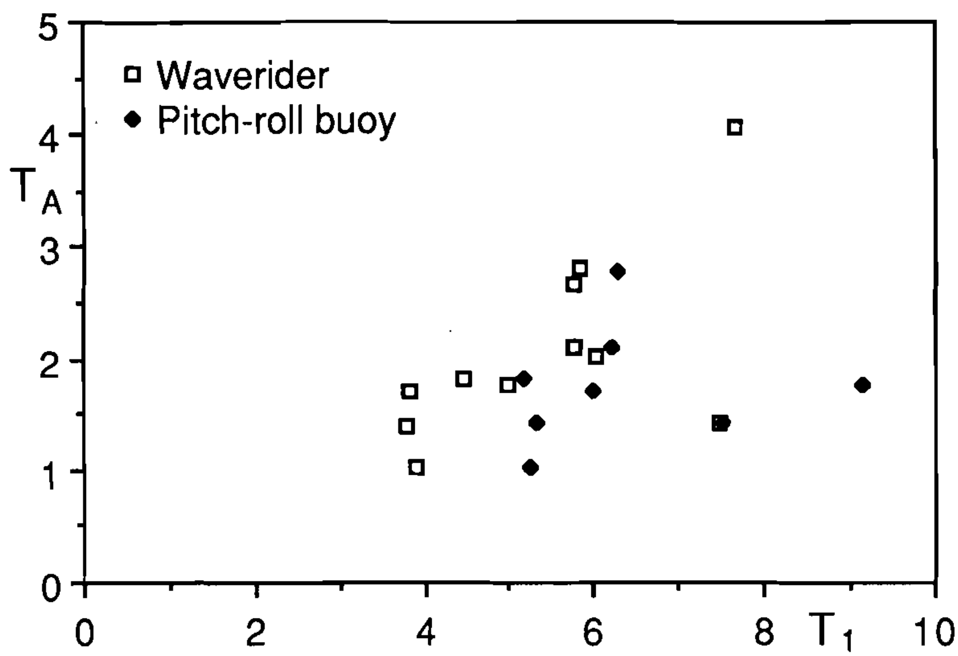

Figure 2. The new altimeter-derived period parameter T_A compared to Waverider and pitch-roll buoy estimates of T_1. Altimeter data from Seasat, *in situ* data from the JASIN experiment. (Period in seconds.)

that this should be measured to the mean sea surface elevation) arises because of the on-board algorithm (known as the tracker) used to estimate the travel time of the radar pulse down to the sea surface and back to the satellite. The tracker estimates the travel time (or, equivalently, the height) by locating the half-power point of the return, which corresponds to the median of the distribution of the specular reflectors. In the Gaussian case the distributions of the surface elevation and the specular points are identical; both being Gaussian, so that their medians and means are equal. Thus, estimating the median of the specular reflectors is equivalent to estimating the mean of the surface elevation (the quantity of interest). However, in the non-Gaussian case, there is a difference between the median of the specular reflectors and the mean of the surface elevation, which depends on λ, δ and H_s. This gives rise to the so-called sea-state bias in the height measurement. By allowing for the effects of wave nonlinearity we can correct for this bias and thus obtain more accurate height measurements. If the height measurement is to be used to estimate the sea surface topography and hence the geostrophic surface currents, this is an important correction as the magnitude of the bias is of the same order as the accuracy to which the height measurement is required (see Srokosz 1986 b, 1987, for further discussion of this effect).

2.4. *Algorithms*

Generally radar altimeters process the return signal to estimate σ_0, H_s and t_0 in real-time using on-board algorithms known as trackers. To date these have been based on the assumption that the statistics of the waves at the sea surface are Gaussian (for the sake of simplicity). Seasat, Geosat and Topex use (or will use) a weighted split-gate algorithm to analyse the return. ERS-1 and Poseidon use Sub-optimal Maximum Likelihood Estimation (SMLE). This is a simplified form of Maximum Likelihood Estimation (MLE), which gives minimum variance, unbiased estimates of the parameters and should allow for more accurate on-board estimation of the parameters.

As noted in the previous section the assumption of Gaussian statistics leads to the problem of sea-state bias and all present on-board trackers cannot account for this effect as they do not estimate λ and δ. Empirical corrections have been applied for sea-state bias (for example, Douglas and Agreen 1983), but it would be much better to re-analyse the return on the ground, using a non-Gaussian model, and obtain a correction explicitly. Srokosz and Challenor (1988) propose that the best approach would be to use MLE, which allows for the effect of the fading noise on the return, to obtain accurate estimates of the parameters. Some re-analysis of altimeter returns on the ground has been carried out for GEOS-3 (Fedor *et al.* 1979), Seasat (Lipa and Barrick 1981, Hayne and Hancock 1982, Guymer and Srokosz 1986) and Geosat (Hayne and Hancock 1987). However, for future altimetric missions, with more stringent requirements on the height measurement, it may be necessary to re-analyse all the returns on the ground to eliminate sea-state bias.

This concludes our brief overview of the estimation of wave parameters from radar altimeter data. We will not go on to consider the uses to which the estimates of the parameters may be put.

3. Studies using the estimated wave parameters

3.1. *Global and seasonal H_s distributions*

Given the global measurement capability of satellite-borne altimeters it is possible to study the global wave climate, in particular, the variations in H_s. The first such description of significant waveheight was given by Chelton *et al.* (1981), who calculated the global distribution of the mean value of H_s for the whole of the 100 day Seasat mission. Subsequently, Sandwell and Agreen (1984) used GEOS-3 data to look at seasonal variations in the global distributions of H_s. Their study was limited due to the incomplete spatial coverage afforded by GEOS-3, but took advantage of the fact that the data acquired covered the three year period of the mission. More recently Carter *et al.* (1990) and Challenor *et al.* (1990) have analysed one year of Geosat data to look at global monthly and annual means of H_s. The results of these studies show the existence of high waves in the Southern Ocean, presumably due to the higher wind speeds there and the much longer fetches over which they are generated. They also show that there is less seasonal variability in the southern hemisphere wave climate than in the northern one. More details can be found in the references cited.

These studies show the potential for mapping the global wave climate using satellite altimeter data. In the future these studies could be extended to look at the variability of H_s, on monthly, annual and interannual time-scales, so that its variance around the mean values computed to date may be examined. This will become feasibile as more altimeter data is acquired over the next few years.

3.2. *The spatial variability of H_s*

Another use of the altimeter measurements is to examine the spatial variability of H_s. This cannot be done from conventional *in situ* measurements of waves, which are generally time series measured at a single point. The spatial variability is of interest for a number of reasons. First, in calibrating and validating the satellite measurements against *in situ* data, it is highly unlikely that the satellite track will pass directly over the *in situ* measuring device or that the time of the overpass will coincide with time at which the *in situ* measurement is made. It is therefore important to know how close, in time and space, the satellite measurement needs to be to the *in situ* measurement, in order that a meaningful comparison may be made. The time variability of the waves can be assessed from the *in situ* data, while the satellite data allows the spatial variability of the waves to be studied.

A second reason for looking at the spatial variability of H_s is to enable the data acquired along the satellite tracks to be interpolated between the tracks. This may be of importance for practical applications, such as ship routeing, and also for the assimilation of the data into wave models, where the grid points in the model will not in general coincide with the satellite measurements. It also allows the appropriate spacing of the grid points for wave modelling to be determined so that the resolution of the model is matched to the scales of variability of the wavefield.

A few studies (Challenor 1983, Challenor *et al.* 1986, Monaldo 1988) of the spatial variability of H_s have been carried out. These show a large-scale (of the order of 1000 km) variation in H_s, associated with wave generation by storms over the oceans, as well as shorter-scale variability. One of the more interestingg observations made with Seasat data is that H_s can change very rapidly in the open ocean over very short distances. Queffeulou (1983) and Guymer and Srokosz (1986) document examples where H_s changes by several metres over 50–150 km. As most wave models have a grid spacing of the order of 200 km, or greater, these relatively rapid changes in H_s would not be resolved. Similarly, data from an *in situ* measuring device on the other side of such a H_s 'jump' from the satellite measurement, would not be useful for validation purposes (usually *in situ* data taken within about 100 km of the satellite track are considered good for validation purposes).

The limitation on the use of altimeter data for this type of study is the size of the footprint (typically about 7 km) which means that only variations on scales greater than that of the footprint can be examined. Nevertheless, this still enables interesting studies to be performed. Here we have only discussed the spatial variability of H_s, but clearly the variability of the other altimeter wave parameters (discussed in § 2 above) could be examined in a similar way.

3.3 *Joint distributions of H_s and wind speed*

In designing offshore structures engineers are interested in the loads imposed by both waves and winds on the structures. A question area of interest is the probability of obtaining a given waveheight and wind speed simultaneously. There is little suitable *in situ* data available to be able to provide an answer to this question. As noted in § 2 the altimeter provides coincident measurements of H_s and wind speed U_{10} (here we reference the wind speed to a height 10 m above the sea surface), so it would seem to give us a means of answering the question.

In order to test this we have carried out a preliminary study of the joint distribution of H_s and U_{10} using Seasat data covering the whole period of the mission. Five 10° squares in different regions of the world's oceans were chosen for study (see figure 3 and the table for their locations). All the Seasat altimeter measurements of H_s and U_{10} within ecach 10° square, acquired during the mission, were binned (into 0·5 m H_s and 1 m s^{-1} U_{10} bins) and the resulting two-dimensional histogram was normalised to produce a joint probability density function (pdf) for H_s and U_{10}. The H_s-U_{10} pdfs were contoured and the results obtained for the five 10° squares are shown in figures 4 to 8.

Before discussing the results shown in the figures some comment needs to be made on how the wind speed U_{10} was obtained from the altimeter return. As noted in § 2 the derivation of wind speed from σ_0 is a subject of some controversy (see Chelton and McCabe 1985). Here we used two different algorithms to convert σ_0 to U_{10}, those of

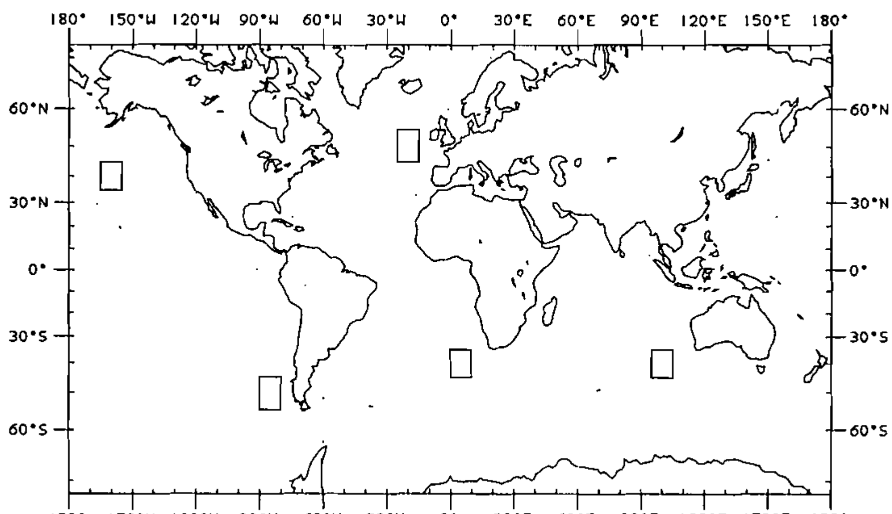

Figure 3. World map showing the locations of the five 10° squares for which Seasat data was analysed in the study of joint H_s–U_{10} pdfs.

The geographical locations of the 10° squares and the means and standard deviations of H_s and U_{10} obtained for those squares, using the Barrick (1974), and Chelton and McCabe (1985) algorithms.

	No. of data points	Latitude	Longitude	H_s(m) mean	H_s(m) s.d.	U_{10}(m s^{-1}) mean (Barrick)	U_{10}(m s^{-1}) s.d. (Barrick)	U_{10}(m s^{-1}) mean (Chelton and McCabe)	U_{10}(m s^{-1}) s.d. (Chelton and McCabe)
N. Atlantic	5388	45–55° N	15–25° W	2·6	1·4	7·4	3·1	7·0	3·6
N. Pacific	5963	35–45° N	135–145° W	1·9	0·9	5·7	2·5	5·0	2·9
S. Atlantic	5094	35–45° S	0–10° E	4·0	1·4	10·6	5·1	10·5	5·4
S. Pacific	5406	45–55° S	80–90° W	4·8	1·3	12·0	4·3	12·2	4·8
S. Ocean	4674	35–45° S	95–105° E	4·9	1·5	10·3	4·4	10·5	5·1

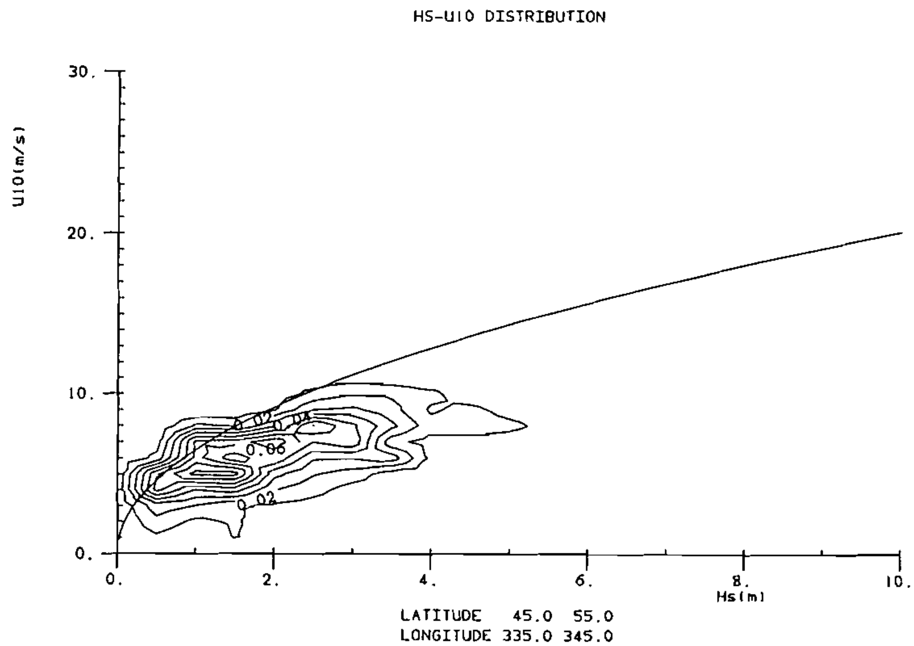

Figure 4. The contoured joint H_s-U_{10} pdf for the North Atlantic 10° square, estimated from Seasat data. The curve is the Pierson–Moskowitz relationship between H_s and U_{10} for a "fully developed sea".

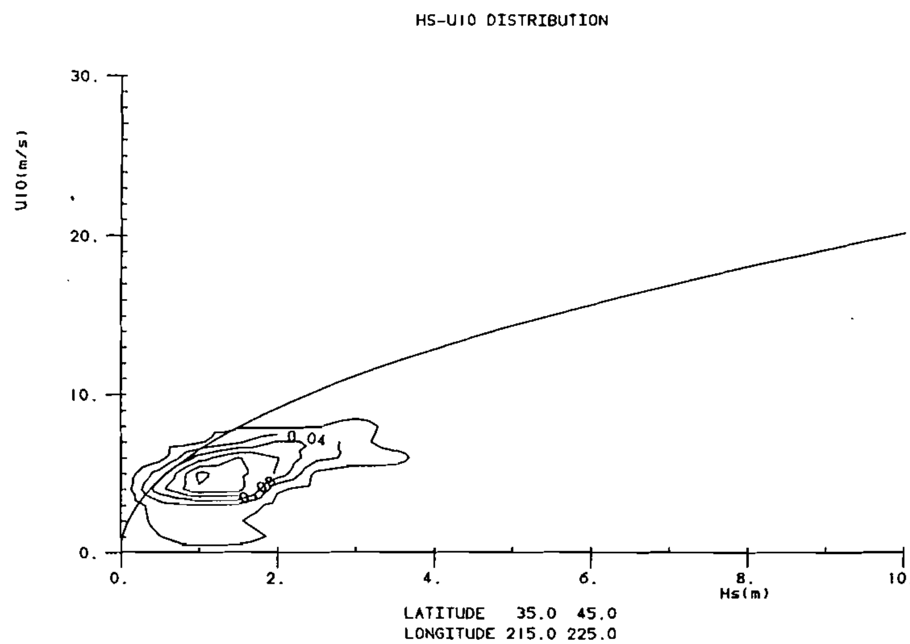

Figure 5. The contoured joint H_s-U_{10} pdf for the North Pacific 10° square, estimated from Seasat data. The curve is the Pierson–Moskowitz relationship between H_s and U_{10} for a "fully developed sea".

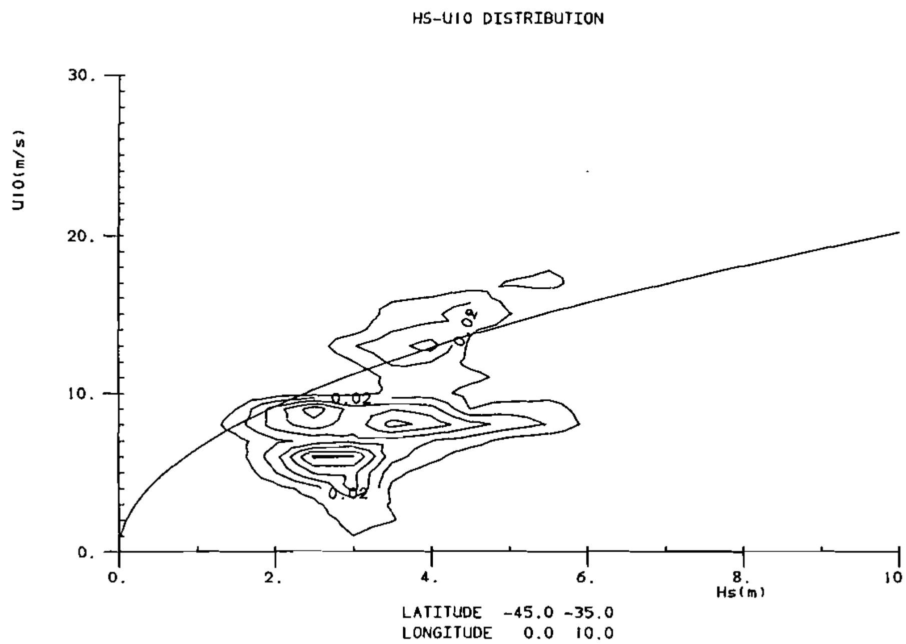

Figure 6. The contoured joint H_s-U_{10} pdf for the South Atlantic 10° square, estimated from Seasat data. The curve is the Pierson–Moskowitz relationship between H_s and U_{10} for a "fully developed sea".

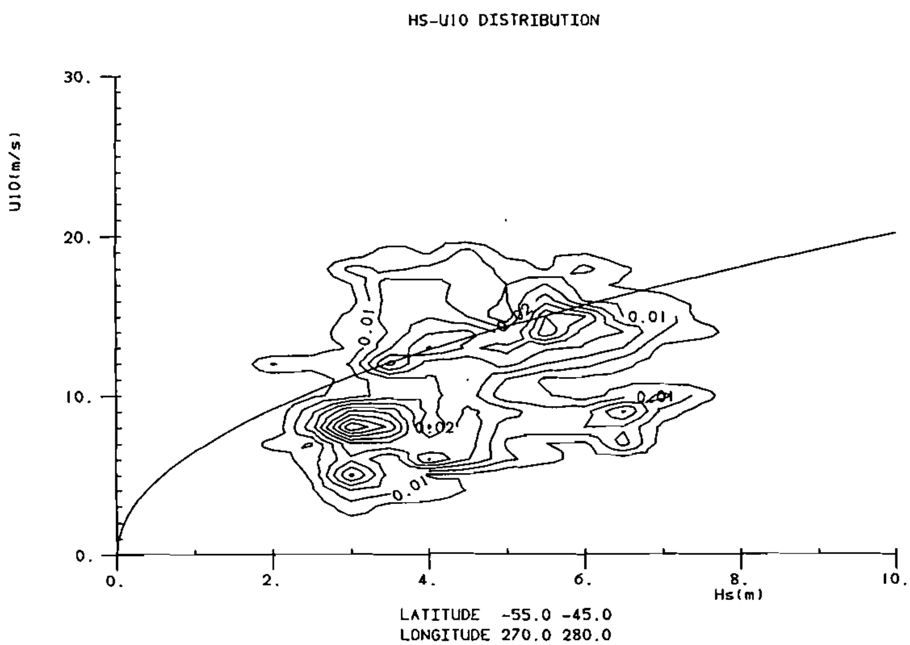

Figure 7. The contoured joint H_s-U_{10} pdf for the South Pacific 10° square, estimated from Seasat data. The curve is the Pierson–Moskowitz relationship between H_s and U_{10} for a "fully developed sea".

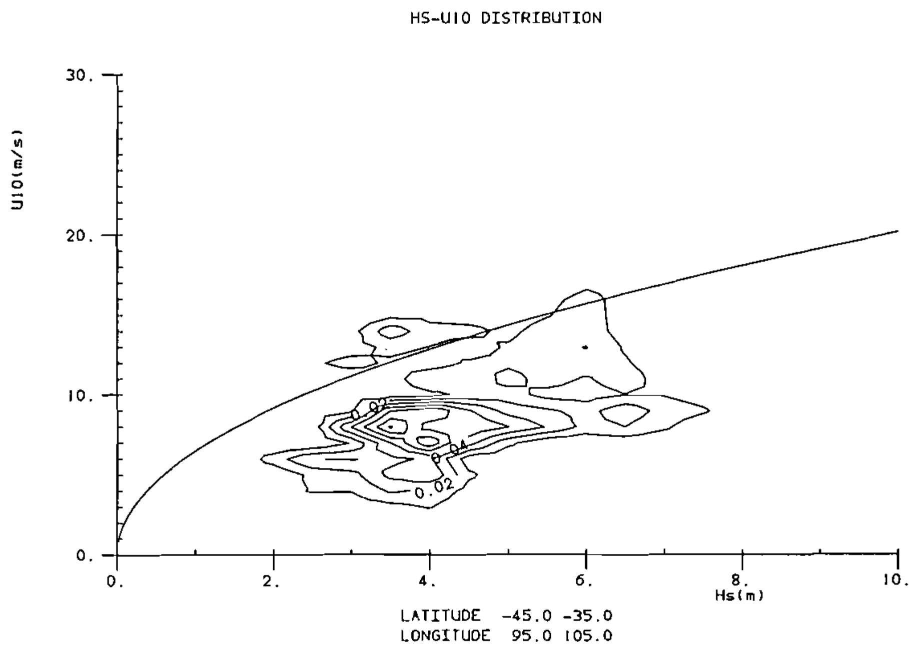

Figure 8. The contoured joint H_s-U_{10} pdf for the Southern Ocean 10° square, estimated from Seasat data. The curve is the Pierson–Moskowitz relationship between H_s and U_{10} for a "fully developed sea".

Barrick (1974) and Chelton and McCabe (1985) (we also applied an AGC to σ_0 correction as suggested in this paper). The table shows that the two algorithms lead to slightly different results for the mean and standard deviation of U_{10}. Comparison of the contoured joint pdfs of H_s and U_{10}, obtained using the two algorithms, show qualitative similarity but some quantitative differences, which reflect the differences evident in the results in the table. In figures 4 to 8 we show results based on the Barrick algorithm.

From figures 4 to 8 and the table it can be seen that during the period of the Seasat mission (27 June to 10 October 1978) the wind speeds and waveheights in the southern hemisphere were larger than those in the northern hemisphere. This, of course, reflects the fact that the data were obtained during the northern hemisphere summer and the southern hemisphere winter. Superimposed on figures 4 to 8 is the Pierson–Moskowitz relationship between H_s and U_{10} for a so-called "fully developed sea" (Pierson and Moskowitz 1964). This curve gives a limit on the height of the waves that can be generated by the local wind. As most of the data have waveheights higher than this limit, for any given wind speed, this shows the predominance of swell (non-locally generated waves) in the measurements. Figures 9 and 10 give the marginal pdfs of H_s and U_{10}, again illustrating the differences between the northern and southern hemisphere data.

From figure 10 it can be seen that there is some double-peaked structure in the southern hemisphere wind speed pdf. This is present whether the Barrick or the Chelton and McCabe algorithm is used and is therefore probably not an artefact of the algorithms. Chelton and McCabe (1985) noted that the original algorithm used for estimating U_{10} from Seasat data was flawed and led to a spurious multiple-peaked

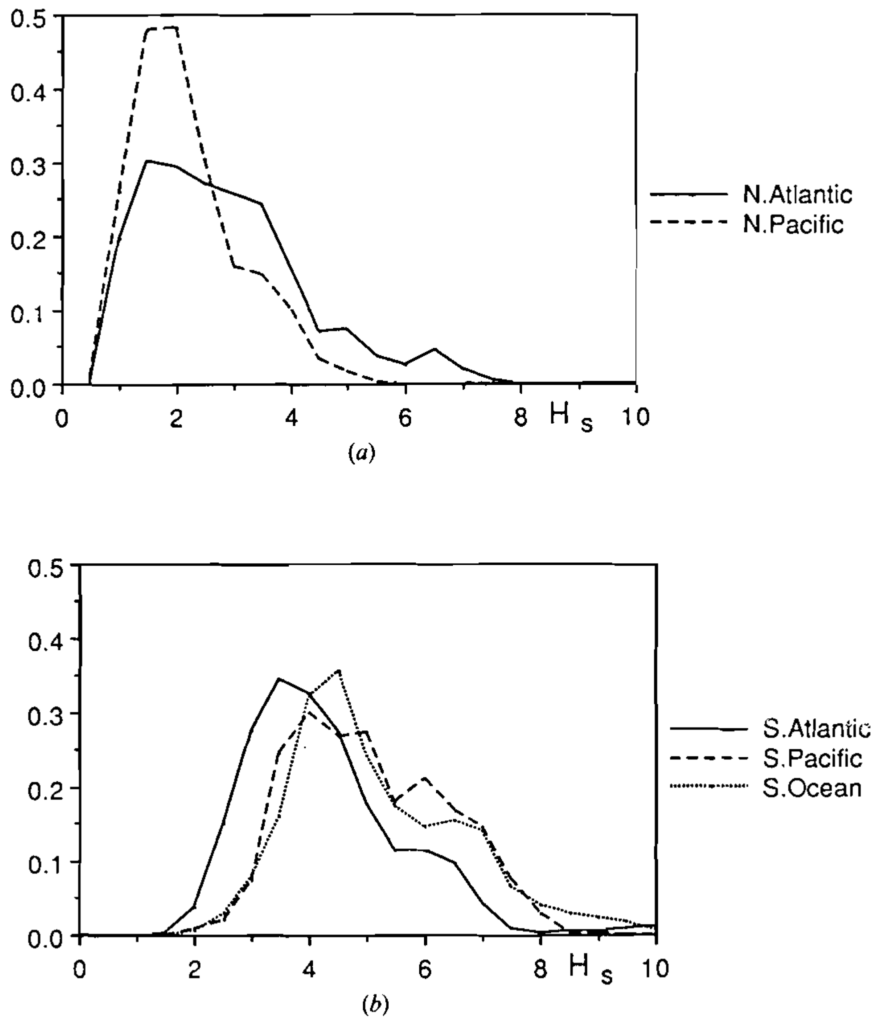

Figure 9. The marginal distributions (pdfs) of H_s (metres) estimated from Seasat data for the regions in (*a*) the North Atlantic and North Pacific and in (*b*) the South Atlantic, South Pacific and Southern Ocean.

structure in the U_{10} pdf. However, we have followed their suggestions for resolving that problem and must therefore conclude that the double-peaked structure is real (unless there is some further as yet unrecognized problem with the data). We have at present no explanation for this observation.

The results on the joint pdfs of H_s and U_{10} presented here must be regarded as preliminary, due to the limitations of the Seasat data. It is hoped to carry out a more comprehensive study using Geosat data and at the same time improve on the derivation of wind speed from altimeter data to obtain more accurate results. There is also the possibility of calculating joint distributions of other parameters (for example, H_s and λ), but these are probably of less practical interest and applicability than those for H_s and U_{10}.

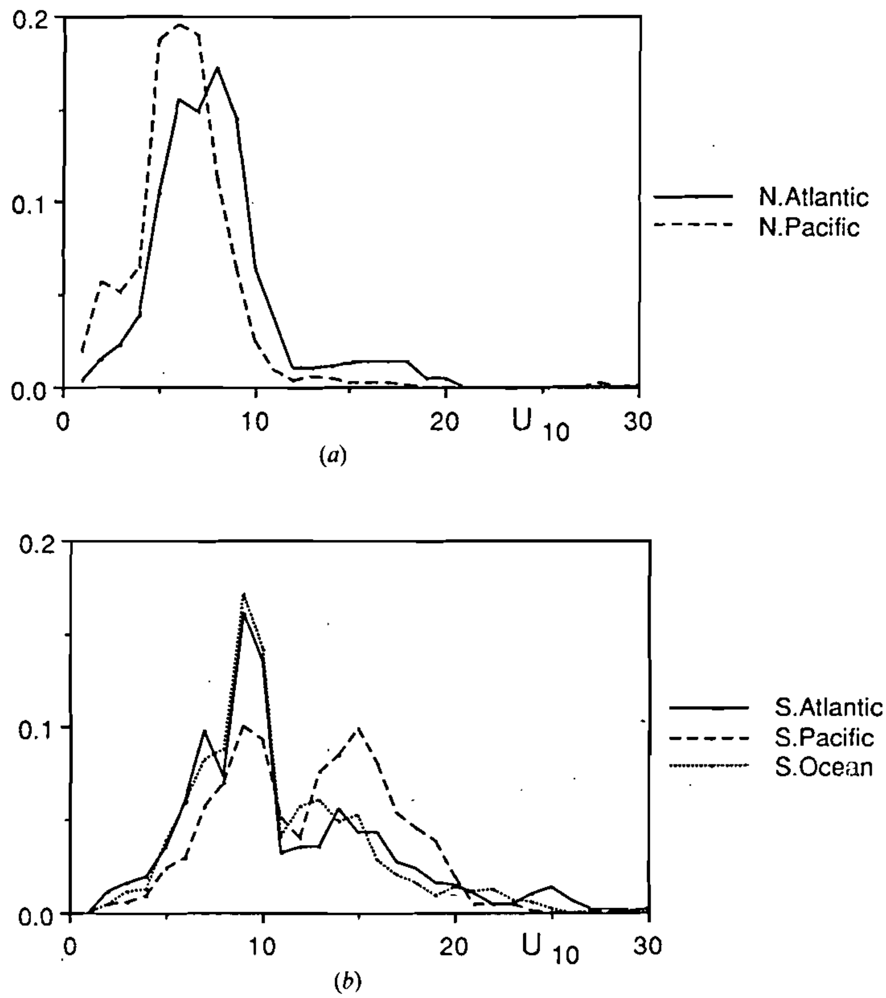

Figure 10. The marginal distributions (pdfs) of U_{10} (metres per second) estimated from Seasat data, using the Barrick algorithm, for regions in (*a*) the North Atlantic and North Pacific and in (*b*) the South Atlantic, South Pacific and Southern Ocean.

3.4. *Other studies*

To conclude this look at waves studied using radar altimeter data we will mention two other types of studies that have been carried out. Parsons (1979) and Mognard (1983) have attempted to use altimeter data to estimate the swell component of the waves. They do this by using the wind speed measurement to estimate the maximum H_s that could be generated by that wind and noting that the remaining energy in the wavefield must be due to swell.

The second type of study that has been pursued is the assimilation of altimeter wave data into wave models. Thomas (1988) has published results which show the potential for improving wave forecasts by the assimilation of H_s data into the British Meteorological Office wave model. In America, Esteva (1988) has shown the beneficial impact of assimilating Seasat H_s data on hindcasts made using the NOAA wave model. The real-time assimilation of ERS-1 altimeter data into wave models is being studied at both the British Meteorological Office and the European Centre for Medium-Range Weather Forecasting, with a view to implementing such a scheme when the ERS-1 satellite is launched in 1990.

4. Conclusions

Having examined the work carried out in the ten years since the launch of Seasat, what conclusions can be drawn about the use of radar altimeter data for wave studies? It is clear that there is potential for further studies in the future. In particular, the present availability of Geosat data, and the future availability of ERS-1 and Topex/Poseidon data in the 1990s, will provide a stimulus to further research. However, there is still a need to increase our understanding of the basic physics of the interaction of the wind and the waves and the consequent effect on the radar return from the sea surface. This is necessary in order to obtain accurate estimates of geophysical parameters from the radar return. Eventually, as our understanding develops and more data becomes available it will become possible to make routine use of the data for practical purposes, such as offshore engineering design, prediction of extreme conditions and wave forecasting (the latter by assimilating the data into wave models). All these applications, and the fundamental studies that will undergird them, require large-scale (global), long-term (decadal), continuous measurements of waves. This need will be met by the data that is being obtained now from Geosat, and that will be obtained over the next few years from ERS-1 and Topex/Poeseidon. The future of wave studies using radar altimeter data appears assured.

Acknowledgments

We are grateful to the Algorithm Development Facility of the U.K. Earth Observation Data Centre for help in preparing Seasat data for use in the study of joint distributions of H_s and U_{10}.

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