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Creep in pelagic sediments and potential for morphologic dating of marine fault scarps

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Abstract. Cores from the flanks of the Galapagos spreading centre show that the top 25 cm of pelagic sediments is highly bioturbated. *Kenyon and Turcotte* [1985] have proposed that bioturbation may cause down-slope creep of the sediment surface and bioturbation may therefore smooth sediment topography in a similar way to hillslope creep on land. Assuming this produces movement consistent with the diffusion transport model, sediment topography across scarp crests, digitised from Deep Tow profiler records, is modelled with a simple analytic solution to the diffusion equation and with a mean diffusion constant k of ~0.007 m^2/yr. This small value suggests creep rates are modest, for example, 2 cm/yr for a 30° slope, and is an upper bound since sediment avalanching and bottom currents probably also remove sediment from scarp crests. Although requiring further work, these results suggest that morphologic dating methods may be applicable to marine fault scarps and could be useful in relatively sediment-free areas, such as mid-ocean ridges, where it is commonly not possible to use datable offset seismic horizons.

Introduction

Developing a quantitative understanding of deep marine sediment transport is hindered by the inaccessibility of these areas, which largely prevents the kind of detailed observation of processes that is possible on land. In particular, creep caused by bioturbation is believed to be widespread in pelagic sediments [e.g., *Stow*, 1986] yet creep rates are difficult to measure directly. It would therefore be useful to develop constraints based on sediment topography, which is widely available from profilers and swath bathymetry systems. This approach of modelling sediment topography is not unambiguous since the models only provide explanations consistent with the results of the processes and are not observations of the processes themselves, though modelling may nevertheless provide useful indications for later, more detailed observation.

Many current questions concerning off-axis faulting on mid-ocean ridge flanks require independent methods to date faults, and therefore establish the sequence and rates of deformation. Morphologic dating may provide a solution. Morphologic methods date scarps by quantifying how much scarp topography has been smoothed by erosion and computing the age based on the smoothing predicted, for example, by a diffusion model for the sediment topography and the sediment's model diffusivity k [e.g., *Nash*, 1980]. Applying the model in the marine environment is potentially more complicated, however, because sediment is transported by bottom currents as well as by gravity processes, and reliable morphologic dating may require modelling of both.

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This study area in the Panama Basin, straddling the Galapagos spreading centre (Figure 1), has generally slow currents and, although currents effects are not fully quantified in the following, it should illustrate the potential for morphologic dating.

Interpretation of Deep Tow Profiler Records

Figure 2 shows sediment and basement reflectors across scarps interpreted from Deep Tow sediment profiler records of the Galapagos spreading centre (SC). The pelagic sedimentation rate here is high, ~50 m/my [*Klitgord and Mudie*, 1974], and sediment is draping the basement except immediately near scarps [*Mitchell*, 1995a]. Figure 2 shows that scarp crests are missing sediment that would otherwise be draping (the "eroded space" in Figure 3). This has been removed by gravity-driven transport, erosion by bottom currents or represents reduced or non-deposition due to enhanced current velocities. Sediment slopes reach 30°-40° over scarp crests and may be close to the angle of repose, so some sediment may be removed by surficial avalanching. Sediment at the base of scarps has a more variable morphology, and is either onlapping, horizontal or reducing in thickness towards scarp faces (e.g., profile 8) which suggests some reduced deposition or

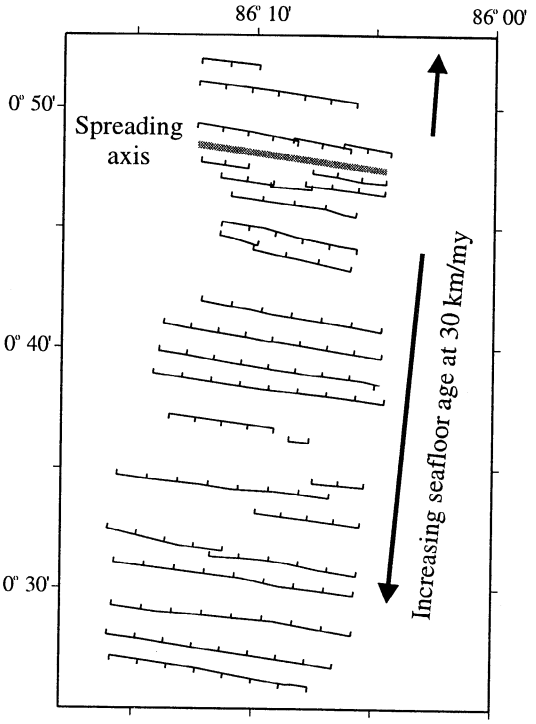

Figure 1. Faults around the Galapagos spreading centre, eastern Pacific, based on Deep Tow data interpretations in *Lonsdale* [1977b] and *Klitgord and Mudie* [1974]. Faults are assumed to increase in age away from the spreading centre, reflecting the increasing crustal age due to seafloor spreading.

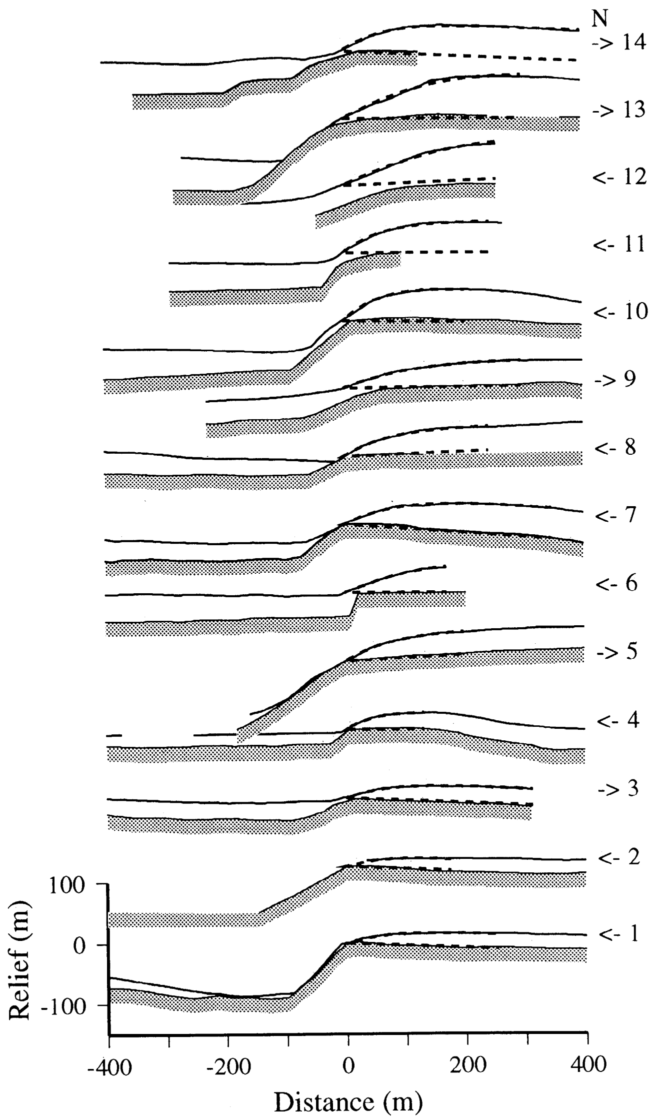

Figure 2. Interpretations of sediment and (shaded) basement surfaces from Deep Tow profiler records across fault scarps around the Galapagos spreading centre [*Klitgord and Mudie*, 1974]. The profiles were projected along a line perpendicular to the spreading axis and are ordered with increasing distance from the axis up the page. Arrows show the north direction and profiles 3, 5, 9 and 12-14 are north of the spreading axis (note there is no systematic difference between north- and south-facing slopes which might otherwise indicate a persistent current flow direction). The dashed straight lines show the unerodable basement of the model and dashed curves show the model sediment topography obtained by fitting equation 4 to the data by minimising the sum of absolute errors with varying k and $h0$, while other parameters were fixed. Curves for the shorter profiles 6 and 12 were computed with $h0$ also fixed. Profiles were selected according to degree of obliquity, lack of hydrothermal deposits and whether the basement and sediment geometry were appropriate for equation 4. In some profiles, the basement is not perfectly flat or is not visible at the scarp edge, such as 10 and 12. These are included to ensure the database reasonably large though were not used to compute mean k for this area.

scouring by currents at scarp bases [*Klitgord and Mudie*, 1974]. Evidently, sediment from scarp crests does not simply accumulate at scarp bases and there are no large volume deposits due to slumps or slides so gravity transport probably does not occur by these mechanisms. Gravity transport by creep due to bioturbation or by small sediment avalanches, however, may intermittently eject small quantities of disaggregated surficial sediment down the scarp face, where the sediment could become resuspended and dispersed by bottom currents, so the lack of deposits at scarp bases does not preclude gravity transport. The following diffusion model describes the creep component of transport and, since some transport from scarp crests probably also occurs by currents and avalanching, this provides an upper bound on creep rates.

Diffusion Equation and Solution

The primary assumption of the diffusion model [*Culling*, 1960] is that the downslope flux of sediment q (kg/m/yr) is proportional to the local slope:

$$q = -\rho k \frac{\partial U}{\partial x} \tag{1}$$

where U is the sediment's topography (m), ρ is sediment density (kg/m) and k is the diffusivity (m^2/yr). Extensive bioturbation is observed in sediment cores from this area [*Borella et al.*, 1983] and surficial creep due to bioturbation may produce downslope transport rates consistent with equation 1 [*Kenyon and Turcotte*, 1985]. A change in q over some distance implies that material is deposited or eroded. Including the rate of supply of pelagics from the water column S (m/yr), which is the mean sedimentation rate for the area, the deposition rate can be expressed in terms of the lateral rate of change of flux using a continuity equation:

$$\frac{\partial U}{\partial t} = -\frac{1}{\rho}\frac{\partial q}{\partial x} + S \tag{2}$$

Differentiating equation 1 and substituting in 2 produces the modified diffusion equation [*Webb and Jordan*, 1993]:

$$\frac{\partial U}{\partial t} = k\frac{\partial^2 U}{\partial x^2} + S \tag{3}$$

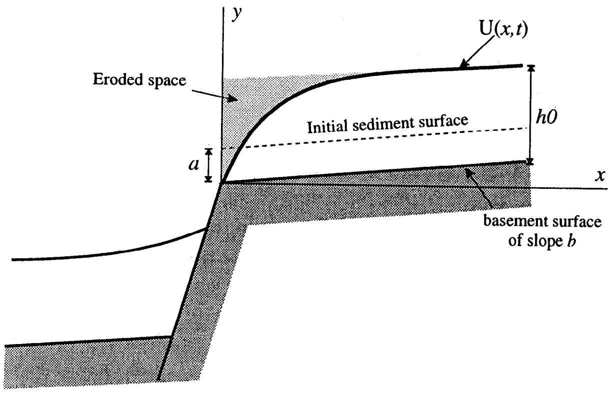

Figure 3. Geometry and terms used to describe scarp profiles. The dashed line shows the sediment surface when the fault formed. The background thickness away from the scarp is $h0$ and $U(x,t)$ represents the sediment topography. The missing sediment area at the scarp face (light grey) is described as "eroded" for brevity. In reality, sediment may also be removed by reduced or non-deposition but this is not easily interpreted from the records.

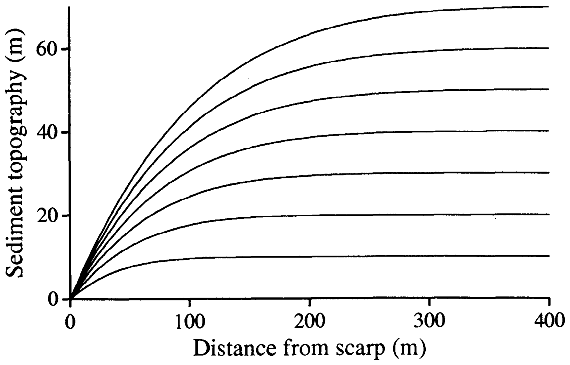

Figure 4. Solutions to equation 4 for a series of 200 ky time steps (b=0, a=0, k=0.01, S=50 m/my). Note that the erosion lengthscale increases steadily while the rate of increase in slope at x=0 is initially rapid but then reduces (also see the model curves in Figure 5b). Hence, scarp age with this geometry is more strongly constrained by erosion lengthscale than scarp slope unlike scarps formed wholly in sediment [*Hanks and Andrews*, 1989].

The geometry of sediment accumulation on scarp crests is shown in Figure 3. The initial sediment topography is $U(x,0)=a+bx$ (dashed line), while the boundary condition for the subsequent evolution is $U(0,t)=0$ (i.e. the basement scarp is effectively unerodable). The appropriate solution to the diffusion equation is [*Carslaw and Jaeger*, 1959, p. 79]

$$U(x,t) = \left(a + St + \frac{Sx^2}{2k}\right) erf \frac{x}{2\sqrt{kt}} + \frac{S}{k}\sqrt{\frac{kt}{\pi}}\, x \exp\left(-\frac{x^2}{4kt}\right) + bx - \frac{Sx^2}{2k} \qquad (4)$$

which is similar to the solution for the repeated faulting problem described by *Hanks et al.* [1984]. The variables in equation 4 are defined in Figure 3 and an example of the evolution predicted by equation 4 is shown in Figure 4. Equation 4 was fitted to the sediment surfaces in Figure 2 (dashed curves) with variable k and $h0$, while b was measured directly from the profiles and held fixed while fitting equation 4. Since a and t are not independent of k, a was set to zero and t was set equal to the estimated basement age (distance from the spreading axis divided by the seafloor spreading rate). The profiles in Figure 2 were digitised from unmigrated profiler records so have distorted slopes. To correct for distortion, the slope of the model curves at the scarp edge (x=0) was converted to true dip in degrees by computing arcsin(slope) and then true slope by computing the tangent of this. k was then recomputed from the adjusted slope [*Hanks et al.*, 1984, equation 14] and this adjusted k was used for the following results.

The erosion lengthscales in Figure 5a were obtained by integrating the model curves in Figure 2 to find the eroded cross-sectional area illustrated in Figure 3 and dividing this area by $h0$, which is the sediment thickness away from the influence of the scarp. Figure 5b shows the sediment slope at the scarp edge (x=0). The circles in Figure 5c show the factor kt obtained from the curve fits, and the crosses show further estimates from profiles over buried basement scarps obtained using the offset-slope method [*Hanks and Andrews*, 1989]. Since this method is designed for scarps formed wholly in sediments and basement scarps may have been initially exposed here, prolonging the initial scarp slope, this method generally underestimates kt. Note that the unfilled circles in Figure 5 are values for profiles without exposed basement which are not properly described by the diffusion model. These were excluded when computing mean k below.

Effects of Bottom Currents

Although not strongly resolved, the erosion lengthscales in Figure 3a generally increase with seafloor age, consistent with a diffusion model. If erosion or non-deposition were due primarily to bottom currents, the erosion lengthscale should be related to scarp height but there is no significant correlation (R=0.05). The erosive potential of bottom currents may be predicted by comparing current meter measurements to erosion experiments on calcareous ooze, which indicate a critical shear velocity of ~4 mm/s (from data of *Southard et al.* [1971] recalculated by

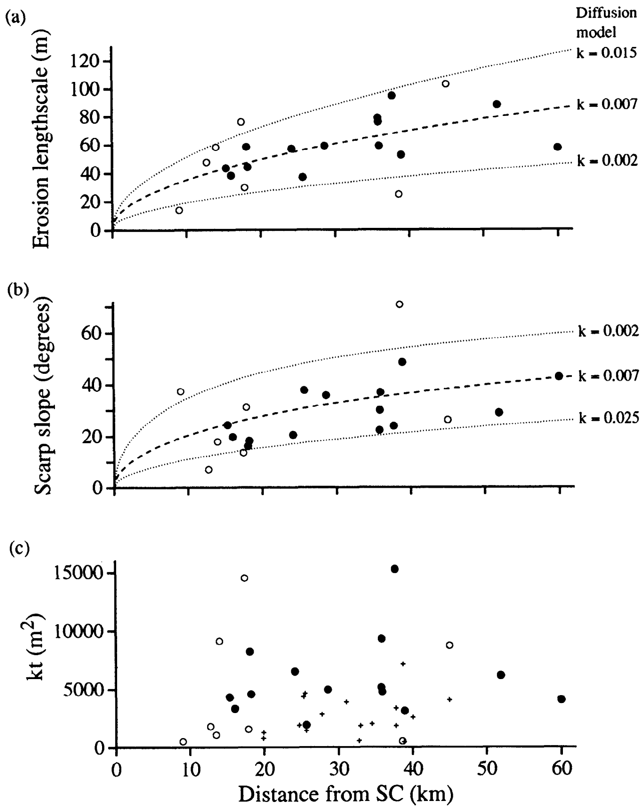

Figure 5. Parameters of the fitted diffusion model curves. The solid circles show values computed from profiles 1-9, 13 and 14 of Figure 2 and some younger scarps not shown. The open circles represent values from profiles 10-12 and others, which are of lower quality or less appropriate geometry for the diffusion equation solution. (a) The erosion lengthscale obtained by integrating U(x,t) over the profile to determine the missing cross-sectional area (light grey in Figure 3), and dividing by $h0$. The three model curves show the lengthscale increase expected from a diffusion model with a=0, b=0 and S=50 m/my (mean sedimentation rate [*Klitgord and Mudie*, 1974]). (b) Slopes of the scarp edge corrected for time-section distortion. This correction enlarges slope errors increasingly with slope so those above 40° are probably inaccurate. (c) The model age, kt, is variable, which may be partly due to data and analysis errors. Assuming errors are small, however, kt varies for multiple crossings of scarps, suggesting variable k rather than t (i.e. variable scarp erosion rates rather than faults forming off-axis). The crosses show further estimates of kt obtained for surfaces over buried basement scarps using the slope-offset method [*Hanks and Andrews*, 1989] which underestimates kt here (please see main text).

McCave [1984]). Extrapolating this to current velocities above the turbulent boundary layer using the Kármán-Prandtl equation [e.g., *Massey*, 1989, equation 8.46] and assuming a boundary layer thickness of 5 m above a smooth seabed, the critical velocity at the height of the current meters is roughly 12 cm/s (see *Southard et al.* [1971] for discussion of problems with in-situ prediction). Measured current velocities are generally less than 12 cm/s suggesting erosion for only limited times: 12.5 cm/s peak in 21 days [*Lonsdale*, 1977a], average 3.5 cm/s and occasional peak 11 cm/s in 28 days [*Lonsdale*, 1977b], and 2.4 cm/s northward average with 4.5 cm/s tidal and occasional 20 cm/s peak in 23 days [*Detrick et al.*, 1974]. The high velocities were associated with strong tidal cycles rather than persistent drift. This argument does not account for reduced pelagic deposition at velocities lower than the erosion threshold [*McCave*, 1984] but at least suggests little or no erosion generally.

Variations in current velocity caused by flow across the seafloor's topography may be qualitatively inferred by analogy with flow experiments across steps. Experiments over negative steps (right to left in Figure 2) typically indicate no change in mean bed shear stress over scarp crests [*Allen*, 1968] so we might expect merely translation of bottom sediments by flow in that direction and no effect on sediment topography. Experiments over positive steps [e.g., *Moss et al.*, 1979], however, typically show a vigorously circulating bubble at the base of the step, which may explain thinning in some profiles of Figure 2. Bunched streamlines occur immediately at scarp crests, indicating enhanced bed shear stress. Although not simple to quantify, this may reduce deposition or cause some erosion at scarp crests and therefore the diffusion model results are upper bounds on downslope transport rates due to creep.

Discussion and Conclusions

Reliable morphologic dating will require further work to resolve the ambiguity caused by bottom currents, although we can at least estimate the data averaging required to achieve reasonable precision. Assuming that the scarps formed at the spreading axis and using only the data shown by the solid symbols in Figure 5c, the range of points suggests k varies from 0.002 to 0.015, with a mean of 0.007 and standard deviation 0.004 m^2/yr (part of this range may be caused by errors due to towfish navigation, profiler record digitising and curve fitting, although these are difficult to quantify). In order to achieve 2σ confidence limits of 10% of mean kt, ~100-200 independent estimates of kt for a scarp would need to be averaged, which should be achievable using data from modern deeply towed swath bathymetry systems.

The rates of movement predicted by these model results are small, for example 40 cm^2/yr or 16 g/cm/yr for typical scarp edge slope of 30° and 0.4 g/cm^3 dry bulk density. Assuming a layer thickness of 25 cm corresponding to the oxidised surface layer [*Borella et al.*, 1983], these values imply creep rates of less than 2 cm/yr and proportionally less for lower slopes. The estimate of k, 0.007 m^2/yr, is lower than in Mid-Atlantic Ridge studies (0.13 m^2/yr [*Webb and Jordan*, 1993] and 0.04-0.11 m^2/yr [*Mitchell*, 1995b]), consistent with the suggestion that movement over the Galapagos SC is mostly by creep due to bioturbation while that in the Atlantic is by repeated resuspension [*Mitchell*, 1995b]. Since some movement or reduced deposition may also occur due to bottom currents and sediment avalanching, the model diffusivity provides an upper bound on creep rates.

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