ORIGIN OF FLUVIAL GRAIN-SIZE TRENDS IN A FORELAND BASIN: THE POCONO FORMATION ON THE CENTRAL APPALACHIAN BASIN

RUTH A.J. ROBINSON* AND RUDY L. SLINGERLAND

Department of Geosciences, Pennsylvania State University, University Park, Pennsylvania 16802, U.S.A.

** Present address: School of Geography and Geosciences, University of St Andrews, St Andrews KY16 9ST, Scotland*

ABSTRACT: **A widely recognized phenomenon of modern and ancient river systems is downstream decrease in grain size. Over the past two decades, theoretical formulations, observations of modern rivers, flume studies, and numerical models have significantly increased our understanding of sediment transport in mixed-grain-size river systems. These have established that downstream fining in modern and ancient rivers can be attributed to a combination of selective sorting, abrasion rate, and accumulation rate. However, no detailed multiple-grain-size sediment transport model has been used to address how much subsidence rate, sediment flux, water discharge, hydraulic geometry, and the mechanics of sediment transport influence grain-size distribution and facies belt development in ancient fluvial systems. We have combined the results of recent empirical and theoretical studies with a multiple-grain-size sediment transport model (MIDAS) to test the sensitivity of downstream fining trends to those controlling mechanisms. Our results demonstrate that subsidence and sediment feed rate are the most important mechanisms controlling downstream fining trends in a foreland basin if the evolution of hydraulic geometry is known. To illustrate the application of this result, we have replicated the textural trends of the Mississippian upper Pocono Formation and Burgoon Sandstone of Pennsylvania. This methodology can be applied to constrain plausible values of accumulation and subsidence rate for ancient alluvial deposits and enhances our ability to interpret paleohydraulic conditions from such facies in a foreland basin.**

INTRODUCTION

It is now accepted that most fluvial deposits decrease in average grain size downstream. Numerous field studies of deposits, both modern (Ashworth and Ferguson 1989; Werritty 1992; Pizzuto 1995; Ferguson et al. 1996) and ancient (Meckel 1967; Paola 1988; Heller and Paola 1989, 1992), have observed downstream fining even though the methodologies vary and their at-a-station samples are far from representative. Using a practice codified nearly forty years ago by the pioneering work of Francis Pettijohn and his students in the Paleozoic fluvial deposits of the Appalachian foreland basin, geologists routinely exploit grain-size trends to infer paleogeography and distance to source terrains (e.g., Yeakel 1962; Meckel 1967). The technique is well illustrated by Pelletier (1958), who used observed grain-size trends in the Mississippian Pocono Formation of Pennsylvania (Fig. 1) to infer the location of the Acadian orogenic highlands. Pelletier assumed that Sternberg's (1875) relationship (as published in Barrell 1925) of grain size versus distance downstream, derived from grab samples of the Rhine River, should also describe textures of the fluvial Pocono Formation derived by repeatedly averaging the ten largest clast sizes over hundreds of meters of accumulation. Sternberg's relationship is

$$D = D_o e^{-\alpha x} \qquad [1]$$

where D is the observed grain size at particular distance x, D_o is the assumed size of the clasts at the headwaters of the stream, and α is the fining rate (length^{-1}). By calculating the fining rate from the observed grain-size trends in the Pocono Formation and assuming a clast size for the sediments in the headwaters (in the manner of Plumley 1948), Pelletier predicted that the source terrain for the Pocono rivers lay near Atlantic City, New Jersey, about 185 km to the southeast of the last Pocono outcrop in southeastern Pennsylvania (Fig. 1; Pelletier 1958).

However, starting with Paola (1988), a new generation of studies (Parker 1991a, 1991b; Ferguson and Ashworth 1991; Paola et al. 1992a; Paola et al. 1992b; Pizzuto 1995) has demonstrated that the exponential downstream grain-size decrease so commonly seen in modern and ancient river sediments is both a cause of and a response to downstream changes in a complete suite of fluvial variables. Numerical models of downstream fining have investigated the effects of selective transport (Parker 1991a; Paola et al. 1992a; Hoey and Ferguson 1994; Cui et al. 1996), abrasion (Parker 1991a), and spatial variations in sediment supply (Pizzuto 1995). Observations have demonstrated that lithologically controlled abrasion rates additionally influence downstream fining trends (Werritty 1992; Mikos 1993; Kodama 1994a, 1994b), although the general consensus is that selective transport is the dominant control on downstream fining in quartz-dominated gravel-bed rivers (e.g., Ferguson and Ashworth 1991; Walthen et al. 1995). Sediment supply is generally thought to be a significant factor influencing selective transport and the texture of bed material. Rana et al. (1973) demonstrated that independently increasing sediment supply and water discharge coarsens bed material for a sand-bed river. Dietrich et al. (1989) modeled the effect of reducing sediment supply of coarse-grained material in a flume and demonstrated that grain size is coarsened at the flume edges while a central zone of bedload transport develops a fine texture. Hoey and Ferguson (1994) similarly demonstrated that decreasing sediment supply produces coarser grain sizes and larger downstream fining rates. Therefore, predictions of distance to a source terrain based on a downstream fining rate do not include the influences of subsidence, discharge, sediment caliber, and sediment supply and may therefore be incorrect.

Can a particular rate of downstream fining tell us anything about a particular ancient fluvial system? Although Paola et al. (1992a) have demonstrated that grain size, water discharge, sediment feed rate, and subsidence rate affect gross stratigraphic patterns in foreland basin sequences, we attempt to address the previous question by using a sediment transport model to explore relationships between downstream fining and a generic river's aggradation rate, hydraulic geometry, sediment and water feed rates, and grain-size distribution in transport. Under the influence of tectonic subsidence, the river is allowed five dependent degrees of freedom: channel width, velocity, water depth, bed slope, and grain size. For aggrading alluvial systems, the definition of channel slope as a fully dependent variable which is a response to other causes (size distribution, sediment feed rate, water discharge, adjustment of hydraulic geometry, and subsidence rate) is important and differs from some previous modeling studies (Rana et al. 1973; Parker 1991b). Our analysis does not include the effects of abrasion, which is a reasonable simplification in a quartz-dominated or rapidly aggrading river system, or the adjustment of channel shape and roughness, both of which influence sediment transport rates. Our conclusion is that, given these omissions and for a particular grain-size distribution, it is subsidence and sediment feed rate that are the first-order controls on downstream fining as long as the evolution of hydraulic geometry is known. We provide an example of how fining rates record an independent estimate of basin subsidence rates and paleohydraulic conditions.

Influence of Aggradation Rate on Downstream Fining

To understand the effect of aggradation on downstream fining, consider the following hypothetical case. A river is supplied with a range of sedi-

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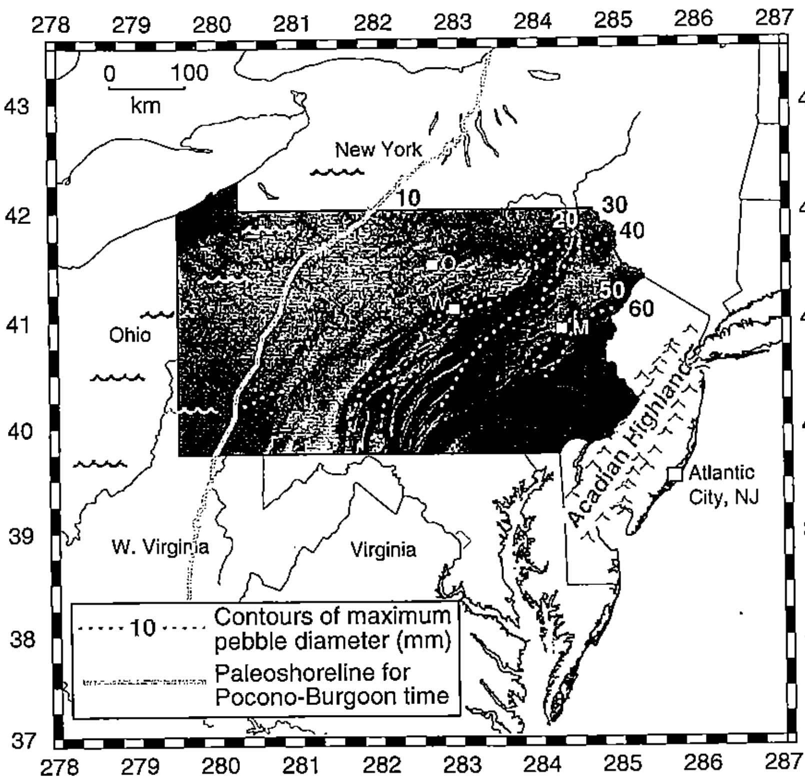

FIG. 1.—Present geography (in black) and Mississippian paleoshoreline and grain-size trends (in gray and white) for Pocono–Burgoon deposition. DEM of Pennsylvania shows valley and ridge topography in eastern-central regions; ridge-forming Pocono outcrops are lightest gray shades (highest elevations). M = Mauch Chunk; W = Williamsport; O = Ogdensburg.

ment sizes that enter at the headwaters (Fig. 2) and are resistant enough such that the effects of abrasion are minimal. Channel width and water discharge increase downstream consistent with standard geomorphic relationships, but an assumption of clear-water tributaries is adopted in order to analyze the effects of sediment supply from one location at the headwaters of the stream, although it is recognized that sediment from tributaries can affect downstream fining rates (Pizzuto 1995; Rice in press). If the initial bed slope is low enough, sediment fed into the stream preferentially accumulates in the upper reaches because the sediment transport rates are less than the feed rates and/or because bed shear stresses are smaller than the threshold stresses, especially for the coarser grains. Through conservation of mass, the river bed aggrades at the rate

$$\frac{\partial h_c}{\partial t} = \frac{1}{(1-\phi)}\frac{\partial q_{s_c}}{\partial x} \qquad [2]$$

where h_c is bed elevation (m), q_{s_c} is volumetric sediment feed rate per meter width of stream (m^2/s), ϕ is porosity, and x is distance downstream (m). Because divergences in sediment flux are greatest in the upper reaches, bed aggradation is greatest there and the longitudinal profile steepens (Fig. 2B). As the slope increases, more and coarser sediment can be transported farther downstream as time progresses, producing a deposit that coarsens upwards and fines downstream. In the absence of other controls, the river profile continues to steepen until its slope is adjusted to transport all the incoming sediment fractions at the feed rate of each fraction. In a classical sense, the river is then at grade (an extension of Mackin's [1948] definition of grade) because it is transporting the sediment feed at the feed rate of each size fraction. Most importantly, the bed does not fine downstream, as is illustrated by the grain-size contours converging at the downstream end by the last time line (Fig. 2B).

Now consider how basin subsidence modifies the preceding scenario. Assume that subsidence can be modeled as a curve rotating around a fixed point downstream (Pitman 1978) and that subsidence is maximum at the upstream end and exponentially decreases downstream. Here too, the stream slope adjusts as more, and proportionally coarser, sediment is accumulated in the upper reaches, where the accommodation space created per unit time is larger (Fig. 2C). Eq 2 becomes

$$\frac{\partial h_c}{\partial t} = \frac{1}{(1-\phi)}\left(\frac{\partial q_{s_c}}{\partial x}\right) - \omega_x \qquad [3]$$

where ω_x is the subsidence rate (m/s). Here too, the bed initially steepens but only until the deposition rate is equal to the subsidence rate at each point downstream. This state of equilibrium is fundamentally different from Mackin's (1948) definition of grade because the river has adjusted to transport only the amount of sediment remaining in transport after the deposition of material due to subsidence. Therefore, this equilibrium condition is a delicate balance between mass removed due to subsidence and mass added due to sediment feed. The bed profile has a lower bed slope than the previous example because there is a decreasing amount of sediment to be transported at each point downstream. Most importantly, the bed fines downstream and at equilibrium the fining rate remains constant in successive chronostratigraphic units. Below, the sediment transport algorithms are quantified in a loose-boundary sediment sorting model called MIDAS.

THE MODEL

MIDAS (van Niekerk et al. 1992; Vogel et al. 1992) is a one-dimensional uncoupled fluid flow and multiple grain-size sorting model that simulates transport, deposition, and erosion in a single-thread, straight channel. Given variations in water discharge and stream width downstream, the gradually varied flow equation calculates longitudinal velocity and water depth. Sediment grains experience both viscous (skin friction) and pressure (form) drag components of bed shear stress, and the importance of each on sediment transport depends on whether a flow boundary is hydraulically smooth or rough. Einstein and Barbarossa (1952) proposed that the vertical velocity profile for a rough boundary is a function of grain roughness (2.5 times the median grain size) and an effective bed shear stress (τ_0') that is reduced

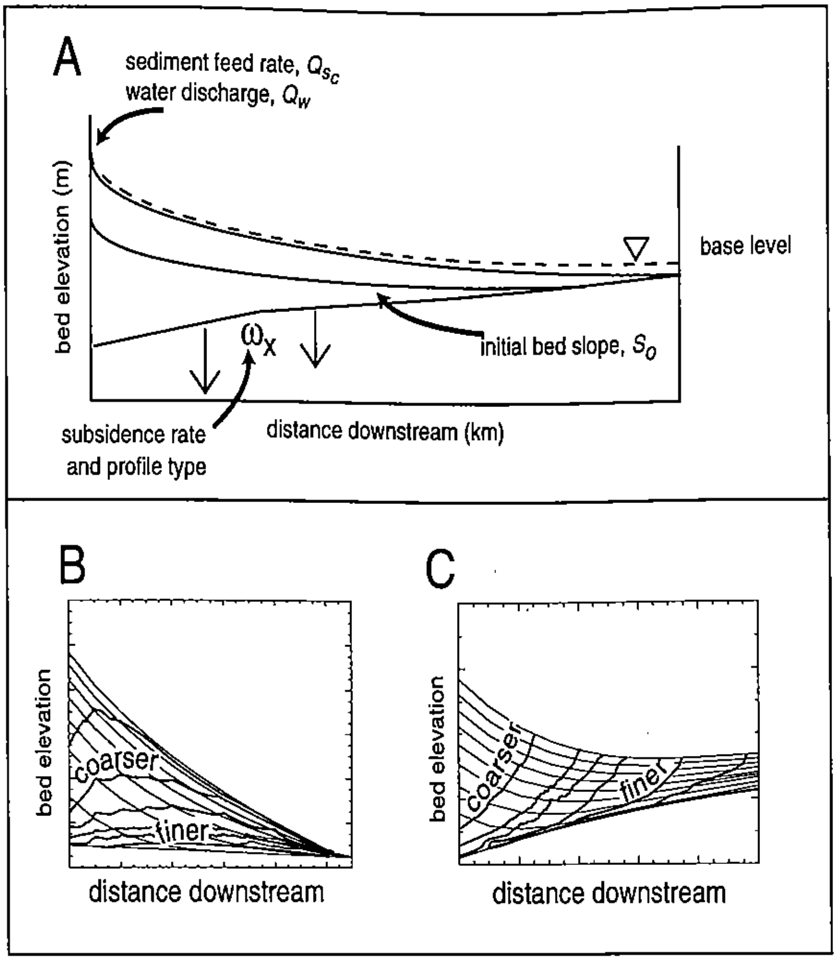

FIG. 2.—Experimental design. **A)** Important variables controlling downstream fining. **B)** Model predictions of bed aggradation and grain-size trends in the absence of subsidence. Thin concave lines are bed elevation through time; thick lines are downstream grain-size trends contoured through time. **C)** Model predictions of trends in the presence of subsidence.

to reflect only the magnitude of drag involved in transport. Instantaneous bed shear stress (τ_k') is distributed in a log-Gaussian manner to represent the turbulent burst cycle. Bedload transport ($i_{b_{ijk}}$) is calculated using a modified Bagnold equation (Vogel et al. 1992),

$$i_{b_{ijk}} = (F_{ij}P_k)\frac{h}{\tan\alpha}(u_{*k}' - u_{*cij})(\tau_k' - \tau_{cij}) \qquad [4]$$

which is scaled by the proportion of each individual size in the bed (F_{ij}) and the time elapsed for each shear-stress interval (P_k), over all shear velocities (u_{*k}'), and for the critical thresholds (u_{*cij}, τ_{cij}) of all grain sizes and densities (ij). The total sediment transport rate is the sum of bedload and suspended-load transport, which is calculated from a vertically integrated Rouse equation.

The model separates the bed and near-bed region into four parts: (1) stacked, accumulated layers of deposited material; (2) an active layer at the top of the bed (defined as two times the median grain size); (3) a moving bed layer of bedload transport, and (4) water depth divided into a number of vertical slices for suspended load calculations. The downstream grain-size distribution within the active layer is updated at every time step and recorded as stratigraphic slices for use if erosion occurs. Bed elevations, channel width, bed slope, and water depth are all calculated and recorded each time step.

For the purposes of this study, it is particularly important that the transport of heterogeneous sediments and the adjustment of hydraulic geometry be accurate. We elaborate on these below; the reader is referred to van Niekerk et al. (1992) for other details of MIDAS.

Transport of Heterogeneous Sediments

The question of mobility of different grain sizes in modern river systems (Parker et al. 1982; Ashworth and Ferguson 1989; Dietrich et al. 1989; Ferguson and Ashworth 1991; Bridge and Bennett 1992; Komar and Shih 1992; Wilcock and McArdell 1993; Paola and Seal 1995) can be reduced to two main questions: how well do small grains hide between and/or beneath larger grains and thus avoid entrainment, and how much more easily can large grains be entrained because they are more exposed in the flow? If hiding and exposure balance each other such that grains of all sizes are entrained in equal proportion to their availability on the bed so that the transported material has the same size distribution as the bed material, a condition termed equal mobility (Parker et al. 1982), then there should be no downstream fining.

Most observational data from modern streams (e.g., Andrews 1983; Ashworth and Ferguson 1989; Komar and Shih 1992) show that τ_{cij}, the fluid shear stress at the bed necessary to mobilize the ith size fraction, can be expressed as a ratio of the size of the individual grain, D_i, to the median grain size in the bed, D_{50}:

$$\tau_{c_{ij}} = \Theta_{c50}(\sigma_j - \rho)gD_i\left(\frac{D_{ij}}{D_{50}}\right)^{-m} \qquad [5]$$

where $\Theta_{c_{50}}$ is the empirically based critical Shields parameter for the median size and is a function of the Reynolds number ($\mathrm{Re_D}$) of a grain (Bridge 1981) but is constant at ~ 0.045 for $\mathrm{Re_D} > 70$, $(\sigma\text{-}\rho)$ is the immersed density, g is gravitational acceleration, and m is a hiding constant that is derived from empirical relationships of field-based data and flume studies (Komar 1987). Accumulation rates and downstream fining trends are sensitive to different m values. In the limit, an exponent of $m = 1$ in Eq 5 means that the critical shear stress for all sizes in a mixture is the critical shear stress of the median grain size (D_{50}) of the bed material (Parker 1990), and in such a scenario, no downstream fining occurs. Since equal mobility was first hypothesized by Parker et al. (1982) from a reanalysis of transport data from Oak Creek, Oregon, many studies of modern gravel-bed rivers have focused on calculating the value of m. Ashworth and Ferguson (1989) and Bridge and Bennett (1992) have demonstrated that $0.65 \leq m \geq 0.9$ includes most of the m values for a suite of modern rivers. Wilcock (1993) has demonstrated that strongly bimodal size distributions show significant deviation from equal mobility, particularly for coarse grain sizes, while weakly bimodal and log-normal distributions are more equally mobile (Wilcock and Southard 1988; Wilcock 1992). Bridge and Bennett (1992) further discuss potential problems with both data collection and data analysis methods in entrainment studies, and question the usefulness of Eq 5. Here we take the view that Eq 5 has been shown to adequately calculate grain-size distributions in several modern datasets (Vogel et al. 1992) and is quite applicable to low-resolution datasets typically derived from ancient river systems. In the present study m is varied between 0.65 and 0.85, which includes values from bimodal-gravel-bed rivers that are similar in character to the ancient rivers being simulated in this study.

Bed Shear Stress Adjustments and Hydraulic Geometry

An under-appreciated problem in modeling fluvial sediment transport and downstream fining is correctly specifying a dynamic hydraulic geometry of the channel. In numerical modeling of rivers, this is an important topic because the sediment transport equation sets that include bed shear stress need to be closed by an equation that relates bed shear stress to the other hydraulic variables of the channel. In a benchmark paper written nearly twenty years ago, Parker (1978) derived an equation for self-formed, noncohesive-gravel-bed rivers by assuming that there is lateral transport of turbulent momentum from regions of high momentum (center) towards regions of lower momentum (banks). Parker (1978) concluded that a stable channel should theoretically configure itself such that the bed shear stress at the center of the channel is about 20% in excess of the critical bed shear stress necessary to entrain the median grain size, or

$$\frac{\tau_0}{\tau_{c_{50}}} = 1 + \epsilon_p \qquad [6]$$

where ϵ_p is 0.2. Eq 6 states that channel width increases or decreases to alter bed shear stress (τ_0) and maintain a constant ϵ_p value. More recently, Paola et al. (1992a) determined ϵ_p to be 0.4 and used this relationship to close their equation sets (all future references will use ϵ, where $\epsilon = 1 + \epsilon_p$). It follows from Eq 6 that if $\tau_0 = rgRS$, and $\tau_{c_{50}} \propto D_{50}$ (i.e., for steady, uniform and hydraulically rough flows), then

$$\frac{\rho g R S}{0.045(\sigma - \rho) g D_{50}} = \epsilon \qquad [7]$$

or

$$\frac{RS}{D_{50}} = k \qquad [8]$$

where R is hydraulic radius. Thus, according to this point of view, downstream fining, in the absence of abrasion or sediment caliber changes from tributaries, can occur only where the product RS decreases downstream.

Figure 3A tests this relationship using the shear-stress data for 235 gravel- and sand-bed rivers (meandering, split, anastomosing, braided-anastomosing, and braided) from Church and Rood (1983). The two distinct populations represent gravel-bed (open circles) and sand-bed (closed circles) rivers; the lack of data between the two populations is typical of fluvial deposits and represents a grain-size "gap" where granules and coarse sand are absent. Bed shear stresses calculated from the numerator of Eq 7 and critical shear stresses calculated from the denominator of Eq 7 for hydraulic data measured after selected two-year floods or bankfull discharge relative to valley-flat elevation in anastomosing, braided-anastomosing, and braided gravel-bed rivers (Fig. 3B) do not show a statistically significant correlation, especially if the very coarse-grained outlier is removed. It is true, however, that certain individual rivers like the Peace River (Fig. 4) exhibit a fairly constant ϵ value of ~ 1.8 in their gravel reaches. Discharge, water depth, and channel width increase downstream, while grain size, slope, and the depth–slope product decrease (Fig. 4B), in accordance with Eq 8. But beyond the gravel–sand transition (hereafter called the gravel front), ϵ increases dramatically to ~ 36.

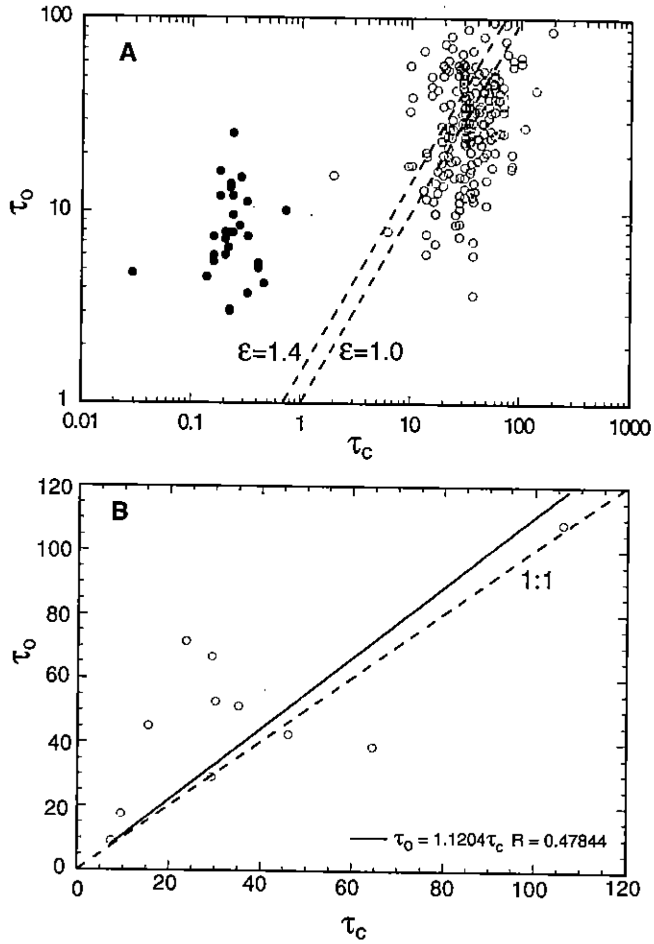

FIG. 3.—A) Excess shear stress relationship for 235 gravel-bed rivers (open circles) and sand-bed rivers (filled circles) for 2-year and 5-year bankfull flood conditions (data from Church and Rood 1983). B) Selected 2-year flood and bankfull discharge (relative to valley-flat elevation) data for "straight" channels.

In the Allt Dubhaig in Scotland (Sambrook-Smith and Ferguson 1995), the gravel–sand transition extends for a distance of ~ 25 m. On the basis of the hydraulic data of Sambrook-Smith and Ferguson (1995) and Ferguson et al. (1996), ϵ decreases from ~ 4 to ~ 1.5 at the gravel front over a distance of ~ 130 m, after which there is a fairly abrupt increase in ϵ as the bed becomes sand-dominated (Fig. 5). This short reach has a slight decrease in discharge toward the gravel front, a decrease followed by a steady increase in width toward the gravel front, and a decrease in width after the gravel front (Fig. 5A). The percentages of sand in the surface and subsurface both increase rapidly from ~ 75 m before the transition (Fig. 5B). Notice that on both the Peace River and the Allt Dubhaig, chan-

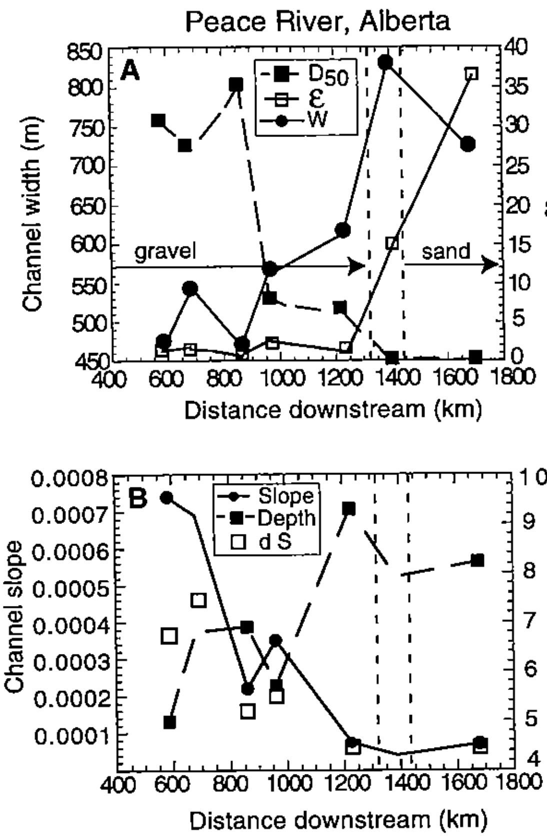

FIG. 4.—Downstream hydraulic data for the Peace River, Alberta. A) channel width, ϵ, and D_{50}; B) bed slope, water depth, and depth-slope product (data are from Shaw and Kellerhals 1982).

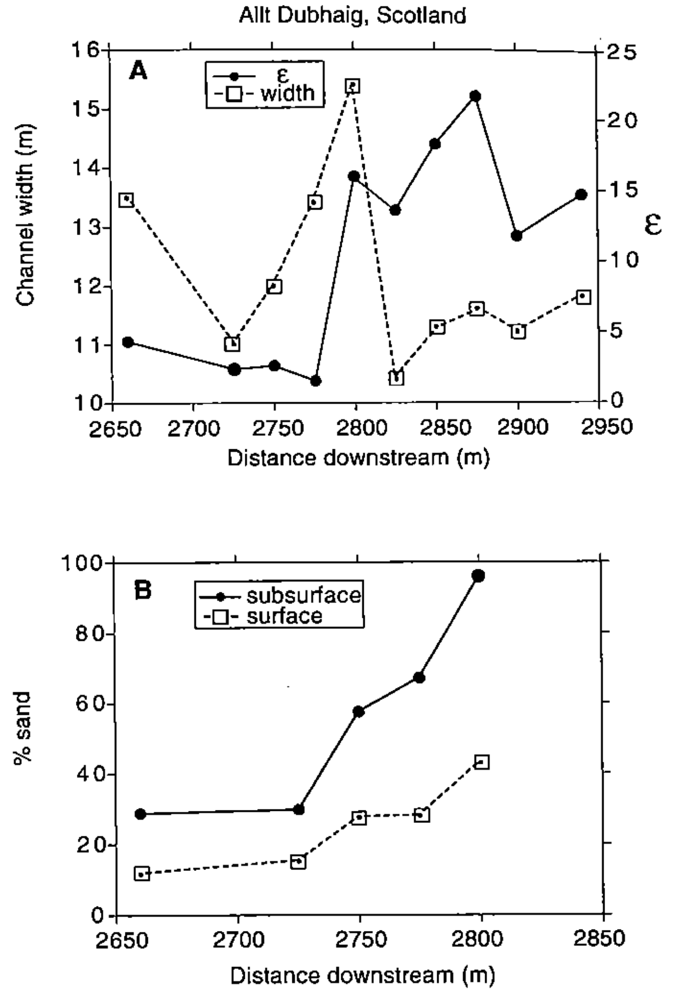

FIG. 5.—Downstream hydraulic data for the Allt Dubhaig, Scotland showing: A) channel width and ϵ over 300 m of the reach including the gravel–sand transition; B) Percentage sand in surface and subsurface over the gravel–sand transition. Data from Sambrook-Smith and Ferguson (1995) and Sambrook-Smith (personal communication 1996).

nel width increases more rapidly towards the gravel front and then decreases where the bed is sand-dominated (Figs. 4A, 5A).

These data suggest that while some rivers may maintain a relatively constant ϵ value in their gravel reaches many do not, and in all cases ϵ increases rapidly, width decreases (or the rate of change of width decreases), and the mobility of gravel drops when a critical sand percentage of the bed is reached. Eq 8 does not predict these changes.

Because the increase in ϵ and changes in width at the gravel–sand transition control downstream fining beyond this location, we explore other channel regime equations to compute channel width. Both empirical (e.g., Pizzuto 1992; Howard et al. 1994) and theoretical (e.g., Diplas 1990; Pizzuto 1990) hydraulic geometry equations have been incorporated into river evolution models before. Most empirical regime equations, e.g., Leopold and Maddock (1953), Park (1977), and as summarized in White (1988), are classified into sand- and gravel-bed rivers. Although the forms of the equations are similar for the two classes (Leopold and Maddock 1953; Lacey 1958), the proportionality constants and exponents are not, reflecting the different physical processes operating within each class. We have selected five regime equations (which are not dimensionally homogeneous) to investigate which relationship produces the best fit to changing ϵ values in natural gravel–sand transitions. The equations are:

(1) a standard empirical regime equation (e.g., Leopold and Maddock 1953);

$$W = aQ_w{}^b \tag{9}$$

(2) a combined Hey and Thorne (1986) empirical equation for water and sediment discharge that includes grain size and slope;

$$W = c(0.7833)Q_w^{0.479}\, S^{-0.1}\, D_{50}^{-0.009}\, D_{84}^{0.084} \tag{10}$$

(3) Bray's (1982) empirical equation that includes grain size;

$$W = dQ_w{}^{0.528}\, D_{50}^{-0.07} \tag{11}$$

(4) Parker's (1978) physically based equation for gravel-bed rivers that includes grain size and slope;

$$W = 6.86Q_w D_{50}{}^{-3/2} S^{1.217} + 2.23y \qquad [12]$$

and

(5) a simple resistance equation for sand-bed rivers;

$$W = \frac{Q_w}{\left[5.75\sqrt{gyS}\,\log\left(\frac{12.2y}{k_s}\right)\right]y} \qquad [13]$$

In the above equations, W is channel width (L), Q_w is bankfull discharge (L^3/T), S is bed slope, and D_{50} (L) and D_{84} (L) are the grain sizes that represent the 50th and 84th percentiles of the bed material, a and b are 2.7 and 0.5, respectively (see Yalin 1992 for a summary of observed a and b values), c is dependent on the amount of bank vegetation (Hey and Thorne 1986), d is equal to 3.83 (Bray 1982), k_s is grain roughness (L), which is

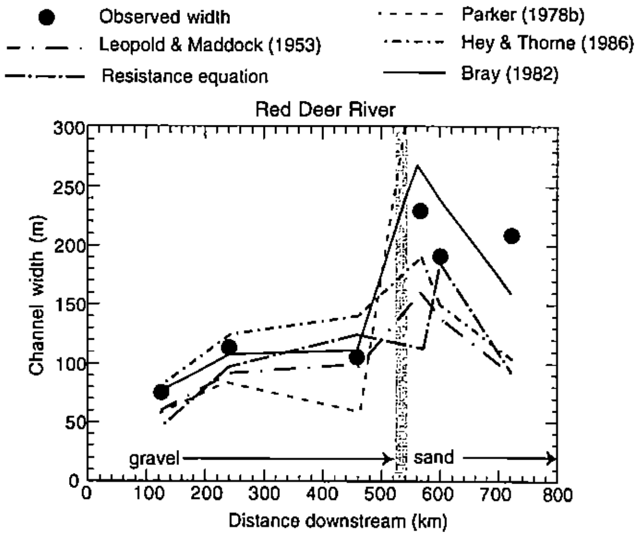

FIG. 6.—Calculated and observed widths for the Red Deer River, Alberta. Data from Shaw and Kellerhals (1982). See text for regime equations.

approximated as $2.5D_{50}$, and y is water depth (L). Because the second term on the right-hand side of Eq 12 is very much smaller than the first term, it is ignored here. The units for the right-hand side of each equation, assuming that a, b, c, and d are dimensionless, are L^3/T (Eq 9), $L^{1.344}/T$ (Eq 10), $L^{1.514}/T$ (Eq 11), $L^{0.5}/T$ (Eq 12), and L (Eq 13).

Figure 6 shows the widths calculated from Eqs 9–13 plotted with the observed discharge, width, depth, slope, and grain-size data for the Red Deer River (Shaw and Kellerhals 1982). The Shaw and Kellerhals (1982) dataset demonstrates how several hydraulic characteristics change downstream for several rivers; the Red Deer river is chosen to test each of the equations, but similar results are obtained for the other rivers.

Predicted values are within one factor of the observed widths for the gravel reach, but diverge downstream of the gravel–sand transition. Bray's (1982) equation is derived from 70 Canadian rivers, of which the Peace River is one, and so its good fit is not surprising. Nevertheless, it is the only equation that increases width in approximate proportion to the data values after the gravel–sand transition. Parker's (1978) equation for gravel-bed rivers overpredicts width for the sand-bed reach; the other equations underestimate width. On the basis of these results, we incorporate the hydraulic equations of Leopold and Maddock (Eq 9), Hey and Thorne (Eq 10), and Bray (Eq 11) into the sediment transport model, but we favor Bray's equation.

NUMERICAL EXPERIMENTS

Eight numerical experiments were designed to test the sensitivity of downstream fining to the controlling variables. Table 1 summarizes the various initial and boundary conditions; the volumetric ratios of water to sediment feed rate assume a sediment density of 2650 kg/m^3 to convert sediment mass to volume. It should be noted that because each simulation was run to equilibrium, the durations and accumulations are different in each case, because equilibrium time is a function of sediment feed rate, water discharge, subsidence rate, and basin length. The experimental input values are intended to represent ancient fluvial systems depositing material over geologic time scales. It is assumed that during 1 million years of simulated time, a bankfull discharge occurs, on average, once every 1.58 years for a duration of about 2–4 weeks (Dury 1973) and that a channel is active for only ~ 10% of the million years for an avulsion rate of 1000 years (Mackey and Bridge 1995) and a floodplain width of ~ 20 km (Hovius 1996). "Geologic time" is therefore compressed. Time to equi-

TABLE 1.—*Initial conditions for numerical experiments.*

Experiment Number / Variable	1) Hydraulic Adjustment	2) Width Coefficient	3) Mobility	4) Sediment Feed Rate
Water discharge, Q_w	50 m^3/s	50 m^3/s	50 m^3/s	50 m^3/s
Width equation	1.1) Eq 9 1.2) Eq 10 1.3) Eq 11	2.1) Eq 11 $d = 1.4$ 2.2) Eq 11 $d = 5$	Eq 11 $d = 3.83$	Eq 11
Water/sediment flux ratio	325	325	325	650
Size distribution	see Fig. 7	see Fig. 7	see Fig. 7	see Fig. 7
Sediment feed rate	407 kg/s	407 kg/s	407 kg/s	203.5 kg/s
Subsidence type and rate	exp; 40 m/myr	exp; 40 m/myr	exp; 40 m/myr	exp; 40 m/myr
Hiding coefficient, m	0.85	0.85	0.65	0.65
Length of reach ($\Delta x = 10$ km; $\Delta t = 0.5$ hr)	100 km	100 km	100 km	100 km
Figure reference	Fig. 7B, C, D	Fig. 8A, B	Fig. 9A, B	Fig. 10A, B
Experiment Number / Variable	**5) Water Discharge**	**6) Subsidence Rate**	**7) Pocono-Subsidence**	**8) Pocono-Sediment Feed**
Water discharge, Q_w	6.1) 100 m^3/s 6.2) 250 m^3/s	50 m^3/s	200 m^3/s	200 m^3/s
Width equation	Eq 11 $d = 3.83$	Eq 11 $d = 3.83$	Eq 11 $d = 3.83$	Eq 11
Water/sediment flux ratio	650	325	595	1188
Size distribution	see Fig. 7	see Fig. 7	see Fig. 13	see Fig. 13
Sediment feed rate	407 kg/s	407 kg/s	890 kg/s	446 kg/s
Subsidence type and rate	exp. 80 m/myr	exp. 80 m/myr	7.1 exp. 43 m/myr 7.2 lin. 43 m/myr 7.3 exp. 22 m/myr	exp; 43 m/myr
Hiding coefficient, m	0.65	0.65	0.65	0.65
Length of reach ($\Delta x = 10$ km; $\Delta t = 0.5$ hr)	100 km	100 km	250 km	250 km
Figure reference	Fig. 11A, B	Fig. 12A, B	Fig. 14A, B, C	Fig. 14D

librium is defined as the time it takes to deposit a volume of sediment exactly equal to the accommodation volume produced by a particular subsidence rate, with a particular combination of sediment feed rate, size distribution, water discharge, and basin length. The deposited volume is tracked through time, and at termination each experiment has reached 1% or less change in volume per timestep.

The Role of Hydraulic Geometry

In this experiment (Table 1; Experiment 1), we explore the extent to which the downstream fining of fluvial deposits is affected by the manner in which the river adjusts its hydraulic geometry. Figure 7 shows the results from Experiments 1.1, 1.2, and 1.3, representing the three different width equations noted in Table 1. Small differences in the hydraulic geometry equations cause significant differences in bed slope, sediment sorting, and textural distribution. The hydraulic equation of Leopold and Maddock (1953) has the smallest proportionality constant ($a = 2.7$) and produces the narrowest widths. Of the three equations, it produces deeper water depths and larger velocities at equilibrium. Compared to the results of Experiments 1.2 and 1.3, bed slope decreases less rapidly in Experiment 1.1 because less sediment is extracted from the flow at each point downstream (because width is much less). The grain sizes prograde farther downstream because the water depth is much deeper and bed shear stresses are large enough to make all available sizes on the bed equally mobile. Grain size is therefore similar everywhere downstream at equilibrium (Fig. 7B) and the large value of ϵ indicates poor sediment sorting relative to Experiments 1.2 and 1.3. The Hey and Thorne (Fig. 7C) and Bray (Fig. 7D) equations both produce wider widths with lower water depths, although their bed slopes are different. Bed slopes in the headwaters are less than in Experiment 1.1 because with increased channel width, the bed must

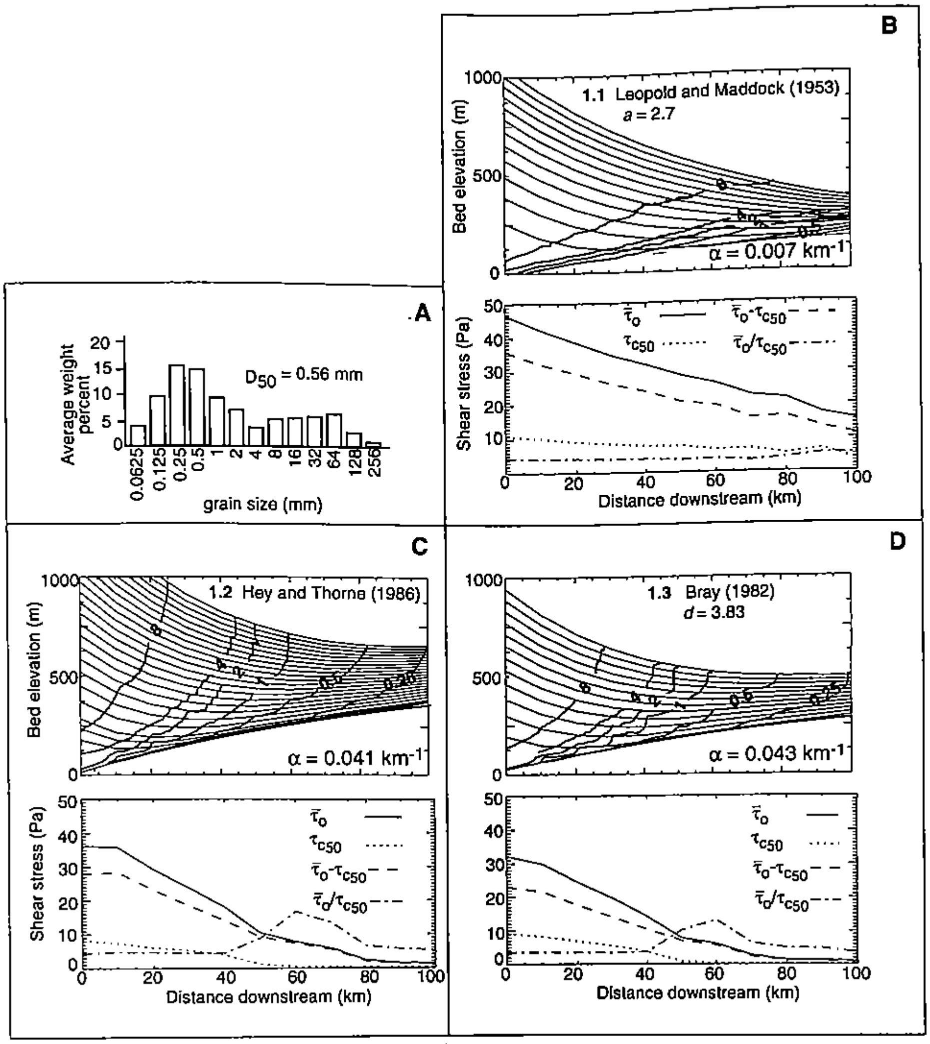

FIG. 7.—Results of Experiments 1.1, 1.2, and 1.3, which compare the influence of regime equations. **A**) Size distribution of feed. **B**) (upper panel) Bed elevation and grain-size contours versus distance downstream plotted every 200,000 timesteps (duration is 2.4 myr). Computation uses Eq 9. Fining rate for last timestep is in lower right corner. (lower panel) Various fluid shear stresses versus distance downstream for last time step. **C**) Same as B (duration is 4.4 myr) but computed using Eq 10. **D**) Same as B (duration is 2.4 myr) but computed using Eq 11.

steepen to transport the sediment at the feed rate of each size. With lower water depths, the bed shear stresses are within the range of critical bed shear stress for the individual grain sizes and therefore there is distinct downstream fining.

Longitudinal variation in excess shear stress ($\tau_0/\tau_{c_{50}}$) or ϵ is similar between Experiments 1.2 and 1.3, being relatively constant with distance in the upstream gravel reach at 4.5 and 3.5 in Experiments 1.2 and 1.3, respectively, although the values appear large for self-formed gravel bed rivers (Figs. 4 and 5). In the downstream sandy reach, ϵ has a relatively constant value of 4–5. In the gravel-sand transition, ϵ values peak at 14–18 because the bed slope is adjusted to transport the small amount of gravel remaining in transport and so is transitional between the higher bed slopes of the dominantly gravel reach and the low bed slopes of the sand reach. It is noted that the Hey and Thorne (1986) equation gives downstream fining results similar to the Bray (1982) equation although the time to equilibrium and bed slopes differ. We conclude that Eq 11 of Bray (1982) predicts width variations that produce the lowest ϵ value in the gravel reach, and it is used in all subsequent experiments.

To better understand the relationship between ϵ variations and downstream fining, Experiments 2.1 and 2.2 were conducted using Eq 11 of Bray (1982) with a smaller (Fig. 8A) and larger (Fig. 8B) proportionality constant, d. As in Experiment 1.1, a smaller magnitude of d in Eq 11 reduces channel width for the same water discharge and therefore increases water depth and sediment feed rate per unit width. Relative to Experiment 1.3, bed slope is less steep in the headwaters but decreases less rapidly (Fig. 8A). There is more progradation of grain-size contours downstream and larger at any particular distance downstream. Conversely, a larger d creates a wider channel for the same water discharge and a decreased sediment feed rate per unit width. The water depths are less and bed slopes decline more rapidly, although more aggradation occurs in the headwaters relative to the narrower width case. In comparison to the equilibrium conditions for Experiment 1.3 (Fig. 7D), the profile exhibits a finer D_{50}, an increased fining rate, and smaller ϵ (Fig. 8B). Although the water–sediment feed ratios (Q_w/Q_s) do not change in Experiments 1 and 2 (Table 1), the equilibrium downstream fining and bed slopes differ because sediment transport is a nonlinear function of bed slope, water depth, channel width, and grain size and because the sediment is distributed differently for each width.

In conclusion, reducing d in Eq 11 by 63% decreases downstream fining rate by 87% and increasing d by 30% increases downstream fining rate by 16%. Thus, downstream fining rate is sensitive to the magnitude of d, particularly for values of d less than 3.83. For the Bray (1982) equation, d values for natural gravel-bed rivers range from 2.08 (Yalin 1992) to 3.83 (Hey and Thorne 1986), but equations similar in form to Bray (1982) have d values that range from ~ 1 to ~ 4.5, and most of this difference is attributed to vegetation (Yalin 1992; Hey and Thorne 1986). In subsequent experiments, we set $d = 3.83$ (Bray 1982) because it is based on numerous data sets of rivers that transport the range of sediment sizes of interest to

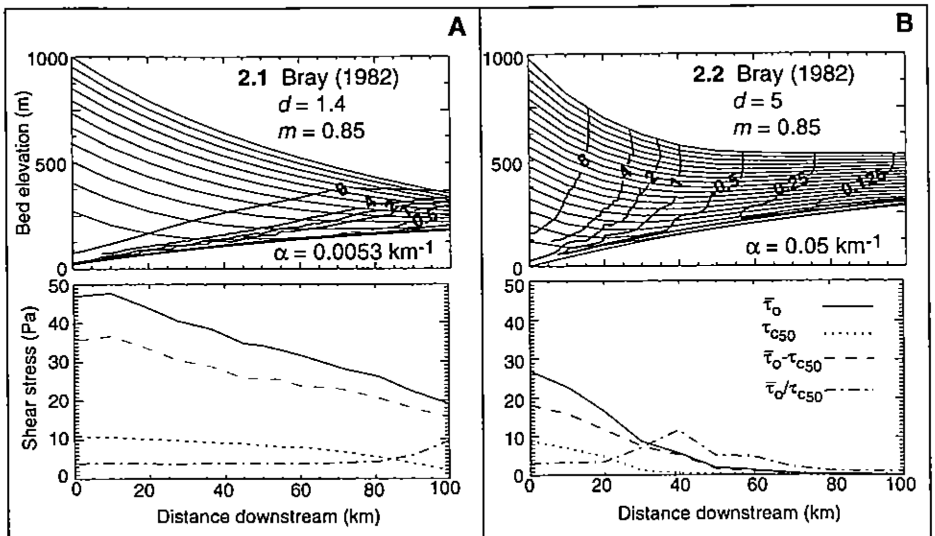

FIG. 8.—Influence of width proportionality coefficient. A) d = 1.4 (duration is 2.2 myr) and B) d = 5 (duration is 3.6 myr). Axes and panels are the same as in Figure 7.

this study. However, we acknowledge that the d value does influence downstream fining rates and therefore careful consideration should be given to the choice of its value. A d value of 3.83 still produces ϵ values that are 2–3 times larger than observed in modern gravel reaches (Figs. 4, 5) and approximately 2–3 times smaller than observed in sand reaches (Fig. 3). This mismatch is addressed in the next section.

The Role of Hiding Exponent

The influence of a smaller hiding exponent, m, is explored in Experiment 3. Changing m from 0.85 to 0.65 (Table 1) has little influence on downstream fining trends (Fig. 9A, B). A 23% reduction in m inversely affects downstream fining rate by only 9%. More importantly, reducing the absolute value of m reduces ϵ to values similar to those observed in Peace River (Fig. 4) and Allt Dubhaig (Fig. 5). A further reduction of m to 0.5 produces an ϵ of 1.45 in the upstream reaches (Fig. 9C). These changes arise because a smaller hiding coefficient increases the range of critical bed shear stresses, $\tau_{c_{ij}}$, necessary to entrain the individual sizes, thereby increasing the degree of sediment sorting. This affects the medium size the least, and the smallest and largest size fractions the most. At equilibrium, the bed material in the headwaters is relatively coarser and is larger (Fig. 9A, B). This coarsening slightly increases the rate of downstream fining and decreases ϵ (Fig. 9). Variations in m exert some effect on bed slope, which steepens to adjust to the decreased mobility of the larger grain sizes.

The use of a smaller hiding exponent is supported by the flume experiments of Wilcock (1993) who deduced the relationship between bimodality, m, and bed sorting. For our distribution, the standard deviation of the modes (Wilcock's $[D_c/D_f]^{0.5}$) and the proportion of material in the modes (P_m) coincide with a bimodality factor (B) of $\sim$ 4.5, corresponding to a hiding factor (m) of $\sim$ 0.6 (Wilcock, 1993). Therefore, for the size distribution, water discharges, and sediment feed rate used in this study, values of d = 3.83 and m = 0.65 best honor field observations of width and excess shear stress (e.g., Church and Rood 1983 and Ashworth and Ferguson 1989) and flume data for the entrainment of individual grain sizes in bimodal mixtures (Wilcock 1993). These values are used in all subsequent experiments.

The Role of Sediment Feed Rate

The influence of sediment feed rate on downstream fining is explored in Experiment 4, in which the feed rate is halved but all other variables are the same as in Experiment 3 (Table 1). Reducing sediment feed rate, and therefore reducing the amount of sediment that each reach must pass at equilibrium, reduces bed slope and therefore ϵ (Fig. 10). The sediment is trapped in the proximal reaches as the bed aggrades to keep up with the increase in accommodation space; subsequently, the basin is underfilled. Note that the values are similar in each case at the headwaters (Fig. 10), but the bed slope and bed shear stresses are smaller, and the downstream fining rate is 83% larger. This result highlights the necessity of correctly specifying sediment feed rate relative to water discharge, and we now explore this in more detail.

The Role of Water Discharge

Doubling water discharge and the ratio of water to sediment feed rate (Table 1; Experiment 5.1) increases channel width and decreases water depth, bed slope, and τ_0 (Fig. 11). This reduces ϵ. These changes occur because increasing water discharge has three effects: (1) channels widen, predominantly controlled by $3.83Q^{0.528}$, and therefore discharge per unit width increases only slightly; (2) water depths decrease, thereby decreasing τ_0; and (3) sedimentation is distributed over a wider channel, the deposit is thinner, and bed slope is less. In contrast to Experiment 2.1, where water discharge per unit width (q_w) was increased and progradation of the grain-size contours occurred, it is rather surprising that increasing water discharge (Q_w) does not cause progradation. Instead, downstream fining rate increases by 21% for an increase in water of 100%. A 400% increase in water discharge increases the fining rate by only 40% in Experiment 5.2 (Fig. 11B). However, bed slopes are reduced by 300%. The sediment feed rate per unit width is decreased in these two experiments as width increases to adjust to the larger water discharge, and the downstream reaches are actually sediment-starved (Fig. 11B). In summary, a larger Q_w/Q_s increases the downstream fining rate and produces a comparatively finer D_{50}, better sorting, and a significantly reduced equilibrium bed slope.

The Role of Subsidence Rate

Previous studies have shown that downstream fining patterns are affected by changes in subsidence rate (Paola et al. 1992a). To further explore this effect, we hold other variables constant and double subsidence rate at x = 0 (Table 1; Experiment 6). Because subsidence decreases exponentially downstream, there is relatively more sediment to be transported at each reach than in the case of reduced sediment feed (Experiment 4). Our results show that a larger ϵ value at equilibrium produces poorer sorting than in the case of reduced sediment feed rate, and consequently the bed material

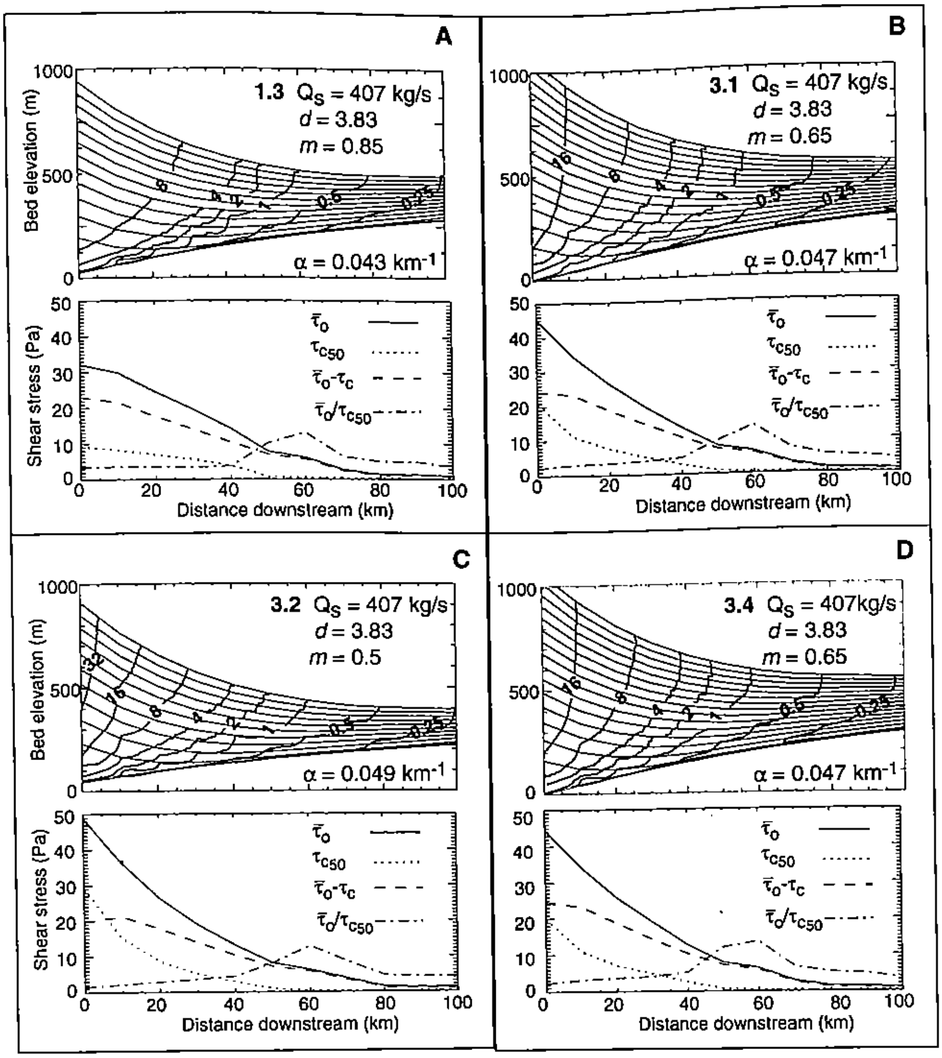

FIG. 9.—Influence of hiding coefficient m in Eq 5. A) m = 0.85 (duration is 2.4 myr), B) m = 0.65 (duration is 3.0 myr), C) m = 0.5 (duration is 2.0 myr), and D) same as in B for comparison. Axes and panels are the same as in Figure 7.

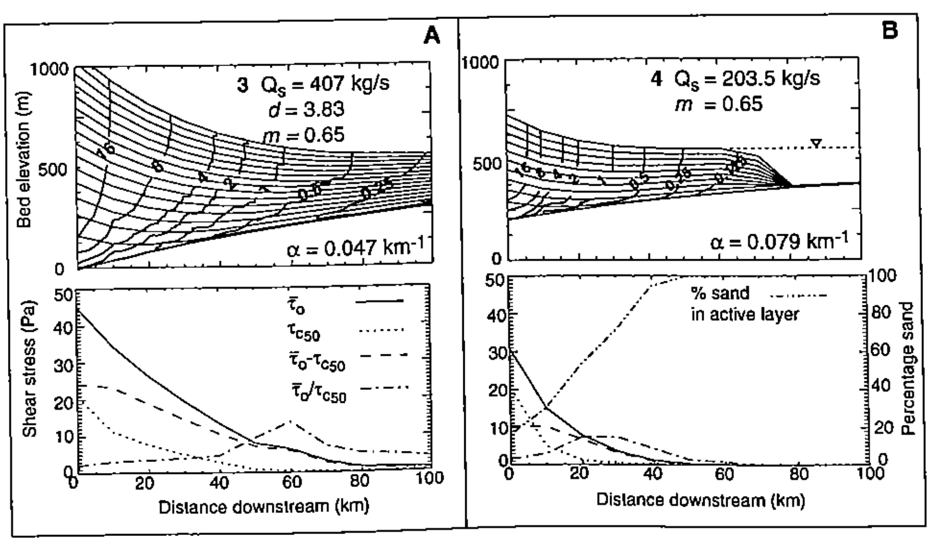

FIG. 10.—Influence of sediment feed rate. A) 407 kg/s (duration is 3.0 myr) and B) 203.5 kg/s (duration is 2.0 myr). Axes and panels are the same as Figure 7.

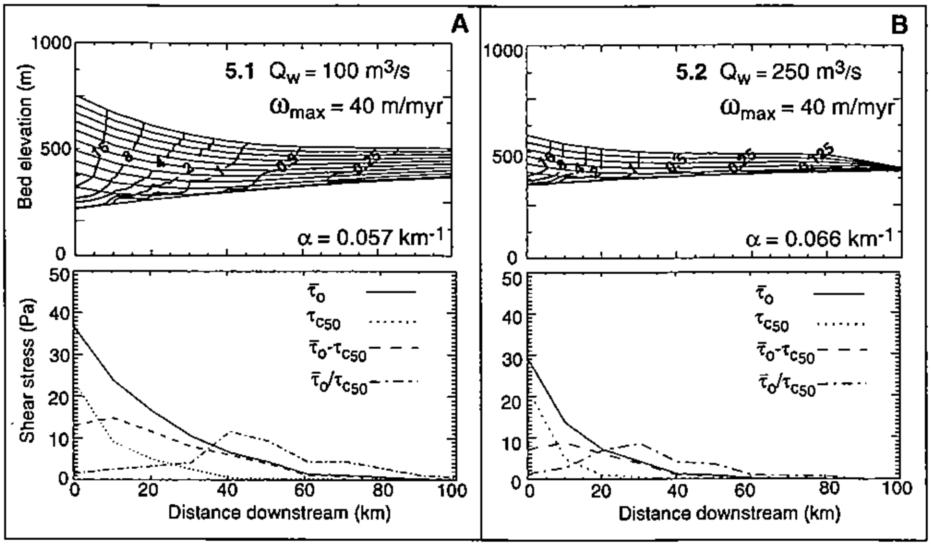

FIG. 11.—Influence of water discharge. **A)** 100 m³/s (duration is 2.0 myr) and **B)** 250 m³/s (duration is 1.2 myr). Axes and panels are the same as Figure 7.

is slightly finer in the proximal regions (Fig. 12), even though the bed slopes are very similar. The ϵ values are different because is almost halved in the subsidence rate experiment, even though the bed shear stresses are the same. The downstream fining rates are increased by 89% for the larger subsidence rate. Thus, on the basis of these two experiments, it can be concluded that downstream fining trends in aggrading modern and ancient rivers are about equally influenced by subsidence rate and sediment feed rate.

It is interesting that halving sediment flux or increasing water discharge does not affect values (lower panels of Figures 10B and 11), while doubling subsidence rate almost halves at equilibrium in the headwater region (lower panel of Figure 12B). On the basis of the previous experimental results, we can see that is most affected by the vertical aggradation rate. The case of decreased sediment feed rate (Experiment 4) is similar to the case of increased discharge (Experiment 5) because in both cases, the sediment feed rate per unit width is low and a thin layer of accumulation is spread over a relatively wide area. Larger subsidence (Experiment 6) and the original sediment feed rate produces increased vertical accumulation rate, thus influencing $\tau_{c_{50}}$.

In summary, these examples demonstrate that the magnitude of the effect on downstream fining rate differs amongst the controlling variables. The following hierarchy of decreasing importance is suggested from our results: (1) subsidence rate and sediment feed rate; (2) hydraulic geometry; (3) water discharge; and (4) hiding exponent. If d values smaller than about 2 are justified (i.e., for rivers with vegetated or cohesive banks), hydraulic adjustment is as important as subsidence and sediment feed rate because it increases the ϵ values and controls downstream fining trends. If water discharge controls channel width, then in self-formed rivers the effects of changing discharge on downstream fining are secondary to those of subsidence and sediment feed rate.

APPLICATION

Here we show how the preceding analysis can be used to interpret the hydraulic characteristics of an ancient fluvial system given the downstream fining of its deposits. The fluvial deposits in question crop out in Pennsylvania, where they are known as the Pocono Formation in the east (Figs. 1 and 13A; Berg and Edmunds 1979; Donaldson and Shumaker 1981;

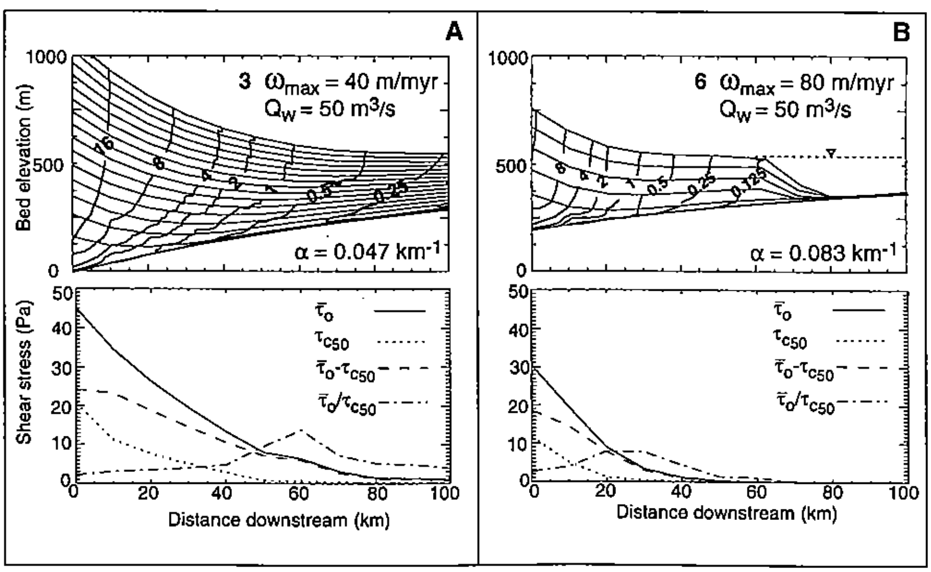

FIG. 12.—Influence of subsidence rate. **A)** 40 m/myr (duration is 3.0 myr) and **B)** 80 m/myr (duration is 1.0 myr). Axes and panels are the same as Figure 7.

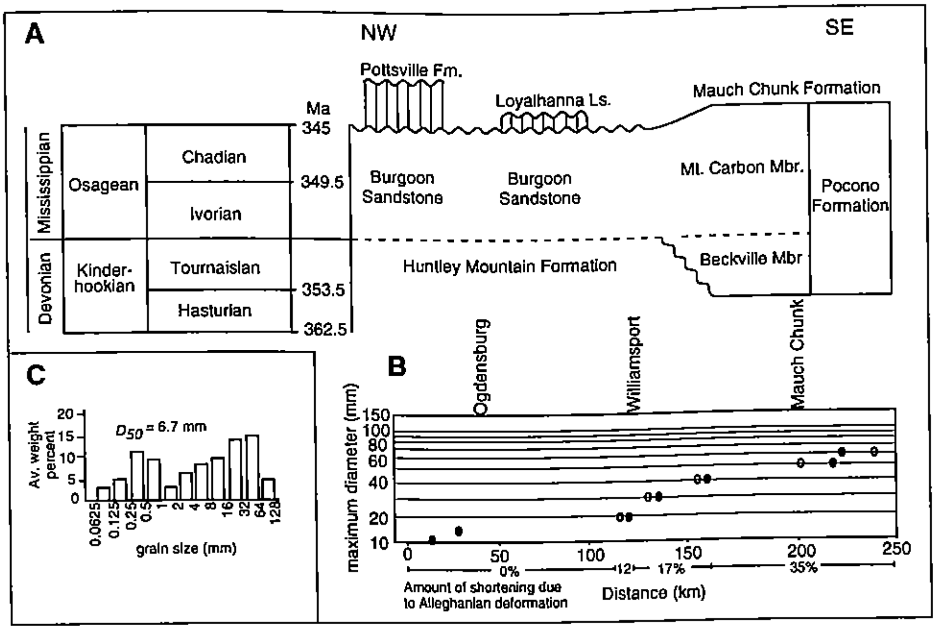

FIG. 13.—A) Stratigraphic correlation chart (modified from Edmunds et al. 1979). B) Pelletier's (1958) graph of downstream fining of largest observed pebble size. C) Simulation grain-size distribution of feed calculated from Pelletier (1957, 1958).

Slingerland and Beaumont 1989) and the Burgoon Sandstone in the west (Berg and Edmunds 1979) and are interpreted to be an extension of a broad alluvial plain that prograded westward during Early Mississippian time from a southeastern source terrain. Limited palynological and paleobotanical dating indicate that these rocks were deposited during the Osagean Age (Fig. 13A; Warg and Traverse 1973; Streel and Traverse 1978; Edmunds et al. 1979). The base of the fluvial Pocono Formation and Burgoon Sandstone is a conformable surface to the southeast near its source but passes basinward into a disconformity of regional extent (Fig. 13A). The top of the section is a regionally unconformable surface that in places represents a relative sea-level rise and elsewhere represents a later sea-level fall at the end of the Pennsylvanian (Sevon 1985).

The observed grain-size trend of the conglomeratic upper Pocono Formation and correlative pebbly sandstones of the Burgoon Sandstone is shown in Figures 1 and 13 across an approximately 250 km depositional dip section of the Acadian foreland basin from Mauch Chunk, Pennsylvania to Ogdensburg, Pennsylvania. The trend is exponential with a fining coefficient of 0.0075 km^{-1}, and a more rapid decrease in grain size occurs at about 100 km downstream from the 60 mm contour (Figures 1, 13B).

Experimental Design

We attempt to match the observed fining rate using forward modeling with estimates of sediment feed rate from the known accumulation of the Pocono deposit and hydraulic variables from modern rivers (Summerfield and Hulton 1994). The most important input values are sediment feed rate, water discharge, tectonic subsidence rate, and the grain-size distribution to be transported by the ancient rivers. Typical drainage basins in modern orogens range in area from 6×10^{11} to 6×10^{12} m^2, and transport solid loads that range between 1×10^1 and 1.5×10^2 m^3/s, with water discharges (runoff times drainage area) of 3×10^3 to 2×10^5 m^3/s (Summerfield and Hulton 1994). Therefore, the volumetric ratios of water to sediment discharge range between 10^2 and 10^3. Since drainage basin area is usually an unknown quantity in modeling ancient sedimentary deposits, discharges can be estimated from measured channel-width data, although this has not been attempted extensively here. An initial water discharge of 200 m^3/s, linearly increasing to 450 m^3/s, is chosen to match typical water–sediment feed ratios mentioned previously, and the approximate evolution of channel widths are calculated from Eq 11. We assume a clear-water discharge increase from tributaries; inherently, since the discharge increase is small, this assumes that the sediment feed accompanying that discharge is negligible. This is a necessary simplification to avoid a complicated input of different size distributions downstream. A constant sediment feed rate of 890 kg/s is calculated from the volume of deposit being simulated (i.e., the integrated areal decompacted thickness of the Pocono times the predicted widths over the "compressed" duration of the deposit). The decompacted sedimentary thickness was calculated using the method of Perrier and Quiblier (1974) with porosity data for the Pocono Sandstone (Griffiths 1967). The overlying material was divided into dated thickness slices using Beaumont et al.'s (1988) summary of stratigraphic packages and were iteratively removed using Perrier and Quiblier's (1974) average porosity curve method.

The grain-size distribution of the sediment feed (Fig. 13C) is calculated by integrating all the Pocono–Burgoon size distributions observed along the transect (Leonard 1953; Pelletier 1957), in a manner similar to Paola (1988) and is based on both maximum and median clast size data. The resulting distribution is quite high in the coarse sand and granule range (Fig. 13C). This probably occurs because the final distribution reflects only one (averaged) measurement from each observed bed and not the variability in each bed. With no dataset to the contrary, we assume that the sediment distribution and sediment flux are constant through time.

The increase in accommodation space is modeled as both a linear and exponential deflection of a beam with maximum subsidence at the headwaters, an obvious simplification of lithospheric response in a foreland basin (Flemings and Jordan 1989; DeCelles and Giles 1996). The basin wavelength is assumed to be 400 km (equivalent to an elastic thickness of $\sim$ 50 km and a flexural rigidity of $\sim 10^{24}$ Nm). A maximum subsidence rate and distribution was calculated by dividing the uncompacted sedimentary thickness of the Pocono–Burgoon by the duration of the Osagean (maximum of 43 m/myr). This value includes vertical aggradation, which is controlled by sediment feed rate and subsidence as well as generation of accommodation space, but serves as an initial estimate.

RESULTS

Four numerical experiments were conducted using ratios of water to sediment feed rates of 595 (Experiment 7.1; Fig. 14A) and 1188 (Experiment 8; Fig. 14D) with an exponential subsidence rate, a linear subsidence

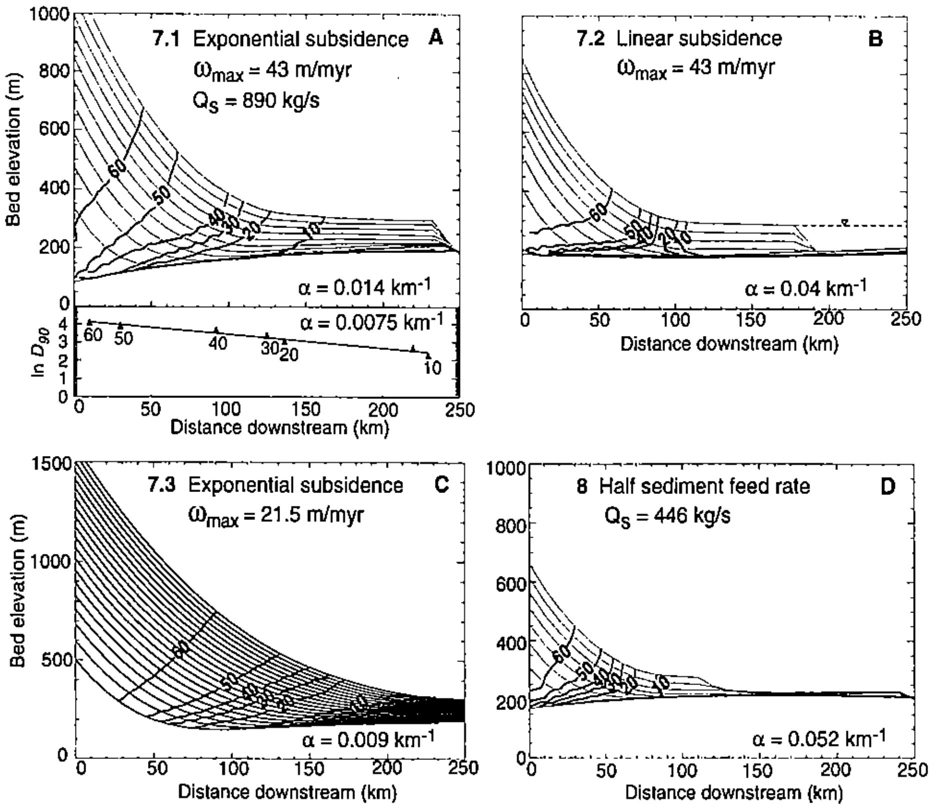

FIG. 14.—Pocono simulations and fining coefficients (α) using **A**) exponential subsidence rate calculated from observed stratigraphic thickness (duration is 4.4 myr). Grain size presented as D_{90} contours, in keeping with Pelletier (1958). Pelletier's (1958) data replotted as a natural log regression for comparison; **B**) linear subsidence (duration is 2.8 myr); **C**) half subsidence rate (exponential profile) (duration is 8.0 myr); and **D**) half sediment feed rate (duration is 2.8 myr).

rate (Experiment 7.2; Fig. 14B), and a decreased exponential subsidence rate (Experiment 7.3; Fig. 14C). Experiment 7.1 best fits the observed fining rate (Fig. 14A). Although the simulated downstream fining rate is larger than Pelletier observed, the general spacing of the grain-size contours is correct. Particularly, we have captured the grouping of the 50–40–20 mm contours as bed slope decreases rapidly (Fig. 14A). We have matched the spacing between the 60 and 50 mm contours although they have prograded too far into the basin. We have not captured the basinward position of the 10 mm contour, however (Fig. 14A). Pelletier (1957) noticed that the low sand/shale ratios in the northwest are misleading because the development of the upper unconformity probably removed some of the coarse material from this region. *Arthrophycus* trace fossils (Pelletier 1957) and more northwesterly marine faunas (Willard 1946) indicate marine influence in this region. Isostatic uplift towards the end of the Acadian orogeny (Slingerland and Beaumont 1989), marine processes, and transgressive reworking of the top surface of the Burgoon Formation could have influenced this region and its grain-size distribution.

A linear subsidence profile extracts more sediment to the bed than exponential subsidence, and therefore the fining rate is too large (Fig. 14B) and the basin is underfilled. Additionally, it does not capture the spacing between all the contours, which underscores how bed slope and grain size are affected by the rate of sediment extraction to the bed (in this case, due mainly to subsidence). Reducing subsidence rate (Experiment 7.3) has the effect of steepening the bed slope and prograding the sediment into the basin. Although the 10 mm contour now lies at about the observed position, it is at the expense of overly prograded 60 mm to 20 mm contours (Fig. 14C). A decreased ratio of water to sediment feed rate produces an underfilled basin and an overly large fining rate (Fig. 14D).

Our calculated bed slopes, ranging from 0.009 in the headwaters to 0.00007 at 250 km downstream, can be compared to slopes obtained by the method of Paola and Mohrig (1996). Our slopes are larger than predicted by their method, and the difference increases downstream. Since their equation for slope does not include subsidence, one would expect our values to be smaller. However, since sediment feed rate also influences bed slope and our ratio of $Q_w/Q_s Q_w/Q_s$ may be smaller than that used to define the excess shear stress (() equation (7) of Parker (1978), then the Paola and Mohrig estimated slopes will be lower. Additionally, the slopes diverge downstream because our *e* increases at the gravel front, whereas it is constant in their method.

On the basis of these results, we suggest that the downstream fining trends in the Pocono–Burgoon river deposits are compatible with a downstream decreasing subsidence rate of ~ 43 m/myr at the headwaters to ~ 15 m/myr at 250 km, and bankfull discharges that ranged from ~ 200 m^3/s to ~ 500 m^3/s downstream. These changes are associated with increases in channel width of 80 m to 200 m downstream, and bed slope decreases of 0.009 to 0.00007 over 250 km from present day Mauch Chunk, Pennsylvania to the Pennsylvania–New York border.

EVALUATION OF RESULTS

How good are these results? One test is to compare our predicted subsidence rate of 43 m/myr to calculations of subsidence necessary to accumulate the preserved sedimentary deposits of the Appalachian basin. Beaumont et al. (1988) predicted a maximum subsidence rate of 41 m/myr (outboard of the estimated load) for the Early Mississippian, very similar to our estimated rate of 43 m/myr. There are no published data on channel width for the Pocono Formation. Limited measurement of channel widths in central Pennsylvania, representing about 150 km downstream in our simulations, have a maximum width of 160 m and a mean of 117 m, which compares fairly well with our calculated value of 146 m from Eq 11. Given the results of our sensitivity studies, we know that hydraulic geometry, sediment feed rate, and subsidence rate influence downstream fining trends and therefore we conclude that our results would be more robust if downstream variations of channel width were available.

However, our sensitivity studies also allow us to bracket how inaccurate our results could be if water–sediment feed rate ratios and hydraulic geometry are different. If we are wrong by a factor of two on water discharge, our fining rates would be about 0.017 km^{-1} for an increase in discharge and 0.0115 km^{-1} for a decrease in discharge; a fourfold reduction would decrease the fining rate to 0.009 km^{-1} but, as in Experiment 7.3, would not give the appropriate spacing seen in the observed grain-size contours (Fig. 14A, B). Increasing water discharge twofold would additionally reduce bed slopes by ~ 56%, producing values closer to those predicted by the Paola and Mohrig (1996) method. If the d value for the hydraulic geometry equation decreases (and f increases), fining rates decrease quite dramatically after values smaller than ~ 2 (reaching 0.0018 km^{-1} for d = 1.4). It is suggested that improved results can be attained by simulating different downstream subsidence distributions, including sediment feed from tributaries, or by obtaining better data on channel width and grain size.

CONCLUSIONS

Our sensitivity studies demonstrate that varying sediment feed rate and subsidence rate exert the strongest influence on downstream fining trends as long as the evolution of hydraulic geometry is adequately accounted for. The effect of varying water discharge on fining rates is balanced by increases in channel width. Scaling water discharge by sediment feed rate reduces the role each plays in determining fining rate, but does not totally remove their influence. The sensitivity studies presented here emphasize the interactions among hydraulic geometry, sediment transport mechanics, and grain-size trends in modern and ancient alluvial settings. Our results demonstrate that in the presence of subsidence, reducing sediment feed rate increases fining rate and produces finer bed material, in contrast to the results of Dietrich et al. (1989) and Hoey and Ferguson (1994). Increasing water discharge has a similar but reduced effect. These contradictions arise because in our study, channel width and bed slope are allowed to dependently vary with water discharge and sediment feed rate.

Despite the stated apparent similarity between physically based and empirical regime equations (see Hey et al. 1982), these equations produce different equilibrium bed slopes and downstream fining patterns because of the sensitivity of sediment transport to small changes in e. We have used the "best-fitting" hydraulic adjustment equation to calculate channel width. With a reasonable (but lower bound) value for the hiding coefficient (m) or a large ratio of water discharge to sediment feed rate, the Bray (1982) equation gives e values comparable to modern gravel-bed river data observed by Paola and Seal (1995).

These results are in general agreement with earlier studies concerning the role of subsidence on gravel deposition in foreland basins (Heller et al. 1988; Paola 1988; Burbank et al. 1988; Heller and Paola 1989; Paola et al. 1992a; Heller and Paola 1992). Our contribution has been to quantify its effects relative to river hydraulics, sediment and water feed rates, and accumulation rate. We have argued that the grain-size trends observed in the Mississippian upper Pocono Formation and Burgoon Sandstone of the Appalachian foreland indicate a maximum subsidence rate of 43 m/myr over the interval of deposition. Typical channel widths of Pocono streams ranged from 80 m to 200 m and bed slopes decreased from 0.009 in proximal reaches to 0.00007 about 250 km downstream. Present-day streams with these characteristics have bimodal sediments and are relatively straight (i.e., the Red Deer and Bow–South Saskatchewan rivers).

ACKNOWLEDGMENTS

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