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Threshold of sediment movement by open ocean waves: observations*

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Abstract—Between 8 and 25 March, 1973, a sequence of bottom photographs, sediment samples, bottom current measurements, and bottom pressure measurements was recorded continuously at several locations on the Washington continental shelf. Concurrently, wave conditions were recorded at 35-m depth on the summit of Cobb Seamount 465 km west of the shelf experiment.

During the sampling period the bottom turbidity at 167 m on the shelf varied significantly in response to the magnitude of wave activity, and these data are used to evaluate the threshold wave conditions necessary to initiate the observed sediment movement.

The bottom oscillatory velocity, orbital diameter of wave motion, and the mean sediment size occurring at the shelf location are compared. For unconsolidated silt-sized sediment the expression

$$\frac{\rho U^2_{1/10}}{(\rho_s - \rho)gD} = 0{\cdot}13\left(\frac{d_0}{D}\right)^{1/2} ,$$

where ρ_s and ρ are the sediment and fluid densities, d_0 is the horizontal excursion of water particles at the sea floor, D is the grain diameter, and $U_{1/10}$ the higher 10% of bottom oscillatory velocities, describes the threshold of grain motion as observed.

INTRODUCTION

SEDIMENT resuspension by oscillatory motions is an important process on continental shelves. In this paper, we report on the correlation between two independent observations that relate bottom turbidity with wave motion. The two experiments were in March, 1973—one on the continental shelf off the State of Washington and the other on Cobb Seamount 465 km seaward of the shelf break and approximately at the same latitude as the shelf experiment. A sequence of bottom photographs was taken on the shelf at depths of 38, 75, and 167 m. On the summit of the seamount, 35 m deep, records of the surface wave field were made thrice daily. The correlation between the two sets of data provides insight into the critical spectral conditions that cause sediment resuspension.

Laboratory studies have verified that the SHIELDS (1936) criterion is an important grouping of variables for describing the threshold of sediment movement. Under steady flow conditions sediment moves when θ reaches sufficiently large values, where

$$\theta = \frac{\tau_0}{(\rho_s - \rho)gD} . \tag{1}$$

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In (1) τ_0 is the boundary shear stress, $\rho_s - \rho$ is the density difference between the sediment grains and water, g is the acceleration of gravity, and D is the sediment diameter. It has been suggested that initial grain motion under oscillatory flow conditions (KOMAR and MILLER, 1973) may be predicted by a relation

$$\theta = \frac{\rho U^2}{(\rho_s - \rho)gD} = f\left(\frac{d_0}{D}\right) , \tag{2}$$

in which U is the maximum bottom oscillatory velocity and d_0 is the horizontal excursion of a fluid parcel at the sediment–water interface. Unfortunately there are many physical models, all of which could support the above functional dependence. KOMAR and MILLER (1973) suggest that $f(d_0/D)$ in (2) is the one-half power. They show data to support this based on laboratory experiments of BAGNOLD (1946) and MANOHAR (1955) and suggest the following relationship

$$\frac{\rho U^2}{(\rho_s - \rho)gD} = K\left(\frac{d_0}{D}\right)^{\frac{1}{2}} . \tag{3}$$

Komar and Miller used the laboratory experiments cited above to evaluate the magnitude of K in (3) at which sediment motion occurred. The results were different for the two laboratory experiments (0·21 and 0·39), and they were averaged to arrive at an intermediate value of 0·30. Our data cover a limited but different range of (d_0/D) than the experiments of Bagnold and Manohar. However, we will apply our field data to (3) in order to evaluate a critical value of K at which sediment motion occurs. Thus, the analyses in this paper are an extrapolation of the above laboratory results applied to field data. Consistent estimates of critical threshold conditions might also be obtained from other functional relationships than those given in (3).

A wave of amplitude a at the surface in water of depth h will be associated with a bottom horizontal velocity maximum of

$$u = a\sigma/\cosh \kappa h, \tag{4}$$

where σ is the wave frequency and κ is the wave number. This may be rewritten

$$u = \frac{\sigma d_0}{2} , \tag{5}$$

where d_0 is as in equation (2). Substitution of (5) into (3) yields

$$\frac{\rho u^{3/2} \sigma^{\frac{1}{2}}}{1{\cdot}41(\rho_s - \rho)gD^{\frac{1}{2}}} = K , \tag{6}$$

where the field value of K is unknown but is expected to have some critical magnitude at which sediment motion occurs. Equation (6) is for a single wave component. In the ocean a spectrum of wave velocities is present. We shall make the plausible assumption that at large depths the velocities are due to swell and that the swell is uni-directional.

The procedure in this paper is to evaluate K in (3) using the wave conditions observed on Cobb Seamount as a data base for the shelf experiment and to evaluate the critical value at which the bottom photographs reveal significant sediment suspension. To our knowledge these data represent the only oceanic field measurements (outside of surf zone) of this kind and hence provide a unique opportunity to investigate the relationship between surface waves and sediment movement.

INSTRUMENTATION AND DATA

The shelf experiment

The experiment carried out on the Washington shelf during March, 1973, was designed to measure bottom current, pressure fluctuation, and to photograph the boundary at three stations located perpendicular to the Washington coast line. The instruments were at depths of 38, 75, and 167 m and were operational between 8 and 31 March, 1973. Station locations are shown in Fig. 1.

Measurements were made at each station with an instrumented tripod that freely descends from the sea surface. It can remain on the sea floor for periods of up to 30 days and (a) continuously measure speed and direction 100 cm off the bed with a Savonius rotor current meter and direction vane; (b) continuously measure differential pressure 2 m off the bed to estimate the tides and pressure fluctuations from surface-wave motion; and (c) take a photograph of the sea floor each 30 min. All data are recorded internally and retrieved with the sensing elements and tripod after the predetermined sampling interval. A complete description of this instrumentation system is described by STERNBERG, MORRISON and TRIMBLE (1973).

Bottom photographs. Bottom photographs are taken with a vertically oriented 35-mm camera located within the tripod-pressure housing approximately 1·8 m off the seabed. The camera field of view is approximately 0·5 m on each side. Three black

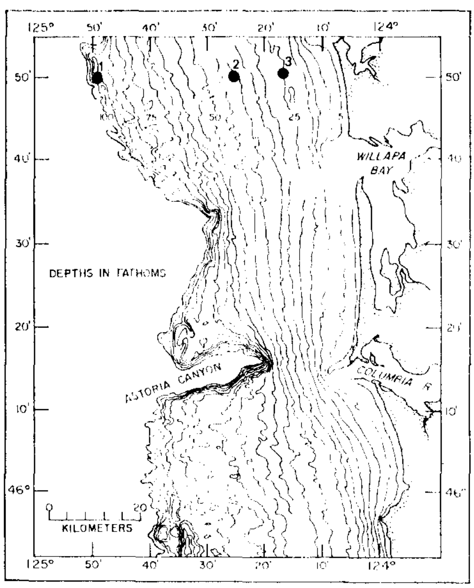

Fig. 1. Sampling locations for the March, 1973, shelf experiment.

and white reproductions of the color bottom photographs taken before, during, and after a period of sediment suspension are shown in Figs. 2(a)–(c).

Photographs obtained from the three shelf locations showed various degrees of turbidity. At Stas. 2 and 3 the level of turbidity was so high that the seabed was not clearly discernible at any time during the sampling period. At Sta. 1, near the shelf break, turbidity levels were generally low with specific periods of high turbidity due to sediment suspensions. Because of the time variability of bottom turbidity at Sta. 1, it is possible to correlate the observed turbidity maxima with other hydrodynamic parameters in an effort to determine the threshold conditions causing sediment movement. Because of the consistently high turbidity at Stas. 2 and 3, these data cannot be used for this purpose; thus, this paper addresses itself to the conditions occurring at Sta. 1.

To obtain a quantitative estimate of the turbidity level at Sta. 1, the color films were analyzed with a photo-resistive type of densitometer using the white compass face as a standard background (Fig. 2a). The results of this analysis are shown in Fig. 3(a), which illustrates the relative variation in turbidity during the total sampling period. This record is not calibrated with respect to particle concentrations; however, it does define the background level of relative turbidity and the periods of significant sediment suspensions.

Pressure fluctuations. The output from the differential pressure transducer on the instrumented tripod indicates the average pressure fluctuation encountered at the

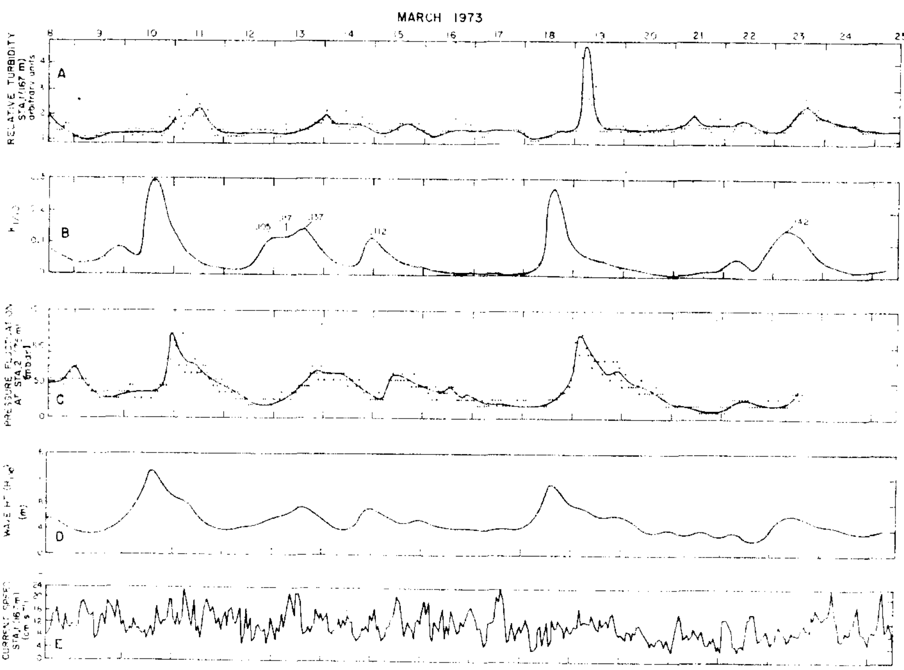

Fig. 3. Composite of data from the shelf experiment and concurrent open ocean wave measurements. A, C, and E are data from the Washington continental shelf; D summarizes the average height of the higher 10% of waves measured at Cobb Seamount; B represents the magnitude of $K_{1/10}$ calculated from Cobb Seamount wave data. Travel time for waves propagating from the seamount to the shelf stations is 9 to 10 h.

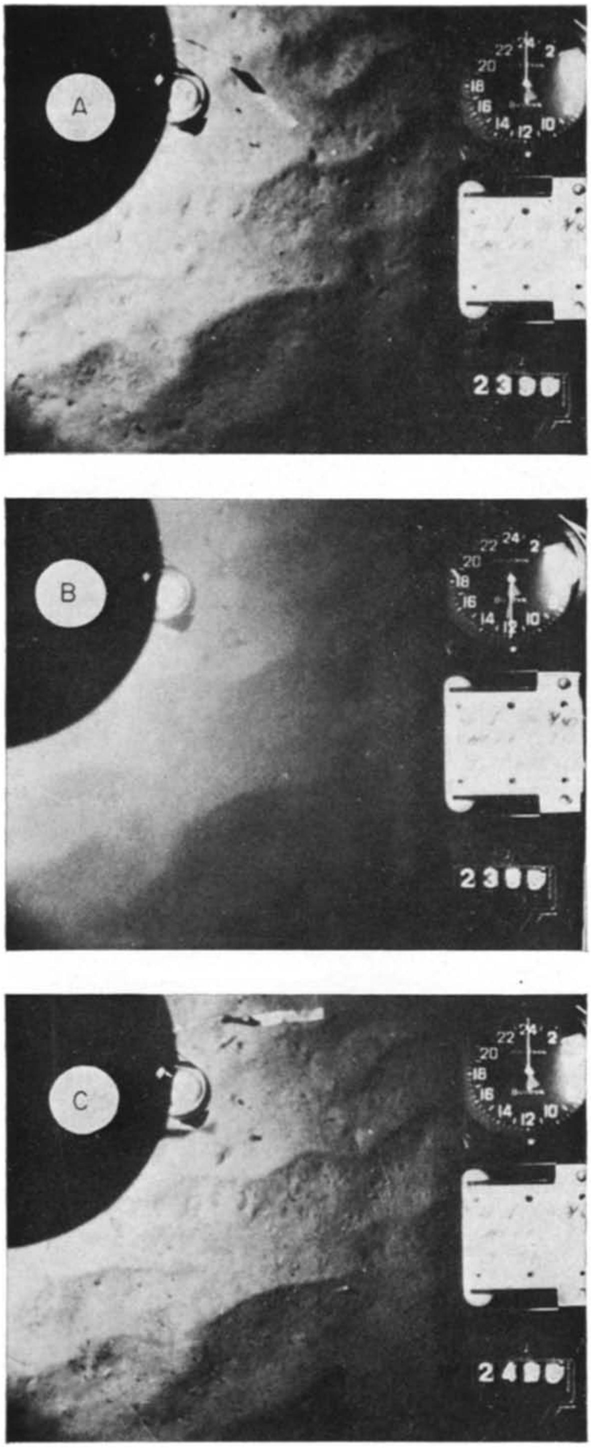

Fig. 2. Bottom photographs taken before, during, and after a period of significant wave motion: 2A taken at 2400 on 23 March; 2B at 1200 on 23 March; 2C at 2400 on 25 March.

bottom; however, the Rustrak recording system does not provide sufficient time resolution to determine the frequency characteristics of the ambient wave motion. The differential pressure record occurring at Sta. 2 during the sampling interval is shown in Fig. 3(c). (The transducer at Sta. 1 was inoperative.) From these data it is seen that pressure fluctuations due to surface waves varied significantly during the sampling period with maximum fluctuations occurring on 10 to 11 March and 19 March and secondary peaks occurring on 14, 15, and 23 March.

Bottom currents. Bottom currents measured at Sta. 1 are shown in Fig. 3(e). The curve represents 1-h averages of the current speed at 1 m above the seabed. Current directions are not included in Fig. 3 because the objective of this paper is to determine the threshold conditions causing sediment motion regardless of flow direction.

The mean bottom current speed varied between 3·2 and 23·0 cm s^{-1}, with an overall mean value of 10·9 cm s^{-1}. A strong tidal component was observed throughout the record. The current metering system was not sufficient to resolve wind-wave oscillations.

Bottom sediment. Gravity core samples were collected at each station while the instruments were being deployed. Surface sediment texture at Sta. 1 is characterized as a bimodal size distribution with prominent modes at 4φ (0·063 mm) and 6φ (0·016 mm) (Fig. 4). Size analysis of the bottom sediment shows a mean grain size of 5·89φ (0·0156 mm) and a sorting coefficient of 3·02φ units.

The seamount experiment

During the time of the shelf experiment described above, a wave recorder was in operation on Cobb Seamount approximately 465 km west of the Washington coast (46°46·4′N, 130°48·8′W). Cobb Seamount has a relatively conical shape with a base diameter of approximately 31 km. It rises over 2750 m above the sea floor to within 35 m of the sea surface. The pinnacle of the seamount stands as a single spire rising 50 m above a surrounding terrace. The side slopes of the pinnacle are quite steep (45°) and the top is flat with a horizontal extent of about 325 m.

On 23 September, 1972, the wave gauge constructed at the University of Washington was placed on the summit of Cobb Seamount for the purpose of recording wave conditions during the following winter. The wave-induced pressure fluctuations at

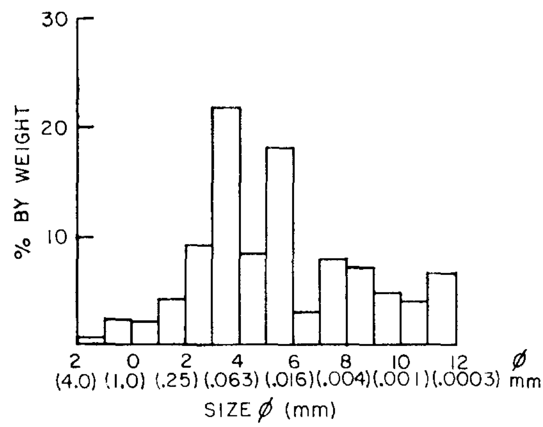

Fig. 4. Size-frequency histogram for bottom sediment from Sta. 1 on the continental shelf.

the seamount were measured by means of a vibrating wire pressure-transducer (vibrotron) that converts the fluctuating pressure filed into a frequency modulated electronic signal. The system will resolve pressure changes equivalent to 7 mm of water-level change. This is approximately the magnitude of pressure fluctuation generated at a depth of 35 m by a 5-s wave of about 1-m amplitude, or a 20-s wave of 8-mm amplitude. This instrument was housed in a pressure housing rated at 200 m that was affixed to a massive triangular frame. It was lowered from the stern of R.V. *Oceanographer* to the seamount summit at a depth of 35 m.

The instrument operated on a 30-min duty cycle each 8 h until 17 April, 1973, at which time the tape supply was exhausted. Recovery was made with the assistance of divers from Bellevue Community College and the Coast Guard Cutter *Iris*. The instrument was brought to the surface on 25 May, 1973.

The basic timing for the system was with a 1·2-mHz quartz crystal oscillator. During the eight months of remote operation of the system, the cumulative time error was 33 s. This error spread over the individual 30-min sample intervals is completely negligible. The data were recorded at intervals of 0·8789 s directly on 1-mil, 1/2-in., 7-track magnetic tape at 200 bits per inch. Only one parity error was found and this occurred near the end of the tape supply. Most probably this was due to a change in tape tension. Thus only one record of 615 was lost.

Wave data. The pressure records collected at Cobb Seamount showed significant wave activity occurring during the month of March. These records have been analyzed by LARSEN and FENTON (1974) to give spectral characteristics of surface waves during each measurement period that the instrument was operating.

The pressure records yielded direct information of the surface wave spectra and enabled reconstruction of the surface wave field. The reconstructed wave field is band-limited depending on the high frequency energy in the signal. In most cases, the surface elevations contain all of the component energy at periods greater than 7 s.

The height of the highest 1/10th of the waves ($H_{1/10}$) for the month of March is shown in Fig. 3(d). The expected value of the wave heights was also estimated using the variance of the wave field and assuming Gaussian statistics. The values for the statistics such as $H_{1/10}$ are about 8% greater when calculated statistically. This discrepancy is in agreement with that found by others in comparing pressure and direct techniques for measuring wave fields.

DISCUSSION

Although sediment movement occurred frequently during the March shelf experiment, mean bottom currents did not exceed the threshold of grain motion for any significant periods of time. From Fig. 5 it is observed that the minimum current (at $z = 100$ cm) required to cause erosion of sand and unconsolidated silt-sized* material approximately 26 cm s^{-1}. Mean bottom currents during the sampling period showed relatively strong fluctuations on a tidal frequency but never exceeded 23·0 cm s^{-1}. Possibly the threshold of motion was exceeded by the mean bottom currents for short periods on 11, 17, 24, and 25 March; however, the duration of these events, if they occurred, was limited.

*Because of the relatively high mobility of the shelf sediments and the frequency with which they are placed in suspension, the bottom sediment is considered to be unconsolidated, in which case the erosion curve in Fig. 5(a) would tend to follow the lower dashed line (SUNDBORG, 1956).

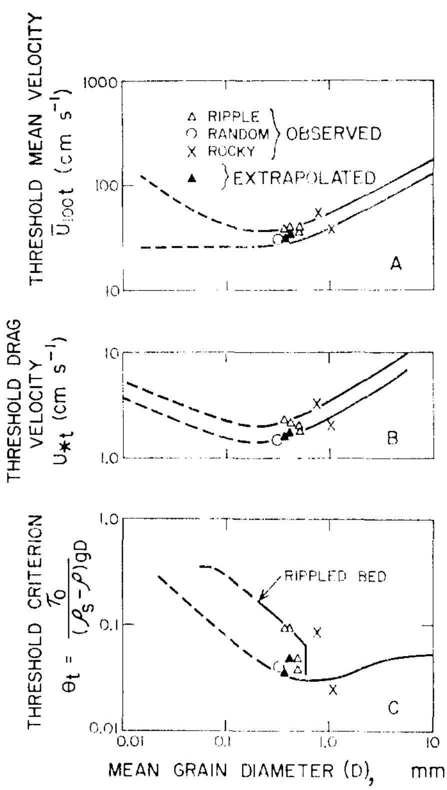

Fig. 5. Threshold of grain motion as related to: A, mean velocity 1 m from the bed ($\bar{U}_{100}$); B, drag velocity (U^*); C, relative stress criterion (θ). Figure taken from STERNBERG, 1971.

A strong correlation exists between the pressure fluctuations observed at Sta. 2 and concurrent observations of bottom turbidity at Sta. 1 (Fig. 3a and 3c). This correspondence leads to the conclusion that the high turbidity levels are the result of sediment suspension caused by wind-wave-induced bottom oscillations. Similarly, the primary justification for using wave measurements from Cobb Seamount, 465 km from the shelf location of interest, is the excellent correlation between the wave climate at the seamount (Fig. 3d) and the differential pressure fluctuations that simultaneously occurred at Sta. 2 (Fig. 3c). The similarities between these records indicate that the same wave conditions occurring over Cobb Seamount are responsible for the pressure fluctuations occurring 9 to 10 h later on the Washington shelf. Estimates of the decay of the larger waves travelling from the seamount to the outer edge of the shelf indicate that changes in significant wave height over the decay distance would be less than 10% (KINSMAN, 1965); thus, the wave statistics from the Cobb wave recorder are being used to estimate directly wave conditions at Sta. 1.

Equation (6) is a dimensionless grouping of the parameters considered to be important in causing sediment resuspension (KOMAR and MILLER, 1973). The pressure records taken on Cobb Seamount enable us to evaluate (6) for the depths at which bottom photographs were obtained. The procedure is to convert the pressure record that was obtained to an equivalent pressure record that could have been taken at

167 m. That is, if $P_c(\sigma)$ is the complex Fourier coefficient of pressure record at Cobb Seamount, then

$$P_s(\sigma) = P_c(\sigma)\frac{\cosh \kappa_s d_s}{\cosh \kappa_c d_c} \quad , \tag{7}$$

where the wave numbers are calculated from the frequency equations

$$\begin{aligned} \sigma^2 &= g\kappa_c \tanh \kappa_c d_c \\ &= g\kappa_s \tanh \kappa_s d_s \quad , \end{aligned} \tag{8}$$

and d_c and d_s refer to the depths of Cobb Seamount and to the bottom at the site of the experiment.

Assuming that all the waves at the depth of 167 m are propagating in the same direction, we can predict the comple × Fourier coefficients of the horizontal velocity field

$$U_s(\sigma) = \frac{\sigma}{g\rho} P_s(\sigma)\frac{\cosh (\kappa_s d_s)}{\sinh (\kappa_s d_s)} \quad . \tag{9}$$

Figure 6 shows $K(\sigma)$ as defined in (6) versus frequency. Estimates of $u(\sigma)$ used in (6) were obtained from (9) as

$$u(\sigma) = [U(\sigma)U^*(\sigma)]^{\frac{1}{2}} \quad . \tag{10}$$

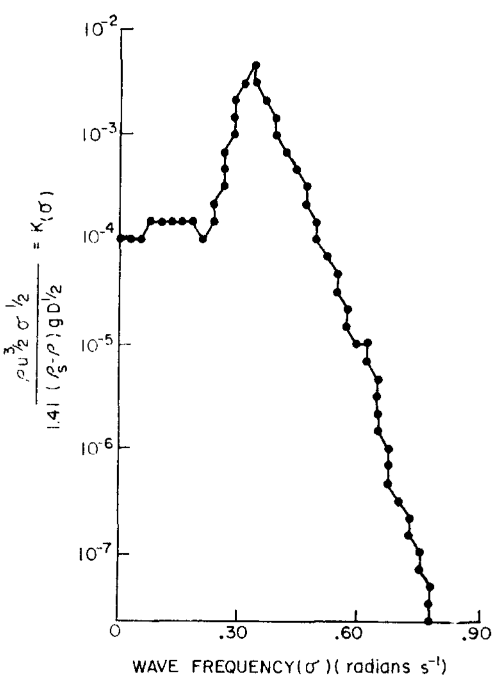

Fig. 6. Plot of $K(\sigma)$ versus wave frequency (σ).

The spectrum of $K(\sigma)$ is seen to peak at a frequency σ_0 which represents the frequency of waves contributing the most to the resuspension of sediment. The value of $K(\sigma)$ calculated for each frequency band, however, does not in itself represent the critical value for sediment resuspension because it does not reflect the total energy present.

To estimate the effect of the total wave field, we chose the following procedure: equation (10) estimates the contribution to the square root of the variance of the velocity spectra at the frequency σ. The total energy in the velocity field was estimated by integrating $u^2(\sigma)$ over all frequencies. Then we assume Gaussian statistics for the velocity field and estimate that

$$U_{1/10} = 3{\cdot}6(E^{\frac{1}{2}}), \tag{11}$$

where E is the energy variance of the velocity field (KINSMAN, 1965). The value of $U_{1/10}$ as estimated by (11) is in agreement with that derived directly by back-transforming the velocity field.

For this study we have used $U_{1/10}$ in determining the critical value of K because we felt that (1) $U_{1/10}$ represents the greater fraction of wave energy accounting for sediment suspension; (2) these oscillations occur frequently enough (16 to 20 waves per 30-min record) to keep material in suspension for the length of time between photographs; (3) these values would compare more favorably with the monochromatic wave conditions used in the laboratory studies considered by KOMAR and MILLER (1973). Similarly, it is felt that most of the energy is at σ_0 and it is also used in (6). This assumption is not too crucial because of the square-root dependence on the frequency (in contrast to the 3/2 dependence on $U_{1/10}$). The value of K obtained in this way is called $K_{1/10}$; however, one could obtain a $K_{1/3}$ or K average or some other value if desired.

The magnitude of $K_{1/10}$ for each wave record obtained during the March experiment is shown in Fig. 3(b). During the sampling period, the magnitude $K_{1/10}$ varied from 0·005 and 0·29 and shows a strong correlation with the wave motion and the magnitude of bottom turbidity.

Because mass concentration of suspended material was not measured, the decision regarding the threshold of grain motion is rather subjective. The background or noise level associated with the turbidity measurements varies between 1 and 1·35 (Fig. 3a). Bottom photographs taken during the times of low turbidity are extremely clear (Figs. 2a and 2c). Periods of major sediment suspension occurred on 11 and 19 March. At these times, the seabed was completely obscured in the photographs and the threshold conditions were far exceeded. It is impossible to correlate the leading edge of the high turbidity peaks on 11 and 19 March with the actual wave conditions causing suspension because of the extremely rapid increase in turbidity (especially on 19 March) and the fact that the wave conditions on Cobb Seamount cannot be exactly phased to the turbidity maxima occurring 465 km away. Phase differences occur because of the long travel time to the shelf (9 to 10 h), and also the wave conditions on Cobb Seamount were measured only once every 8 h.

An alternative for estimating threshold wave conditions is to explore the secondary maxima in both the turbidity and wave records. Strong wave action occurred at 0000 on 14 March, 0800 on 15 March, and 1600 on 23 March (Fig. 3a and e). Bottom turbidity in excess of background noise occurred at these times also. The numerical value of $K_{1/10}$ at these critical times is displayed above each data point in Fig. 3(b).

During the 13 March peak, $K_{1/10}$ varied from 0·105 to 0·137, and increased bottom turbidity occurred somewhere between the times of the 0·117 and 0·137 values. On 15 March, $K_{1/10}$ reached a peak value of 0·112, and the turbidity level showed a slight increase above background. It should be noted that relatively high tidal currents also occurred at this time on 15 March, so although the threshold bottom conditions were slightly exceeded, waves and mean bottom currents may have augmented one another to cause this movement.

On 23 March the peak value of $K_{1/10}$ was 0·142 and a significant increase in turbidity also occurred. The bottom photograph taken at 1630 on 23 March shows that the bottom is visible but only the gross features are discernible (Fig. 2b). This is in contrast to the high resolution photographs obtained before (Fig. 2a) and after (Fig. 2c) threshold was exceeded. The primary bed configuration was altered only slightly by the 24-h period of sediment motion. The ripple pattern was modified to the extent that the crests were not realigned, but they were made more distinct or peaked. The sediment placed in suspension by the wave action probably consisted of the 6φ mode, whereas the basic bed configuration includes the sandy fractions of the bed material which were not moved.

From the comparison of $K_{1/10}$ and bottom turbidity on 13, 15, and 23 March, it appears that the threshold of motion was for values of K between approximately 0·12 and 0·14. Comparison of bottom photographs during these periods shows significant levels of suspension and slight degrees of bed modification. This value of $K_{1/10}$ (0·13) represents about one-half the value suggested by KOMAR and MILLER (1973). Some representative values of the maximum horizontal oscillatory velocities associated with the wave conditions, in which $K_{1/10}$ varies from 0·12 to 0·14, are shown in Table 1. From this it is seen that the critical value of $U_{1/10}$ that causes the initiation of grain motion is on the order of 7·2 cm s^{-1}, which is somewhat less than the 10 cm s^{-1} suggested by KOMAR, NEUDECK and KULM (1972), yet quite close, considering the differences between the laboratory and field experiments.

CONCLUSIONS

Observations of surface wave motion, bottom turbidity, and sediment texture are used to estimate the threshold oscillatory conditions that initiate sediment movement

Table 1. Some representative values of the maximum horizontal oscillatory velocities associated with the wave conditions.

Date	Record Number	$K_{1/10}$	$U_{1/3}$	$U_{1/10}$	U_{max}
10 Nov 72	175	0.127	5.51	7.01	9.91
26 Nov 72	224	0.124	5.45	6.93	9.80
25 Dec 72	311	0.125	5.39	6.85	9.69
11 Mar 73	540	0.146	5.99	7.61	10.80
14 Mar 73	548	0.137	5.82	7.40	10.50
24 Mar 73	577	0.142	5.91	7.52	10.60

on the Washington continental shelf. These data suggest that the expression

$$\frac{\rho U^2_{1/10}}{(\rho_s - \rho)gD} = 0{\cdot}13\left(\frac{d_0}{D}\right)^{\frac{1}{2}}$$

can be used to predict the threshold of grain motion due to oscillatory flows. This relationship is thought to apply for grain diameters less than 0·05 cm where the flow in the boundary layer is still laminar. For the range of conditions measured, the magnitude of $U_{1/10}$ at which threshold occurred was approximately 7·2 cm s^{-1}.

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