QUANTIFYING THE DISTRIBUTION AND TRANSPORT OF PELAGIC SEDIMENTS ON YOUNG ABYSSAL HILLS

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Abstract. The post-depositional transport of pelagic sediments on rough seafloor is approximated as a nonlinear diffusive process. The topographic slope distribution of the sediment-water interface depends primarily on the apparent sediment diffusivity $\kappa$, the average sediment load $L$, and the root-mean-square (rms) height $H$. These parameters have been determined for the ARSRP corridor on the west flank of the Mid-Atlantic Ridge by fitting theoretical models to the slope distribution observed from center-beam Hydrosweep data. The average value of the apparent diffusivity is $\kappa = 0.13 \pm 0.04 \text{ m}^2/\text{yr}$. $H$ varies between 128 and 285 m, with a mean of 230 \pm 7 m. $L$ is estimated to be as thin as 10 m on 5-Ma crust near the ridge crest, thickening to values greater than 50 m at 10–15 Ma and then decreasing to less than 30 m on 25-Ma crust in the western portion of the corridor. The decrease in sediment load with age may be associated with a shallower carbonate compensation depth prior to the late Miocene.

INTRODUCTION

Pelagic sedimentation modifies the seafloor on the flanks of the mid-ocean ridges, smoothing and eventually burying the abyssal hills generated by seafloor spreading. Thin pelagic sediments are notoriously difficult to study, however, because the scattering by rough abyssal-hill topography obscures the wide-beam acoustic records obtained from surface ships. This report outlines a new methodology for quantifying the sediment distribution and transport that utilizes only bathymetric data routinely collected by narrow-beam echosounders.

Considered in detail, the processes that govern deep-sea sedimentation, such as the generation of small-scale turbidity currents, the interactions of bottom currents with pre-existing topography, and bioturbation, are complex. But averaged over geological time, they act to transport sediments from topographic highs to topographic lows at all scales [e.g. Marks, 1981]. We therefore model sedimentation as a simple diffusive process in which the lateral flux of sediments is proportional to the local topographic gradient [Culling, 1960]. Diffusion models have been used to explain the evolution of scarp-like landforms [Hanks et al., 1984], as well as the stratigraphy found in foreland basins [Flemings and Jordan, 1989] and shallow marine settings [Kaufman et al., 1991]. To our knowledge, however, sedimentation in pelagic environments has not been addressed from this perspective.

FORWARD MODELING AND PARAMETER ESTIMATION

Let $h(x,t)$ be the height of a durable basement $b(x)$ covered by an erodible sediment layer of thickness $s(x,t)$, let $F(x,t)$ be the sedimentation rate, and let $\kappa$ be the diffusivity of lateral sediment transport. The sediment transport equation is

$$\frac{\partial s(x,t)}{\partial t} = \kappa V^2 h(x,t) + F(x,t) \tag{1}$$

where $V^2$ is the surface Lapacian. Eqn. (1) is solved for $h(x,t) = b(x) + s(x,t)$ subject to the constraint $s(x,t) \geq 0$. The inequality makes the problem highly non-linear. Complexity is also introduced by the structure of the basement topography $b(x)$, which has a fractal character.

We have obtained representations of the sediment-water interface by numerically solving (1) on a square grid using a lattice-gas algorithm. We start from basement topographies generated as realizations of the stochastic process of Goff and Jordan [1989, 1990]. The Goff-Jordan model represents the seafloor as a two-dimensional, spatially stationary, anisotropic Gaussian random field specified by five parameters: the root-mean-square (rms) height $H$, the characteristic length $s$, and a fractal dimension $D$. For the applications discussed here, we fixed the scale lengths and fractal dimension at values typical of the western flank of the Mid-Atlantic Ridge [Goff and Jordan, 1990; J. Goff, personal communication, 1992]; $\lambda_s = 22 \text{ km}$, $\lambda_d = 5.4 \text{ km}$, and $D = 2.2$.

The computations were performed on the Naval Research Laboratory's CM-2 Connection Machine using a 100 km \times 100 km horizontal grid and a grid point spacing of $\Delta x = 100 \text{ m}$. The flux rate $F$ was assumed to be constant in space and time, so that the total sediment load for a seafloor of age $T$ was $L = FT$. The diffusivity then scales as $\kappa = KL^2/T$, where $K$ is dimensionless. Synthetic topographies were calculated for $L \leq 100 \text{ m}$ for a wide range of $K$, $T$, and $H$.

The effect of varying the diffusivity is illustrated in Figure 1. For low values of $\kappa$, the sediments drape hillsides, while for higher values, they are transported more rapidly downslope, exposing peaks and creating flat sediment ponds. This behavior is reflected in the slope distribution, which is a sensitive indicator of deviations from the Gaussian character assumed for the basement topography [Shaw and Smith, 1990]. A profile of constant orientation sampled at a constant horizontal interval $u$ has a cumulative slope distribution function $S(\theta,u)$, defined to be the fraction of profile length with a slope angle, $\arctan[(h(x+u) - h(x))/u] < \theta$. $S(\theta,u)$ can be estimated directly from single-channel, narrow-beam echograms, and its shape is diagnostic of $H$, $L$, and $\kappa$ (Figure 2). For $K >> 1$ and $L/H >> 1$, the ponds are well defined, and $S(\theta,u)$ displays a sharp kink at a slope angle $\theta_0$ that separates the ponds from hillsides: i.e., $S(\theta,u)$ measures the fraction of the profile occupied by ponded sediments. This fraction decreases as $L$ decreases and the rate of this decrease is most rapid when $L/H$ is small. The slope of $S(\theta,u)$ at $\theta = \theta_0$ is sensitive to $L/H$, and the slope of $S(\theta,u)$ at $\theta >> \theta_0$ depends primarily on $H$. Hence, by least-squares fitting the theoretical values of $S(\theta,u)$ to the observed values, we can determine $H$, $L$, and $\kappa$. Although the results are conditional on the values assumed for $\lambda_s$, $\lambda_d$, and $D$, which we have fixed.
Fig. 1. Plan-view relief images (top) and vertical cross-sections (bottom) of two models of sedimented seafloors, showing the effect of different diffusivities: \( \kappa = 0.02 \text{ m}^2/\text{yr} \) (left), \( \kappa = 0.8 \text{ m}^2/\text{yr} \) (right). Models have the same age (\( T = 25 \text{ My} \)), sedimentation rate (\( P = 4 \text{ m/My} \)), sediment load (\( L = 100 \text{ m} \)) and basement topography (\( H = 250 \text{ m}, \lambda_y = 22 \text{ km}, \lambda_n = 5.4 \text{ km}, \xi_0 = 35^\circ, D = 2.2 \)). Lower diffusivity produces a morphology where sediments drape hillsides, while higher value yields more rapid downslope transport, exposing the peaks and creating flat sediment ponds. Dimensions of the images are 30 km x 30 km; colors change at elevations of -200, 200, and 600 m. Cross-sections are taken along the dashed lines and plotted with a vertical exaggeration of 7.5:1. Sections show basement topography, sediment-water interface, and two intermediate horizons corresponding to isochrons at 5 My (\( L = 20 \text{ m} \)) and 12.5 My (50 m).

in our inversions, numerical experiments show that \( S(\theta,\kappa) \) is only weakly dependent on these additional parameters.

APPLICATION TO THE ARSRP CORRIDOR

This procedure has been tested on a data set collected during Cruise 9208 of the R/V Maurice Ewing (14 Jul–18 Aug 92) as part of the Office of Naval Research (ONR) Acoustic Reverberation Special Research Project (ARSRP). The study area, shown in Figure 3, occupies 75,000 km\(^2\) on the west flank of the Mid-Atlantic Ridge north of the Kane Fracture Zone and comprises crust with ages ranging from 3 to 30 My. Track lines were run on a WSW-ENE trending grid subparallel to plate flow lines with a spacing that varied from 4-6 kilometers near the ridge crest to 8-9 kilometers in deeper water.

The primary data were the interpolated depth values logged each minute from the center beam of the Atlas Hydrosweep system. The average data-point spacing was ~300 m, which set the value of \( a \) in calculating the slope statistics. Based on a detailed examination of the swath maps, we grouped the track segments into 19 small areas (1000–3000 km\(^2\)) where the statistical character of the abyssal hills was more or less homogeneous. We avoided regions where the abyssal hills were interrupted by ridge-segment boundaries, which are typified by deeper bathymetry, thicker sediments, and inhomogeneous statistics. The track lengths in each area varied from 130 km to 360 km, averaging about 200 km. Mean crustal ages were determined from magnetic-anomaly data processed by M. Tivey (personal communication, 1993).

\( H, L, \) and \( \kappa \) were estimated for each area by fitting the center-beam slope distributions using a weighted least-squares algorithm; the results plotted on Figure 3. Figure 2 shows how the data for one area, centered at 26.0°N, 46.5°W, are matched by the theoretical model. The best estimates are \( H = 220 \pm 7 \text{ m}, L = 34 \pm 3 \text{ m}, \) and \( \kappa = 0.09 \pm 0.02 \text{ m}^2/\text{yr} \). The relative sizes of the standard errors (4%, 10%, and 20%, respectively) are typical of the other areas. In examining the covariances among parameters, we found that the uncertainties in \( H \) and \( L \) are positively correlated owing to the smoothing of the topography by the sediments; e.g., a large value of \( S(\theta,\kappa) \) may correspond either to thin sediments on low-
amplitude basement or to thick sediments on high-amplitude basement. The tradeoff between $L$ and $\kappa$ is small, however.

In Figure 3, $H$ varies from 188 m to 285 m, with a mean $\bar{H} = 230 \pm 7$ m. The variation is not large, but it does correlate with the ridge-segment boundary that passes through the center of the corridor. The average abyssal-hill amplitude in the seven areas north of the segment boundary is significantly higher ($\bar{H} = 255 \pm 12$ m) than in the twelve areas to the south ($\bar{H} = 214 \pm 5$ m), implying a persistent discontinuity in the ridge-crest constructional processes. In contrast, the variation in $L$ is primarily in an east-west direction, along the plate flow lines. In Figure 4, we plot the sediment load as a function of crustal age. It is thin on the young crust near the ridge crest, increasing westward to maximum thicknesses as high as 55 m on crust with ages of 10-15 Ma. The data out to 12 Ma indicate an apparent sedimentation rate of $\sim 3.4$ m/My.

On older seafloor, $L$ decreases with age, dropping to values as low as 20 m at 25 Ma. This inference agrees with the 3.5-kHz and single-channel seismic data collected during the ARSRP survey. A detailed discussion of the trends in Figure 4 is beyond the scope of this initial report, but we note that carbonates dominate the upper-ocean sediment flux in this region, and the non-monotonic character of $L$ may therefore be related to a deepening of the carbonate compensation depth (CCD) (B. Tucholke and G. Jaroslow, in preparation, 1993). Tucholke and Vogt [1979] concluded that the CCD in the central North Atlantic increased rapidly during the mid-Miocene, which corresponds to the sediment peak in Figure 4.

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**Fig. 2.** Plot of $S(\theta, u)$ for the area centered at 26.0°N, 46.5°W. Dots are data obtained from 170 km of Hydrosweep center-beam soundings. Heavy solid curve is the best-fitting theoretical model. Blue and green curves show the effect of varying $\kappa$, and red lines show the variation with $L$. Small circle on the $L = 100$ m curve denotes $\theta_a$, the apparent slope angle that separates sediment ponds from hillsides.

**Fig. 3.** Map of the ARSRP corridor, showing bathymetry, track lines, and model parameters estimated for the 19 areas analyzed in this report. Track lines used in slope distributions are highlighted. $L$ (in meters) is given by large number centered on each area; $H$ (in meters) and $\kappa$ (in m²/yr) listed as small numbers above and below. Color changes at water depths of 3200, 4000, and 4800 m. Region of deep bathymetry (light brown) trending WNW through the center of the corridor is a ridge-segment boundary separating higher abyssal hills to the north ($\bar{H} = 255 \pm 12$ m) from lower hills to the south ($\bar{H} = 214 \pm 5$ m). Sediment load $L$ ranges from 11 to 55 m and is a non-monotonic function of crustal age (see Figure 4); maximum values occur in the center of the corridor ($T = 10-15$ Ma). Significant differences in the apparent diffusivity are obtained from the data analysis ($\kappa = .04-.26$ m²/yr), but no spatial pattern is evident. Averaged over all 19 areas, $\bar{\kappa} = .13 \pm .014$ m²/yr.
Fig. 4. Estimates of sediment load L as a function of crustal age for the 19 areas analyzed in this report. Line corresponds to a constant sedimentation rate of 3.4 m/My. Shaded region delimits the interval of CCD deepening discussed by Tucholke and Vogt [1979]. We speculate that the non-monotonic behavior is due to CCD deepening prior to this time.

Taken at face value, the decrease in L between 15 and 25 Ma suggests that the CCD at the end of the Oligocene was even shallower than Tucholke and Vogt have estimated, and that the period over which it deepened included the early Miocene.

The individual areas show apparent diffusivities ranging from 0.04 to 0.26 m²/yr. Some variation in this parameter is evidently warranted by the data; attempts to fit the data with a constant value of \( \lambda \) led to unsatisfactory results, consistent with the error analysis, which demonstrates that the observed order-of-magnitude change should be resolvable. The physical significance of the differences, if any, is unclear. No pattern emerges from Figure 3. The mean values north and south of the segment boundary are statistically identical (0.15 \pm 0.020 vs. 0.12 \pm 0.019 m²/yr), as are those east and west of 47°W (0.13 \pm 0.017 vs. 0.13 \pm 0.023 m²/yr). The estimates of L and H, like those of L and H, appear to be spatially uncorrelated.

The mean value of the apparent diffusivity is determined with a small uncertainty: \( \lambda = 0.13 \pm 0.014 \) m²/yr. For comparison, we note that this value is about two orders of magnitude greater than that for compacted sediments in arid subaerial regions, where \( \lambda = 10^{-3} \) m²/yr [e.g., Hanks et al., 1984], but several orders of magnitude less rapid than in deltaic and near-shore marine settings, which yield values \( \sim 10^4 \) m²/yr [Flemings and Jordan, 1989].

**DISCUSSION**

Our methodology obtains good estimates of the sediment load, apparent diffusivity, and rms amplitude of the basement topography from single-channel, narrow-beam echosounding profiles in rough abyssal-hill terrains where standard sediment profiling techniques cannot be applied. Although the diffusion approximation used to invert for these parameters undoubtedly oversimplifies the processes that govern post-depositional transport, it represents a first step towards a dynamical model of sedimentation on young oceanic crust. Application of the statistical techniques presented here to dense, high-quality data sets like those being collected in the ARSRP corridor will lay the foundation for future improvements. In the short term, however, using these techniques to process existing narrow-beam data may be helpful in mapping the distribution and transport of sediments on the flanks of the mid-ocean ridges.

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**REFERENCES**


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