The marine sediments of the Guaymas Basin are unusually rich in organic matter (~2% organic carbon), mostly of planktonic origin (*20*). Furthermore, the presence and vertical distribution of acetate and propionate in the Guaymas Basin hot sediments indicate that these organic acids are thermocatalytically produced from organic matter in deeper sediment layers at temperatures greater than 200° to 300°C. They appear to be consumed only in the top sediment layers where the temperatures fall below 110° to 120°C (*21*). Hydrothermally generated methane as well as aliphatic and aromatic hydrocarbons also rise from deeper layers by pore fluid transport and are abundant in the surface sediment (*22*). Although mesophilic use of C_{12} to C_{18} straight-chain aliphatics has been identified as a potential source of substrates for bacteria that reduce sulfate (*23*), the metabolic availability of hydrocarbons for anaerobic bacteria at higher temperatures has not yet been investigated.

It has not been uncommon to detect a microbial process before the isolation and description of the responsible organism. Yet, it was unusual that the first discovery of sulfate reduction at 90°C in the hot seabed of Guaymas Basin (*15*) was immediately followed by the isolation from the same sediment material of *Archaeoglobus profundus*, a sulfate-reducing bacterium growing in temperatures up to 90°C (*13*). Attempts to isolate the organisms growing at temperatures >100°C have so far been unsuccessful (*24*). The radiotracer technique we applied for measuring sulfate reduction at temperatures >100°C is sensitive and free from temperature-dependent artifacts, such as isotopic exchange or chemical reduction, within the temperature interval studied here. Enrichment culture and ultimate isolation of the presumed organisms are, in contrast, more uncertain and unpredictable without knowledge of substrate specificity and other growth requirements. As long as pure culture evidence for the existence of the bacteria is lacking, the radiotracer technique is thus an effective and reliable alternative to detect this microbial process.

The metal sulfides at the mid-oceanic ridges are by far the largest recent deposits on the global scale (*3*). Our results expand the temperature tolerance of bacterial sulfide production to a range where it approaches the lower temperature limit of thermolytic sulfate reduction (*25*). At such temperatures, this microbial process becomes important for an interpretation of the formation of sulfide deposits and their sulfur isotope distribution.

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17. Sediment sampled and collected at depth intervals of 5 cm from a core 15 cm wide by *Alvin* was used to measure the temperature dependence of sulfate reduction. Sediment was mixed in a 1:2 ratio with anoxic, artificial seawater, and 10-ml subsamples were dispensed under N_2 into test tubes and injected after temperature equilibration with 2 to 20 μCi of $^{35}SO_4^{2-}$. The tubes were incubated at more than 30 different temperatures in a stable temperature gradient of 3° to 120°C (±0.5°C) for 1 day (sample from 0 to 5 cm) or 7 days (sample from 10 to 15 and 15 to 20 cm). At temperatures >100°C, tubes were sealed by screw caps and thus attained equilibrium vapor pressure of >1 bar. Sulfate reduction rates were determined from the formation of ^{35}S-labeled, reduced sulfur as analyzed by the single-step chromium reduction technique (*18*). Although incubations were done at around 1 bar rather than at in situ pressure, a pressure release from 2000 m seems generally not to change metabolic activity of heterotrophic bacteria significantly (*26, 27*).
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Downstream Fining by Selective Deposition in a Laboratory Flume

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There has long been debate about the relative importance of abrasion versus selective deposition of the coarsest clasts in causing downstream fining of sediment in river systems. Although high fining rates observed in many natural rivers seem to require strong selective deposition, the ability of selective deposition to produce downstream size sorting has never been measured under controlled conditions. In an experiment using a long flume and a poorly sorted, bimodal gravel feed, downstream fining was produced by a factor of 1.3 in median size and 1.8 in 90th percentile size, over a distance of 21 meters. The experimental conditions rule out abrasion effects. Selective deposition appears to be a natural consequence of the transport and deposition of sufficiently poorly sorted or bimodal gravels and appears to be capable of accounting for fining rates observed in natural gravel rivers.

Going downstream most natural rivers become finer grained, higher in discharge, gentler in slope, and more sinuous. These changes have important effects on vegetation, flood characteristics, ecological habitats, and so forth. Although there has been some debate as to which of these effects are primary and which are secondary, it is clear that downstream changes in grain size and discharge must be among the basic driving factors. There are two broad explanations of the observed downstream decrease in grain size (downstream fining): abrasion, a nonconservative mechanism that converts large clasts into smaller ones; and selective deposition of the coarser clasts, in which the sizes of individual clasts are conserved and downstream fining results from a sedimentary fractionation process. Although it has long been recognized that selective

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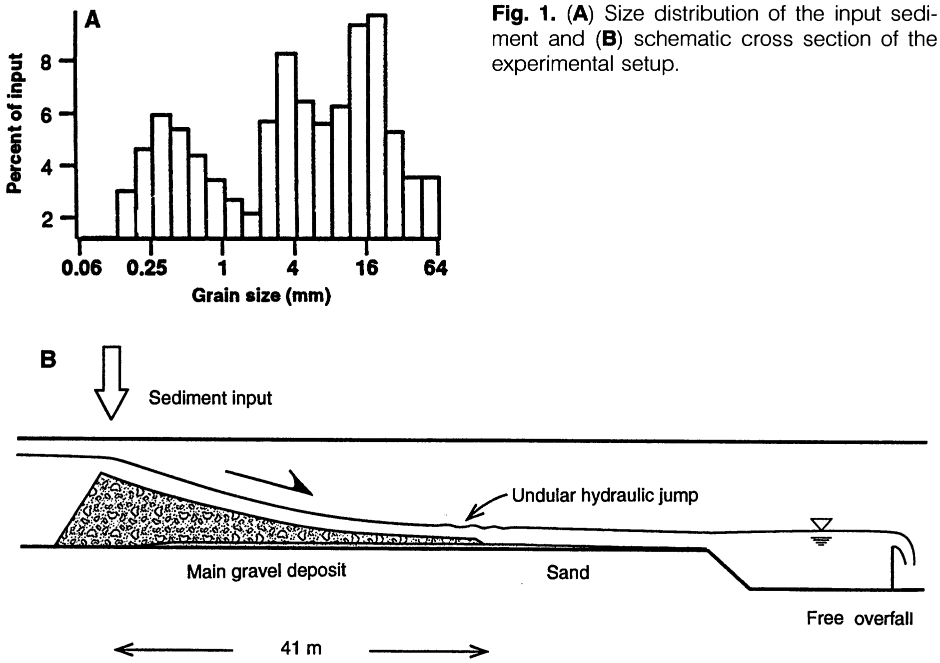

Fig. 1. (**A**) Size distribution of the input sediment and (**B**) schematic cross section of the experimental setup.

deposition is probably an important fining process (*1–4*), historically abrasion has received more attention. The principal evidence for abrasion is usually either that clasts of different lithology, but similar density, fine downstream at different rates or that the reach in question is not depositional (*5–10*). Sternberg (*11*) proposed a simple abrasion model in which the rate of clast attrition was assumed to be proportional to clast size, leading to an exponential law for downstream size decrease:

$$D = D_0 e^{-ax} \tag{1}$$

where D is some characteristic grain size (for example, the median size), D_0 is its initial value, a is an empirical coefficient, and x is downstream distance.

There is little doubt that abrasion is an important mechanism of downstream fining. However, values of the fining coefficient a estimated from abrasion experiments in rolling mills are orders of magnitude lower than many values observed in the field, and the discrepancy becomes stronger as the rate of deposition increases (*10*). Although some of this discrepancy could be explained by positing more elaborate abrasion mechanisms (*12–14*), the observed dependence on deposition rate suggests that selective deposition must be important, at least in aggradational rivers.

Selective deposition would seem a natural consequence of the fact that small grains are easier to transport than large ones. Nonetheless, a considerable body of field, experimental, and theoretical work suggests that this is not necessarily so. During transport of sediment comprising a range of grain sizes, the larger grains protrude farther into the flow and roll over a surface that is relatively smooth, whereas the finer grains tend to become entrapped among the larger grains. The sum of these effects appears, in many cases, to render all the grains in the mixture nearly equally mobile (*15–18*), although other workers have argued for varying degrees of deviation from perfect equal mobility (*19–24*). The limited experimental data available on gravel deposition have added to the uncertainty, showing downstream coarsening rather than fining (*25, 26*).

Progress in determining the relative importance of each fining mechanism has been hampered by the difficulty of isolating either mechanism in the field and of performing experiments of sufficiently large scale to observe natural fining processes. Thus, we have undertaken a series of laboratory experiments aimed at producing gravel deposits under controlled conditions in a system large enough to allow detection of the effects of selective deposition but short enough that abrasion can be unambiguously ruled out.

We fed a poorly sorted gravel mixture into the upstream end of a channel 45 m long, 1.2 m deep, and 0.3 m wide (Fig. 1). The material in the feed ranged from 0.125 to larger than 64 mm, and the feed distribution was essentially bimodal with primary modes at about 0.35 and 16 mm and a minimum around 2 mm. This bimodality with a minimum around 2 mm is common in natural gravel-bed streams (*10*), where it is often much more pronounced than in our experiment. The gravel comprised well-indurated, unfractured clasts of various lithologies but with no significant variation in density.

Water discharge during the run was held constant at 49 liter s^{-1}. We fed sediment manually at the upstream end of the chan-

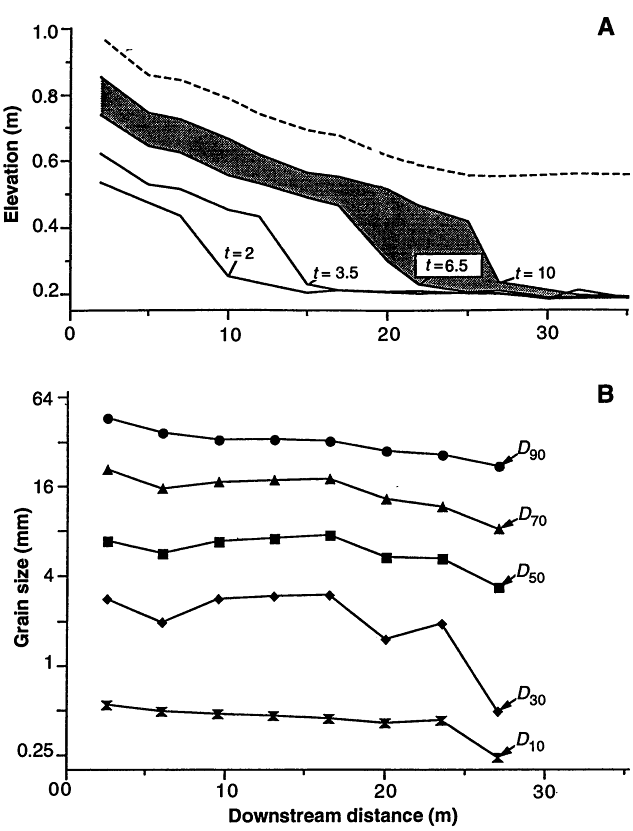

Fig. 2. (**A**) Profiles of the bed elevation measured at four different times (measured in hours) during the experiment. The water surface at t = 10 hours is given by the dashed line. The shaded area was sampled after the experiment for measurement of size distribution. (**B**) Variation in five percentile measures (the percentile D_x is defined as the size such that x percent of the distribution is finer than D_x) of sediment size along the sampled section, showing fining across the entire size distribution.

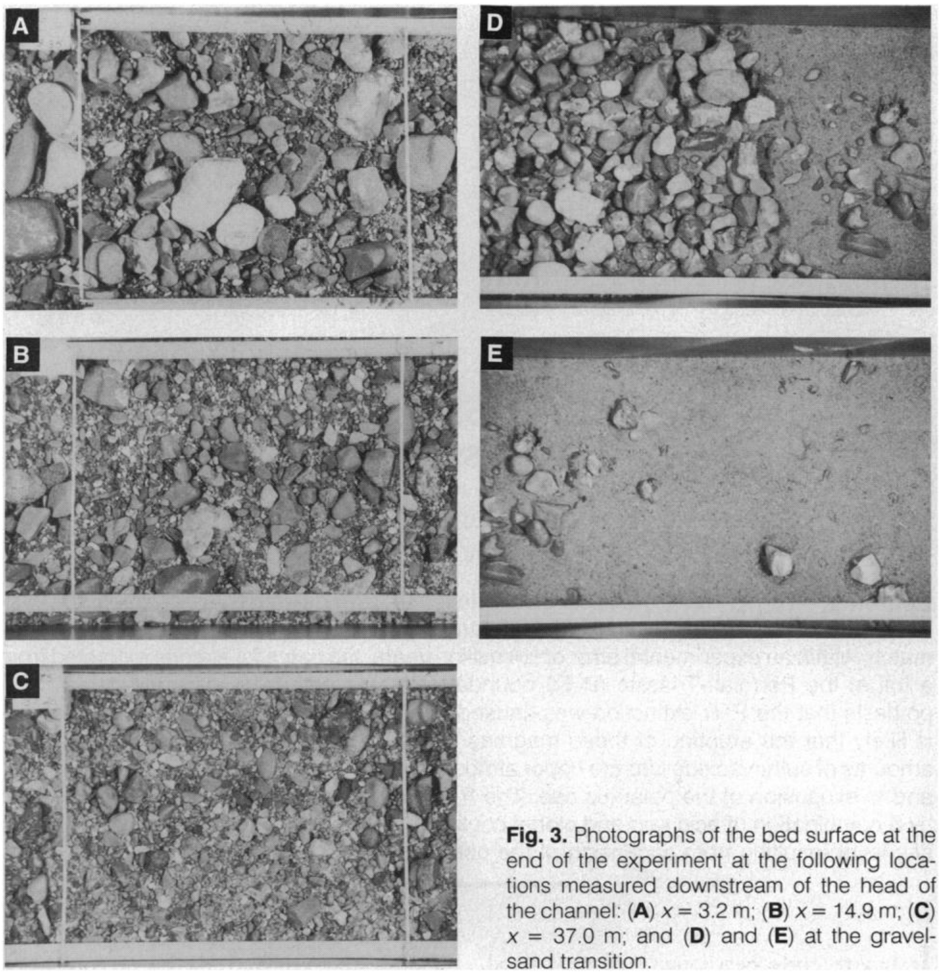

Fig. 3. Photographs of the bed surface at the end of the experiment at the following locations measured downstream of the head of the channel: (**A**) $x = 3.2$ m; (**B**) $x = 14.9$ m; (**C**) $x = 37.0$ m; and (**D**) and (**E**) at the gravel-sand transition.

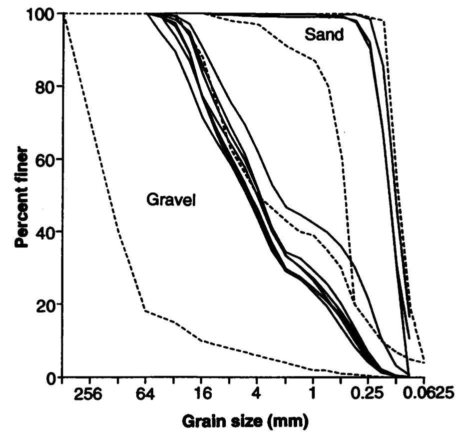

Fig. 4. Cumulative distributions of samples (solid lines) from the main gravel deposit and downstream sand deposit of the experiment. Each line shows the percentage of grains in the sample that are finer than a specified size. These data are compared with the envelopes bounding the "gravel" and "sand" size distributions (dashed lines) measured in the field by Shaw and Kellerhals (*10*). The division into "gravel" and "sand" distributions is based on sample median size D_{50}.

nel at a constant rate of 11.3 kg min^{-1} for 16 hours 50 min. A wedge-shaped gravel deposit with a well-defined front developed almost immediately on the horizontal initial surface and then propagated down the channel, eventually reaching a length of 36 m. There was no significant local topography or sorting on the sediment surface. Bed profiles (Fig. 2A) show the motion of the front and a weak but persistent upward concavity of the gravel deposit upstream of the front.

We measured the variation in grain size for the deposit between the 6.5- and 10-hour time lines (Fig. 2, A and B). Surface character is shown in Figure 3. The data show consistent decreases in all the measured size percentiles, although the rate of decrease is highest for the higher percentiles, representing the coarse tail of the size distribution. Not including values measured nearest the front, D_{90} declined from 46.7 to 26.6 mm, D_{50} from 7.0 to 5.3 mm, and D_{10} from 0.57 to 0.43 mm over a distance of 21 m. The short length and duration of the experiment, together with the durability of the clasts, rule out any possibility that these results are due to abrasion. The observed change in D_{50} gives a fining coefficient a (Eq. 1) of 13.2 km^{-1}, larger than the maximum values reported from alluvial fans (5 to 7 km^{-1}) (*10*). The high a value is what one would expect if the mechanism responsible for selective deposition in the field were constrained to operate in a short system.

We have also observed a similar pattern of downstream fining from the input point to the gravel front in the deposit between the 2- and 5-hour time lines of Fig. 2A. Evidently, this fining pattern is set up very early in the run and then maintained as the bed aggrades and the front migrates downstream.

The fining pattern is also reflected in the persistent upward concavity of the sediment surface upstream of the front (Fig. 2). This concavity reflects a downstream decrease in the slope required for transport as a result of downstream decreases in both total sediment flux (due to deposition) and grain size (due to selective deposition). It was clear during the experiment that the bed slope constructed by the system at its upstream end was sufficient to transport even the largest clasts provided in the feed.

The percent sand (grain size <2 mm) in the deposit generally increases downstream, especially near the downstream end of the deposit where there is an abrupt transition from a gravel-sand mixture to pure sand. The point of transition is just downstream of the gravel front. It appears (Fig. 3D) that the grain size of the gravel increases slightly near the transition. The transition was associated with a weak undular hydraulic jump (Fig. 1). Both were evident throughout the run.

A fairly abrupt transition from gravel- to sand-bed morphology is also common in natural streams (*10, 27*), although the hydraulics of the transition in natural streams probably differ from those of our experiment. The gravel-sand transition has often been associated with pronounced bimodality in the bed material (Fig. 4). This bimodality is associated with a relative paucity of sediment in the range 1 to 4 mm; samples are either composed predominantly of gravel with a median grain size in excess of 4 mm or exclusively of sand with a median grain size below 0.5 mm. This is true even though sediments with median sizes in the intervening range could be produced from the material available.

In our experiment, the grain-size change at the gravel-sand transition is much more pronounced than the weak difference between the gravel and sand modes in the feed mixture (Fig. 4); evidently the gravel-sand transition develops because of sorting processes and is not simply a reflection of bimodality in the sediment supply. We suggest that there is a cutoff size in the medium-sand range above which grains cannot be suspended; they can roll or saltate as bedload up to and on to the gravel front but cannot be transported any farther in the zone of reduced shear stress downstream of the undular hydraulic jump. Grains below this cutoff size can be transported in suspension across the front and through the jump. They are then available for deposition farther downstream, giving rise to the sand tail (Fig. 1).

Our results show that significant downstream fining can be produced over short distances in the laboratory through selective deposition. We cannot specify precisely the critical factors required to produce well-developed fining but, in our view, the most important respects in which our experiment differed from previous ones were the wide size range and bimodality of the feed mixture and the channel length. In a similar series of detailed experiments performed in a smaller channel, Wilcock and McArdell (*28*) showed that there were strong deviations from equal mobility using the sediment mixture we used in our experiment; fractional transport rates vary inversely with grain size, and the critical shear stress to initiate motion of individual sizes increases with grain size over an order of magnitude. It is likely that both the bimodality of the grain-size distribution and its large standard deviation are important in producing the size-dependent variation in mobility (*29*).

Our results also indicate some factors that are apparently not required to produce strong downstream fining; neither variation in discharge (temporal or spatial) nor preexisting slope variations were present in our experiment. For the simple experimental geometry reported here, downstream fining by selective deposition appears to require only the transport and deposition of a sufficiently poorly sorted or bimodal gravel.

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Synchronism of the Siberian Traps and the Permian-Triassic Boundary

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Uranium-lead ages from an ion probe were taken for zircons from the ore-bearing Noril'sk I intrusion that is comagmatic with, and intrusive to, the Siberian Traps. These values match, within an experimental error of ±4 million years, the dates for zircons extracted from a tuff at the Permian-Triassic (P-Tr) boundary. The results are consistent with the hypothesis that the P-Tr extinction was caused by the Siberian basaltic flood volcanism. It is likely that the eruption of these magmas was accompanied by the injection of large amounts of sulfur dioxide into the upper atmosphere, which may have led to global cooling and to expansion of the polar ice cap. The P-Tr extinction event may have been caused by a combination of acid rain and global cooling as well as rapid and extreme changes in sea level resulting from expansion of the polar ice cap.

It has recently been suggested that flood volcanism is produced by melting of the head of starting plumes that originate at the core-mantle boundary and grow by entrainment as they rise. Thus, by the time they reach the top of the mantle they form giant disks with a diameter of ~2000 km (*1*). This hypothesis offers a simple explanation for the vast lateral extent (typically 2000 by 2000 km) and short duration (1 to 2 million years) of flood volcanism (*2*). Flood volcanic provinces represent the most dramatic outpourings of basalt in the Phanerozoic record, with eruptive rates that are one to two orders of magnitude greater than those associated with normal oceanic hotspots (*3*).

At least ten mass extinctions have been recognized during the last 250 million years. Nine of these, including the two largest at the Cretaceous-Tertiary (K-T) and P-Tr boundaries, may be temporally associated with episodes of flood volcanism (*4*). Recent data for the P-Tr boundary and Siberian Traps, in particular, permit a correlation between the two events. The age of the P-Tr boundary is 251.1 ± 3.6 million years (2σ) based on ion probe analysis (*5*) of zircons extracted from a bentonite layer from the Meishan section at Changing, China. This value agrees with an earlier averaged estimate of 249 ± 4 million years from K-Ar data (*4*). Renne and Basu (*6*) reported whole-rock and plagioclase ^{40}Ar-^{39}Ar incremental heating data for flows from the Siberian flood basalts that indicate that flood basalt volcanism commenced at 248.4 ± 2.4 million years (2σ). Dalrymple *et al.* (*7*) obtained ages of 243.5 ± 1.8 and 244.9 ± 1.8 million years by ^{40}Ar-^{39}Ar laser incremental heating of plagioclase separates from apparently fresh flood basalt drill core from within 300 m of the base of the basalt sequence. These results were duplicated by Baksi and Farrar (*8*) who obtained similar ages of 243 to 244 ± 1.2 million years by whole-rock ^{40}Ar-^{39}Ar laser incremental heating of a sample from the lowest flows of the same drill hole. The discrepancy between these results and those of Renne and Basu arises largely from differences in the age ascribed to the neutron flux monitor MMhb-1 by the different laboratories: Renne and Basu used 520.4 ± 1.7 million years, whereas Dalrymple and colleagues and Baksi and Farrar used 513.9 ± 2.3 million years (*9*). If this

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