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ON THE DETECTION OF UNDERWATER BOTTOM TOPOGRAPHY BY IMAGING RADARS

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ABSTRACT

A simple theoretical model explaining basic properties of radar imaging of underwater bottom topography in tidal channels is presented. The surface roughness modulation is described by weak hydrodynamic interaction theory in the relaxation time approximation. In contrast to previous theories on short wave modulation by long ocean waves, a different approximation has to be used when describing short wave modulation by tidal flow over underwater bottom topography. The modulation depth is in this case proportional to the relaxation time of the Bragg waves. The large modulation of radar reflectivity observed in SEASAT-SAR imagery of sand banks in the Southern Bight of the North Sea can be explained by assuming that the relaxation time of 34 cm Bragg waves is of the order of 30-40 seconds.

1. INTRODUCTION

It has been known for more than 10 years that real aperture radar (RAR) imagery taken over sea areas with strong tidal currents (tidal channels) sometimes shows features that seem to be related to underwater bottom topography (de Loor and van Hulten, 1978; de Loor, 1981; McLeish et. al., 1981). The same phenomenon has also been observed in synthetic aperture radar (SAR) imagery obtained by the SEASAT satellite (Fu and Holt, 1982; Lodge, 1983; Kenyon, 1983; Lyzenga et. al., 1983). The sea-floor topography (bathymetry) causing these radar signatures sometimes lies tens of meters below the sea surface. An example of such imagery is the SEASAT-SAR image shown in Fig. 1.

FIG.1: Digitally processed SEASAT-SAR image of the Southern Bight of the North Sea from orbit 762 (Aug. 19, 1978, 6:46 UT) with frame center at 51°19'N, 1°52'E. The land area in lower left hand corner is the English coast near Ramsgate. The V-shaped feature in the center are the sand banks South Falls and Sandettie. South Falls is about 30 km long, 600-800 m broad and rises from the sea floor of a depth of about 30-40 m to within 7 m of the sea surface.

The imaging of underwater bottom topography by real or synthetic aperture radar seems not to be understandable at first sight, because the penetration depth of the electromagnetic waves emitted by the radar into sea water is only of the order of millimeters to centimeters. Therefore, surface effects related to the sea-floor topography must be responsible for the radar imaging.

Since radar signatures of bathymetry are only observed when strong (tidal) currents are present, we are led to the hypothesis that these radar signatures are caused by surface current variations associated with underwater bottom topography. Variable surface currents modify the short-scale surface roughness, and this roughness variation is sensed by the radar. Since most imaging radars operate at incidence angles between 20 and 70 degrees the radar backscattering is dominated by Bragg scattering (Valenzuela, 1978). Consequentely, cross-section modulation results from the modulation of the spectral energy of the Bragg waves.

Cross section modulation determines the image intensity or grey tone level variations in real aperture radar imagery. However, in case of synthetic aperture radar, in addition to this amplitude modulation, also phase modulation or velocity bunching contributes to the imaging (Larson et.al. 1976; Alpers and Rufenach, 1979; Alpers et.al. 1981). In most cases of SAR imaging of bathymetry, however, the contribution of velocity bunching to the imaging mechanism is small (Alpers and Hennings, 1984). Therefore we will discuss in this paper only amplitude or cross section modulation.

2. BATHYMETRY - CURRENT INTERACTION

The interaction of a 3-dimensional time-variable current field with 3-dimensional underwater bottom topography (bathymetry) can sometimes be a very complex process which does not allow a simple mathematical description. Nevertheless, in this paper we make the simplest possible assumption that the current flow over the bathymetry is laminar, free of any vertical current shear and only weakly time dependent. Furthermore, we assume that the tidal velocity component $U_{\perp}$ normal to the direction of the underwater bank obeys the continuity equation

$$U(x_\perp)d(x_\perp) = \text{const} \tag{1a}$$

and that the parallel component $U_{\parallel}$ remains constant,

$$U_{\parallel} = \text{const} \tag{1b}$$

Here $d(x_\perp)$ denotes a depth profile along a line perpendicular to the ridge direction $(x_\perp)$.

Current measurements on and off the sand bank South Falls by Venn and Olier (1983) in the Southern Bight of the North Sea (north-eastern approach to the English Channel), where the tidal current flows across the bank at an oblique angle, have confirmed that equations (1a) and (1b) are acceptable first order approximations. (However, deviations from these simple relations are observed, and a more refined theory should account for them).

3. CURRENT - BRAGG WAVE INTERACTION

Since the variation of the surface current due to interaction with bathymetry has space and time scales that, in general, are small compared with the space and time scales of the Bragg waves, the current -Bragg wave interaction can be described by a WKB (Wentzel - Kramers - Brillouin)-type interaction theory. In this theory the transport equation, which describes the variation of the spectral energy density of short waves in a slowly varying current field, is the action balance or radiation balance equation (Hasselmann et al., 1973; Keller and Wright, 1975; Alpers and Hasselmann, 1978; Wright, 1978). This equation reads

$$\frac{dN}{dt} = \mathcal{L}[N] = \left[\frac{\partial}{\partial t} + \dot{\underline{x}}\frac{\partial}{\partial \underline{x}} \quad \dot{\underline{k}}\frac{\partial}{\partial \underline{k}}\right] N = S(\underline{x},\underline{k},t) \tag{2}$$

where

$$N(\underline{x},\underline{k},t) = E(\underline{x},\underline{k},t) / \omega' \tag{3}$$

is the action spectrum, $E(x,k,t)$ the wave spectrum, ω' the intrinsic frequency of the wave in a reference system which is locally at rest, $\underline{x}$ the space variable, $\underline{k}$ the wavenumber and $S(\underline{x},\underline{k},t)$ a source function. The waves propagate along trajectories in 4-dimensional phase space which are given by the ray equations

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$$\dot{\underline{x}} = \frac{\partial \omega}{\partial \underline{k}} \quad , \quad \dot{\underline{k}} = - \frac{\partial \omega}{\partial \underline{x}} \tag{4}$$

where

$$\omega(\underline{x},\underline{k},t) = \omega'(\underline{k}) + \underline{k} \cdot \underline{U}(\underline{x},t) \tag{5}$$

denotes the wave frequency in the moving medium with variable velocity $\underline{U}(\underline{x},t)$.

We assume that the variable surface current only leads to small deviations of the action density from equilibrium. Therefore, we write the action density N and the surface current U as sums of a constant equilibrium term and a time dependent perturbation term

$$N(\underline{x},\underline{k},t) = N_0(\underline{k}) + \delta N(\underline{x},\underline{k},t) \tag{6}$$

$$\underline{U}(\underline{x},t) = \underline{U}_0 + \delta \underline{U}(\underline{x},t) \tag{7}$$

Furthermore, we approximate the source term S by a diagonal operator

$$S(\underline{x},\underline{k},t) = \mu \, \delta N(\underline{x},\underline{k},t) \tag{8}$$

where μ is a parameter with dimension $(\text{time})^{-1}$. μ is called relaxation rate and $\tau_r = \mu^{-1}$ the relaxation time. Physically, τ_r is a system constant describing the response of the wave system to small deviations from equilibrium caused by surface current variations. It is determined by the combined effect of wind excitation, energy transfer to other waves due to conservative resonant wave - wave interaction, and energy loss due to dissipative processes like wave breaking. No measurements of the relaxation time in the open ocean exist. However, from theory we expect that τ_r is of the order of 10 to 100 wave periods. Applied to SEASAT-SAR Bragg waves, which have a wavelength of 34 cm and a wave period of 0.47 s, this means that τ_r should lie in the range between 4.7 and 47 seconds. In this paper we consider τ_r (or μ) as a free parameter. Inserting eqns. (6), (7) and (8) into eqn. (2) and only keeping first order terms yields

$$\left[\frac{\partial}{\partial t} + (\underline{c}_g + \underline{U}_0) \frac{\partial}{\partial \underline{x}} + \mu \right] \delta N = \underline{k} \cdot \frac{\partial \underline{U}}{\partial \underline{x}} \cdot \frac{\partial N_0}{\partial \underline{k}} \tag{9}$$

The time scales of the three terms on the left hand side are given by the local time T, the acvection time τ_a, and the relaxation time τ_r. The local time is of the order of the period of the semi-diurnal tide divided by 2π, which is $12.5/2\pi$ h $\approx$ 2h, and the advection time is given by

$$\tau_a = \left| (\underline{c}_g + \underline{U}_o) \cdot \underline{K} \right|^{-1} \qquad (10)$$

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where $\underline{K}$ is the wavenumber of the bottom topography.

For tidal flow over large-scale underwater bottom topography, such as sand banks, both, the local time T and the advection time τ_a, are usually much large than the relaxation time τ_r. Therefore the first two terms on the left hand side of eqn. (9) can be neglected in comparison with the third term.

This approximation is applicable for discribing short surface wave modulation by tidal flow over bathymetry, but it is completely different from the approximation applicable for discribing short wave modulation by long surface waves. In this case, the local time is given by Ω^{-1}, where Ω is the radian frequency of the long surface wave. Ω^{-1} is typically of the order of 1 or 2 seconds, which is small compared to the relaxation time τ_r. Therefore, the first term in eqn. (9) is dominant in this case.

We now insert eqn. (4) and the dispersion relation for short surface waves into eqn. (9) and consider only those waves which travel towards or away from the antenna look direction (Bragg waves). If we define the projection of the antenna axis onto the horizontal plane as x-direction, then in case of tidal flow over large-scale bottom topography, we obtain the result

$$\frac{\delta E}{E_0} = - \frac{4+\gamma}{\mu} \frac{\partial U_x(\underline{x})}{\partial y} \qquad (11)$$

and in case of long surface waves

$$\frac{\delta E}{E_0} = - \frac{4+\gamma}{\Omega} \frac{\partial U_x(\underline{x})}{\partial x} \qquad (12)$$

Here γ denotes the ratio between the group and phase velocity of the short waves. For gravity waves we obtain $\gamma = 0.5$, and for capillary waves $\gamma = 1.5$. In deriving eqns. (11) and (12) we have assumed that the spectral energy of the short waves is proportional to $|\underline{U}|^{-4}$.

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Note, that according to Bragg scattering theory,

$$\frac{\delta\sigma}{\sigma_0} = \frac{\delta E}{E_0} \qquad (13)$$

where $\frac{\delta\sigma}{\sigma_0}$ denotes the relative cross section modulation.

Comparison of eqns. (11) and (12) shows that in the first case the modulation is proportional to $\mu^{-1} = \tau_r$ and in the second case proportional to $\Omega^{-1} = (2\pi)^{-1} T_W$, where T_W is the period of the long surface wave. We will show below that for the SEASAT-SAR Bragg waves τ_r is of the order of 30 to 40 seconds. This means that for the same current gradient, the modulation by tidal flow over bathymetry is a factor 15-40 larger. On the other hand, the surface current gradient generated by tidal flow over bathymetry is usually much smaller than the surface current gradient generated by the orbital motion of long surface wave (typically one order of magnitude smaller). Therefore we obtain the net result, that the cross section modulation ("modulation depth") associated with tidal flow over sand banks is of the same order of magnitude, or even larger than the modulation associated with long surface waves.

Physically, this result is a consequence of the fact that in the case of tidal flow over bathymetry the weak interaction can act sufficiently long on the short wave system. The limit is given by the relaxation time. In case of short wave modulation by long surface waves, the short waves have not sufficient time to build up a strong modulation because their dwell-time in flow regions with positive or negative velocity gradients is determined by the period of the long surface waves.

If the partial time derivative and advection terms in eqn. (9) are neglected, then eqn. (9) states, that the relaxation of the action density perturbation is balanced by "straining" exerted on the wave system by the spatially varying current $\underline{U}$. The right hand side of eqn. (9) represents a refraction term. It originates from the fact that the spatially variable current $\underline{U}$ refracts the short waves, i.e., changes their wavenumber. A wavenumber change causes a local perturbation of the equilibrium action (or energy) density spectrum of the short waves, because the short wave spectrum varies as a function of wavenumber. Thus, the steeper N is, as a function of wavenumber, the larger is the modulation. Note also that the modulation is proportional to the current gradient.

In those cases where the advection time τ_a is not small compared to the relaxation time τ_r, the second term on the left hand side of eqn. (9) has to be retained. The inclusion of this term results in the addition of a low-pass filter to the imaging process. This is discussed in detail in Alpers and Hennings (1984).

4. RELATIONSHIP BETWEEN BATHYMETRY AND CROSS SECTION MODULATION

By combining the results of sections 2 and 3 we can derive an expression relating the cross section modulation to the large scale bottom topography and the tidal current field. Inserting eqns. (1a),(1b) and (11) into eqn. (11) yields

$$\frac{\delta\sigma}{\sigma_0} = \frac{4+\gamma}{\mu} |U_0| d_0 \cos\varphi \cos^2\phi \frac{\text{grad}_\perp d}{d^2} \qquad (14)$$

Here ϕ denotes the angle between the flight and the sand bank direction, φ the angle between the $x_\parallel$- and the (undisturbed) flow direction, $\text{grad}_\perp d$ the gradient of the depth profile in direction perpendicular to the bank crest, d_0 the water depth outside the bank area, and $|\underline{U}_0|$ the modulus of the undisturbed current velocity.

Inspection of eqn. (14) shows that radar signatures of sand banks always have double sign, which means that the radar image is composed of image elements having both enhanced and reduced grey levels relative to the local mean. The sign of the modulation is such that increased radar reflectivity occurs always on the downstream side, and reduced radar reflecting on the upstream side of the sand bank.

Furthermore, eqn. (14) shows that the modulation depth increases with tidal velocity and decreases with water depth. The modulation pattern is not correlated with the depth profile d, but with $d^{-2}\,\text{grad}_\perp d$.

In particular, eqn. (14) predicts that the cross section modulation vanishes when
- the tidal velocity is zero ($|\underline{U}_0| = 0$)

- the bank crest is aligned parallel to the current direction ($\psi = 90°$)
- the band crest is aligned parallel to the radar look direction ($\phi=90°$)

(However, we do not expect that the last statement fully holds, because it is a consequence of the fact that our hydrodynamic interaction model does not include the interaction between short surface waves travelling in different directions)

According to eqn. (14) the modulation depth depends on the relaxation time $\tau_r = \mu^{-1}$. We expect that, to first order, τ_r is independent of wind direction. However, it is likely that τ_r decreases with wind speed.

In order to obtain an estimate of the relaxation time we analyzed digitally processed SEASAT-SAR imagery and made several image intensity scans across sand banks in the Southern Bight of the North Sea (north eastern approach to the English Channel) and compared them with bathymetry. One scan was made across South Falls along the line shown in Fig.1. The arrow indicates the direction of the tidal flow at the time of the overflight. The tidal current velocity was 0.60 m/s and the wind was blowing from 135^{o}N at 4 m/s. This image was processed on the MDA (McDonald Dettweiler and Associates) digital SEASAT processor by the Deutsche Forschungs- und Versuchsanstalt für Luft- und Raumfahrt at Oberpfaffenhofen, FRG. The resolution is 25m * 25m (4 looks).

Our analysis shows that we can explain the measured modulation depth if we assume a relaxation time of the order of 30-40 seconds for the SEASAT-SAR Bragg waves (for a wind speed of 4 m/s). This value corresponds to 60-80 wave periods of 34 cm waves.

5. CONCLUSIONS

We have confronted the predictions of our imaging theory with existing experimental data. It seems that the predictions are largely confirmed. However, it should be stressed that we consider this theory only to be a first-order theory.

Finally, we want to suggest the use of radar imagery of bathymetry in tidal channels as a practical means for measuring the relaxation time of short surface waves in the open ocean.

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REFERENCES

Alpers, W., and K. Hasselmann, 1978: The two-frequency microwave technique for measuring ocean wave spectra from an airplane or satellite, Appendix B, Boundary-Layer Metorol., vol. 13, 215-230.

Alpers, W. R., and C. L. Rufenach, 1979: The effect of orbital motions on synthetic aperature radar imagery of ocean waves, IEEE Trans. Antennas Propagat., AP-27 (5), 685-690.

Alpers, W. R., D. B. Ross, and C. L. Rufenach, 1981: On the detectability of ocean surface waves by real and synthetic aperature radar, J. Geophys. Res. 86, 6481-6498.

Alpers, W., and I. Hennings, 1984: A theory of the imaging mechanism of underwater bottom topography by real and synthetic aperture radar, submitted to J. Geophys. Res..

de Loor, G. P., 1981: The observation of tidal patterns, currents and bathymetry with SLAR imagery of the sea, IEEE, Geoscience and Electronics, (USA), OE-6, 124-129.

de Loor, G. P. and H. W. Brunsveld van Hulten, 1978: Microwave measurements over the North Sea, Boundary-Layer Meteorology, 13, 113-131.

Fu, L.-L., and B. Holt, 1982: SEASAT views oceans and sea ice with synthetic-aperature radar, JPL publication 81-120, Pasadena, California.

Hasselmann, K, et.al., 1973: Measurements of wind-wave growth and swell decay during the joint North sea wave Project (JONSWAP), Deut. Hydrogr. Z., 8 (suppl. A.).

Keller, W. C. and J. W. Wright, 1975: Microwave scattering and straining of wind generated waves, Radio Sci., vol. 10, 139-147.

Kenyon, N. H., 1983: Tidal current bedform investigated by SEASAT, in T. D. Allen, Ed., Satellite Microwave Remote Sensing, Ellis Horwood, Series in Marine Science, Chichester, England, Chapter 16.

Larsen, T. R., L. I. Moskowitz, and J. W. Wright, 1976: A note on SAR imagery of the ocean, IEEE Trans. Antennas Propagat. vol. AP-24, 393-394.

Lodge, D. W. S., 1983: Surface expressions of bathymetry on SEASAT synthetic aperture radar images, Int. J. Remote Sensing, Vol. 4, 629-653.

Lyzenga, D. R., R. A. Shuchman, E. S. Kasischke and G. A. Meadows, 1983: Modeling of bottom-related surface patterns imaged by synthetic aperture radar, 1983 International Geoscience and Remote Sensing Symposium (IGARSS 83), San Francisco, Aug. 31 - Sept. 2, 1983, Digest, Vol. 2, 7.1 - 7.10.

McLeish, W., D. J. P. Swift, R. B. Long, and D. Ross, 1981: Ocean surface patterns above sea-floor bedforms as recorded by radar, Southern Bight of North Sea, Marine Geology, 43, M1-M8.

Valenzuela, C. R., 1978: Theories for the interaction of electromagnetic and ocean waves - a review, Boundary-Layer Meteorol., 13, 61-85.

Wright, J. W., 1978: Detection of ocean waves by microwave radar; the short gravity-capillary waves, Boundary-Layer Met., 13, 101-102.