Transience and persistence in the depositional record of continental margins

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[1] Continental shelves and coastal plains are large persistent depositional landforms, which are stationary (nonmigrating) at their proximal ends and characterized by relatively steady long-term growth. In detail, however, their surface form and stratigraphic record is built of transient freely migrating landscape elements. We derive the timescales of crossover from transient to persistent topographic forms using empirical scaling relations for mean sediment accumulation as a function of averaging time, based upon tens of thousands of empirical measurements. A stochastic (noisy) diffusion model with drift predicts all the gross features of the empirical data. It satisfies first-order goals of describing both the surface morphology and stratigraphic completeness of depositional systems. The model crossover from noise-dominated to drift-dominated behavior corresponds to the empirical crossover from transport-dominated (autogenic) transient behavior to accommodation-dominated (subsidence) persistent behavior, which begins at timescales of 10^2–10^3 years and is complete by scales of 10^4–10^5 years. Because the same long-term scaling behavior emerges for off-shelf environments, it is not entirely explicable by steady subsidence. Fluctuations in sediment supply and routing probably have significant influence. At short-term (transient) scales, the exponents of the scaling relations vary with environment, particularly the prevalence of channeled sediment transport. At very small scales, modeling sediment transport as a diffusive process is inappropriate. Our results indicate that some of the timescales of interest for climate interpretation may fall within the transitional interval where neither accommodation nor transport processes are negligible and deconvolution is most challenging.

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1. Introduction

[2] The continental shelf of passive margin settings and the adjoining coastal plain form a large, long-term, leaky trap for terrigenous sediment. At scales of millions of years, the trap grows by persistent tectonic subsidence of the continental margin, augmented by the growing load of sediment. In detail, the aggrading top of the trapped pile of sediment responds to a cascade of transitory events and unsteady processes. Severe weather events, biotic changes, and climate fluctuations, for example, combine with the internally generated (autogenic) variability in deposition and erosion [*Lyons*, 2004; *Kim et al.*, 2006] to generate a spectrum of transient landscape elements. These are the ripples, dunes, sand waves, channels, and bars, whose migratory dynamics can be directly monitored on human timescales. With long-term net deposition, these features determine the form of buried sediment layers that build up larger and more persistent landscape elements such as fans, deltas and whole shelves. At what scale does transience give way to persistence in this landscape? The answer lies beyond direct observation but is recoverable from the nested hierarchy of beds and bedding planes in the buried stratigraphic record. As revealed from outcrop scale to seismic profiles of whole shelves, the bedding planes are a patchwork record of former landscape surfaces.

[3] Stratigraphers recognize fractal properties in the vertical succession of sedimentary layers and buried bedding planes [*Plotnick*, 1986; *Korvin*, 1992; *Schlager*, 2004; *Bailey and Smith*, 2005]. In particular, mean accumulation rates have a strong negative dependence on the averaging time [*Gilluly*, 1949; *Reineck*, 1960] that can be described by power law functions [*Sadler*, 1981, 1993]. The marine shelf maintains shallow depths for periods that encompass enormous numbers of beds. Because the bedding surfaces form during intervals of erosion or nondeposition, the beds between them must have accumulated at rates of fill that exceed the long-term trapping rate (accommodation) of sediment at the seafloor. Individual beds record transient surface features; in aggregate, they record more persistent features. The transition to persistence may be sought in the scaling relations for mean accumulation. It seems reason-

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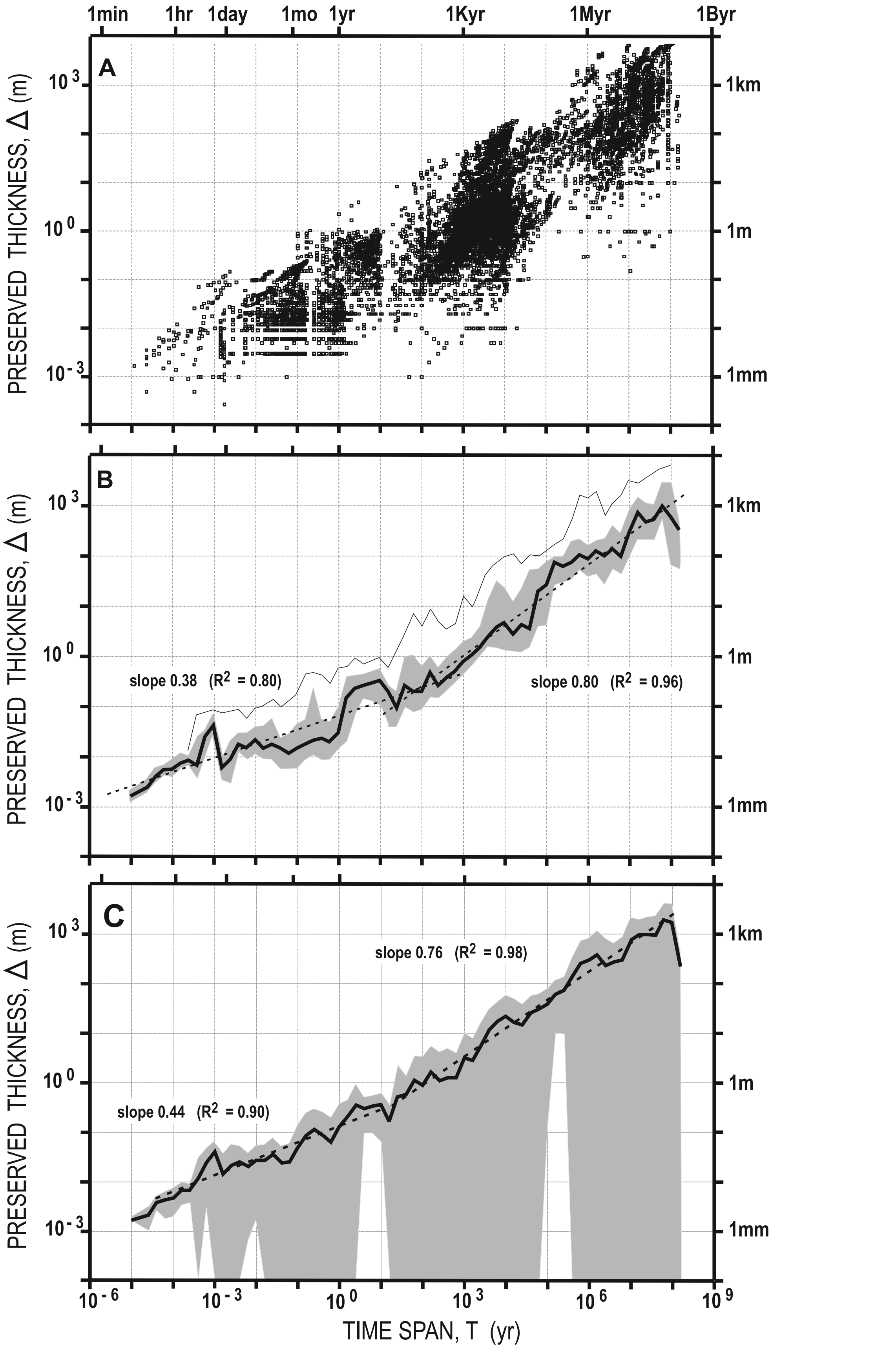
1min
1hr
1day
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1Kyr
1Myr
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A
B
C
PRESERVED THICKNESS, Δ (m)
10^3
10^0
10^{-3}
1km
1m
1mm
slope 0.38 (R^2 = 0.80)
slope 0.80 (R^2 = 0.96)
slope 0.44 (R^2 = 0.90)
slope 0.76 (R^2 = 0.98)
10^{-6}
10^{-3}
10^0
10^3
10^6
10^9
TIME SPAN, T (yr)

Figure 1

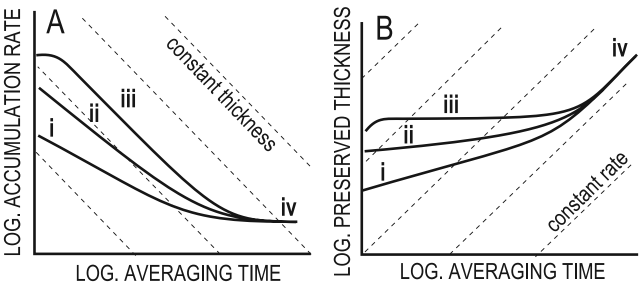

Figure 2. Comparison of scaling curves for plots of (a) mean net accumulation rate, $\partial\eta/\partial t$, and (b) mean net preserved thickness, Δ, against averaging time, T. Model curves are for (curve i) random walks [e.g., *Strauss and Sadler*, 1989]; (curve ii) noisy diffusion [e.g., *Pelletier and Turcotte*, 1997]; and (curve iii) sinusoidal models [e.g., *Sadler and Strauss*, 1990; *Sadler*, 1999]. Each model is added to a steady trend or "drift" (curve iv), to which mean values roll over in the long term.

able to expect that a single model may be able to account for the both the morphology of the surface landscape and the incompleteness of sedimentary deposits built by the aggrading surface.

[4] We present empirical scaling relations that summarize thousands of published measurements of sediment aggradation on coastal plains and marine shelf bottoms (e.g., Figure 1). Single power laws do not satisfactorily fit the full range of these data. Rather, there is a rollover between two asymptotic regimes with different slopes at the long-term and short-term ends. We recognize corresponding crossovers in the dominant processes and in the topographic forms. With increasing timescale, the determinative process changes from sediment transport to accommodation. The surface topography changes from transient fully mobile bed forms (e.g., ripples, bars, and leveed channels, which have both erosional and depositional portions) to more persistent forms that are anchored at their upstream ends (e.g., fans, deltas and whole shelves).

[5] The following sections first explain the nature of the empirical data, then review three approaches to modeling such data: one from sedimentology, another from stratigraphy, and a third that combines elements of both. We show how a noisy diffusion model, previously fit to short-term floodplain deposits [*Pelletier and Turcotte*, 1997], can reproduce the rollover at longer time spans. By comparing floodplain data with those from marine shelf and deeper off-shelf environments, we gain further insight from shortcomings of the model.

2. Data and Model Representation

[6] Power laws that describe stratigraphic, geomorphic, and paleontologic change have traditionally been illustrated by plotting mean rates of change (e.g., sediment surface aggradation rate, $\partial\eta/\partial t$) against the time spans (T) of measurement [*Sadler*, 1981; *Gingerich*, 1983; *Gardner et al.*, 1987]. Plotting a fraction (length/time) against its denominator (time) upset some researchers, but it effectively revealed that no single rate measurement could reliably be used as a general conversion factor between observed change (e.g., preserved stratigraphic thickness) and elapsed time for an unsteady process. With enough measurements to determine the slope of the power law, however, real insight about the nature of the unsteadiness may emerge [*Sadler and Strauss*, 1990; *Sadler*, 1999]. In order to avoid spurious correlation coefficients [*Kenny*, 1982] and the condensation of the length scale [*Sheets and Mitchell*, 2001], we present plots of the primary variables: preserved sediment thickness (or elevation change in the case of accommodation processes), Δ, against time span. The exponents of power laws in the two graphical formats have a straightforward relationship (Figure 2). Add 1.0 to the exponent of the rate–time span plot to determine the corresponding exponent on the length–time span plot. For example, the −0.5 slopes that characterize stationary random walks on a rate–time span plot ($\partial\eta/\partial t \sim T^{-0.5}$) transform to a +0.5 gradient on the length–time span plot ($\Delta \sim T^{0.5}$).

3. Empirical Scaling Relations

3.1. Nature of the Empirical Data

[7] Figures 1 and 3–5 summarize empirical estimates of scaling relations for sediment accumulation and accommodation at the continental margin. They are part of an active database project that currently holds about 350,000 rate estimates for processes crucial to understanding the construction of the sedimentary stratigraphic record [*Sadler*, 1994, 1999]. Although a few measurements may suffice to indicate that a power law likely describes the relationship of mean accumulation and averaging time, individual rate measurements vary considerably even within a narrow window of averaging time. Consequently, huge compilations are required to achieve stable estimates of the scaling exponents. Data must be combined from numerous localities and subdivision of the data by environment or process must remain very coarse. The resulting plots describe the average stratigraphic record for major subdivisions of continental margins. The actual records at real places will surely depart from these averages which serve as baseline expectations and a target for general models. A few more comments about the nature of these data are in order as a check against overinterpretation.

[8] Measurement method and ease vary with time span. Most estimates for time intervals from seconds to a century come from direct observations or the comparison of historic charts. Radiocarbon dates support estimates for periods from centuries to about 50,000 years. From this scale to intervals of a few 100,000 years, thermoluminescence dates

Figure 1. Empirical data for sediment accumulation on marine shelf sea bottoms. (a) Scatterplot of individual measurements. (b) Median (thick line) with 95th (thin line), 75th (top of gray band), and 25th percentiles (base of gray band) values. (c) Arithmetic mean (thick line) and one standard deviation (gray band) values. Mean, standard deviation, and percentile values in Figures 1b and 1c are determined for five time span units per logarithm cycle, with no overlap (i.e., no smoothing). Mean and standard deviation are determined for untransformed values (not their logarithms).

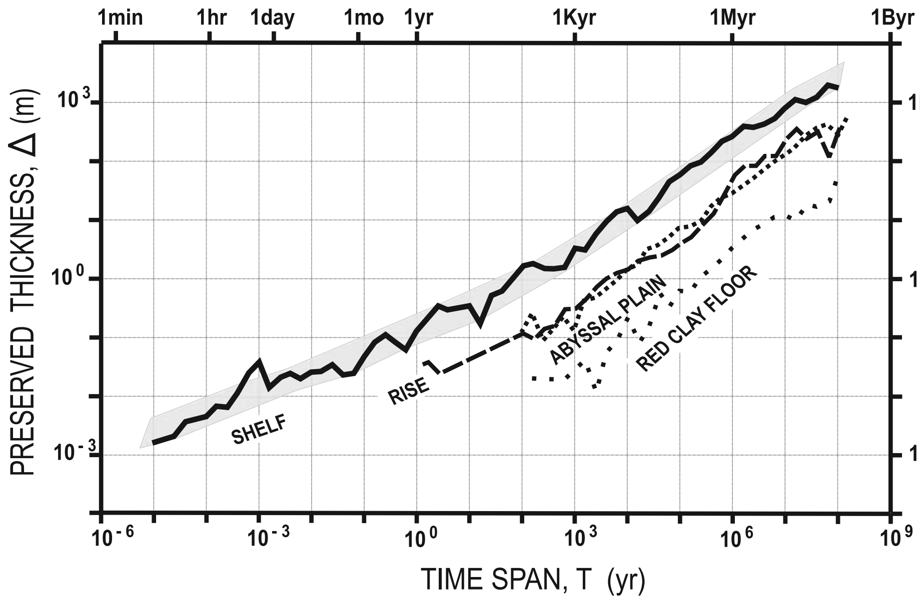

Figure 3. Empirical scaling relations for mean aggradation on marine shelf systems compared with off-shelf accumulation on continental rises, the abyssal plains, and the remote ocean floors where red clays accumulate. Arithmetic means are determined for five nonoverlapping bins per log cycle; gray envelope is fit subjectively to mean shelf values, exclusive of outliers. Data sources are as discussed by *Sadler* [1993, 1999].

and calibrated isotopic fluctuations are the most common source of age information. Calibrated paleomagnetic reversals are useful from these scales to tens of millions of years, where they become increasingly supplemented by biostratigraphy calibrated using K-Ar and U-Pb ages. Excluded are rates based solely on assumptions about uniform periodicity in cyclic sediments and age estimates that assume steady accumulation.

[9] Data are relatively sparse at scales of 50,000 to 200,000 years (Figure 1a), which fall near the limits of readily available dating techniques. For continental shelf sediments, intervals of 20–50 years are also underrepresented, but this difficulty vanishes for alluvial plains, which are more readily accessible to direct observation. Data become sparse a both extremes: there is a short-term limit to useful measurements and a long-term limit to the persistence of continental margin environments. The short-term

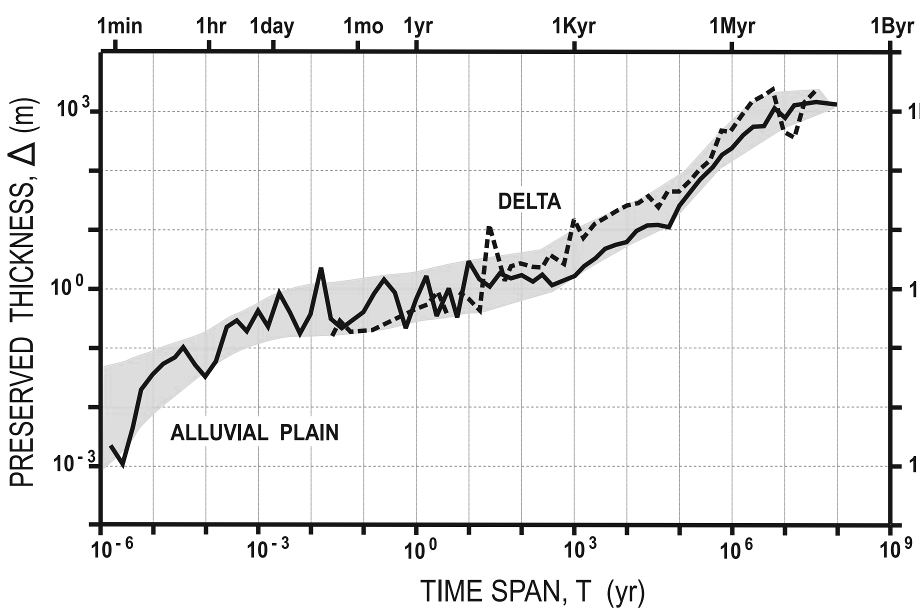

Figure 4. Empirical scaling relations for mean aggradation on channeled coastal plains (solid line) and deltas (dashed line). Means are determined for five nonoverlapping bins per log cycle. Gray band is fit subjectively to include all but outlier means. Data sources are as discussed by *Sadler* [1993, 1999].

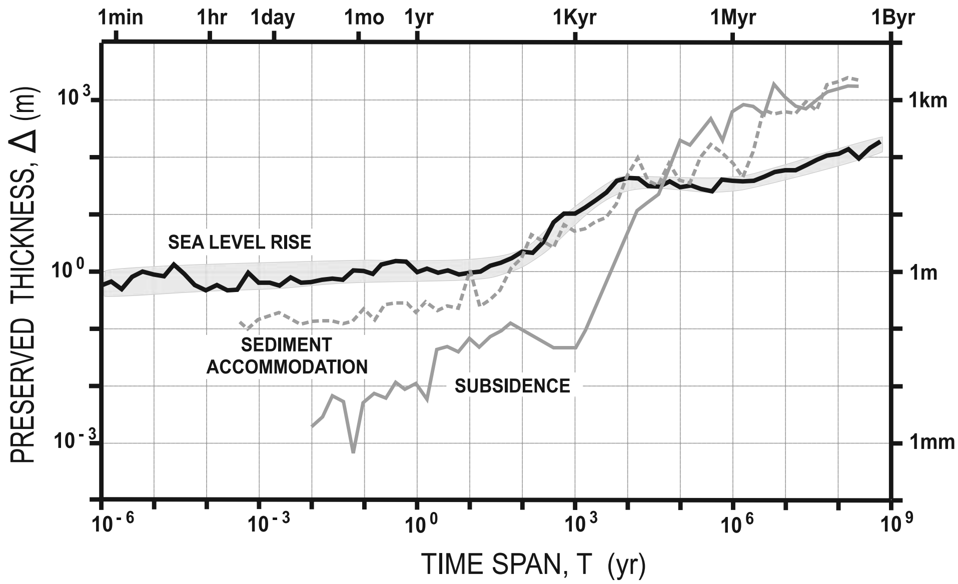

Figure 5. Empirical scaling relations for mean sea level rise (black line and light-gray band), net sediment accommodation determined from the depth of peritidal deposits (dashed gray line), and subsidence (solid gray line). Lines and band are determined as in Figures 3 and 4. Data sources are as discussed by *Sadler* [1994].

limit of practical measurement tends to increase as water depth increases and as accumulation rate decreases. Obviously, zero rates and immeasurably slow rates are underrepresented at all scales, but to what extent? That question motivated the original compilation effort: the significance of the zero rates at any timescale is seen in the reduced mean values at longer timescales. Intervals of near-zero or negative accumulation (bedding surfaces) at one scale lie within the averaging window at longer spans. They explain the negative association of mean accumulation with time span which leads to estimates of the incompleteness of the stratigraphic record [*Reineck*, 1960; *Sadler*, 1981]. In order to manage the changes in data density and source from one time span to another, our analyses are based on arithmetic means of the observed values in narrow logarithmic bins.

[10] Time span cannot be fully independent of age in these data. Closely spaced, high-precision dates are best obtained from young rocks. Measurements for long time intervals necessarily include old rocks. The Holocene is overrepresented in the short-term data. It has been a time of relative sea level rise in which the least amounts of terrigenous sediment bypass the shelf system into deeper water. For very short-term processes, accommodation is not an issue and the bias becomes irrelevant. Time spans of 5–15 kyr are resolved in the Pleistocene, so data capture a full range of sea level dynamics. The weak dependence of time span upon age cannot explain the first-order trends in the data [*Sadler*, 1981, 1994].

3.2. Scaling Relations for Accumulation

[11] For steady sediment accumulation described as a constant (nonfluctuating) drift, the cumulative sedimentary thickness would increase linearly with time ($\Delta \sim T^1$). This case would represent an idealized basin having steady subsidence where deposition is equal to accommodation everywhere. We might expect natural depositional systems at equilibrium to tend toward this behavior. Unsteadiness in deposition causes deposit thickness to increase more slowly than linearly with time ($\Delta \sim T^{<1}$), where the nature of the unsteadiness determines the scaling exponent. In the regime where fluctuations dominate, we expect such a less-than-linear scaling (see below). We anticipate then that at least two processes, fluctuating transport and subsidence, exert influence on the accumulation of sediment at short and long times, respectively (Figure 2). Such a process transition might be reflected in the natural data of sediment accumulation in basins.

[12] No single diagram can resolve all aspects of these large empirical data sets. We provide an illustrative suite of diagrams for the marine shelf data (Figure 1). Figure 1a shows the clustering of the measurements that is imposed by uneven opportunities for dating. The frequency distribution of values within these dense clusters is better represented by median, percentile, mean and standard deviation values (Figures 1b and 1c). Even though means are more sensitive to extreme values than medians, the trend of arithmetic mean values generally parallels the 95th, 75th, and 50th percentile contours. Other environments and processes are summarized by the mean values only (Figures 3–5).

[13] The empirical data plot as lines that curve gently, but significantly. Single power law trend lines fit these data with high correlation coefficients, however the residual misfit is not random. Rather, the empirical trends are systematically steeper at the long-term end than at the short-term end (Table 1). An acceptably random distribution of residuals is most simply achieved by fitting two power law segments with a curved crossover at scales from 10^2 to 10^4 years (e.g., Figure 1). The exponents of the pairs of trend lines fitted to the data are sensitive to the selection of the crossover time span. Nevertheless, they place limits on the gradient of asymptotes that will be used to evaluate different numerical models of accumulation. In the long-term limit, models

Table 1. Summary of Model Power Law Exponents and Empirical Regressions on Plots of Mean Recorded Thickness Against Averaging Time Span

Environment/Process	Short Term		Long Term		Overall		Data Count
	Exponent	R^2	Exponent	R^2	Exponent	R^2	
Abyssal red clay			0.83	*0.95*			2215
Abyssal plain			0.85	*0.99*			5638
Continental rise	0.37	*0.62*	0.86	*0.97*	0.65	*0.94*	6129
Continental slope			0.80	*0.98*			8202
Shelf	0.44	*0.90*	0.76	*0.98*	0.61	*0.98*	9632
Shore	0.23	*0.56*	0.78	*0.77*	0.43	*0.88*	10942
Delta	0.44	*0.73*	0.75	*0.91*	0.63	*0.94*	2994
Floodplain	0.48	*0.96*	1.00	*0.84*	0.74	*0.91*	1169
Channeled alluvial plain	0.17	*0.51*	0.86	*0.97*	0.46	*0.89*	11445
Total							58366
Shelf and shore	0.30	*0.84*	0.81	*0.98*	0.47	*0.93*	28776
Shelf, shore, and coastal plain	0.37	*0.97*	0.78	*0.98*	0.48	*0.93*	41390
Subsidence			0.88	*0.95*			6998
Sediment accommodation	0.10	*0.14*	0.70	*0.92*	0.57	*0.87*	10339
Sea level change	0.02	*0.05*	0.24	*0.82*	0.20	*0.70*	77674
Models							
Random walk with drift	0.50		1.0				
Noisy diffusion with drift	0.25		1.0				
Fixed diffusion with drift	variable		1.0				
Sinusoid with drift	1.0–0.0		1.0				

must have a gradient steeper than about 0.75 ($\Delta \sim T^{>0.75}$), perhaps a steady trend ($\Delta \sim T^{1.0}$ and $\partial\eta/\partial t$ = constant). This long-term exponent varies little between environments; *Sadler* [1981, 1994] attributed it to accommodation. The short-term limit may be no steeper than about 0.45 ($\Delta \sim T^{<0.45}$), with short-term exponents varying significantly with environment.

3.2.1. Marine Shelf and Distal Environments (Figures 1 and 3)

[14] The data for marine shelf systems combine measurements from a broad spectrum of landscape elements that are traditionally categorized as a function of their size and shape. In approximate order of increasing size and life span, they are ripples, megaripples or dunes, sandwaves, bars, barrier islands, deltas, and shelves. Each may be incorporated into the surface form and the buried deposits of larger elements. All generate clinoform bedding surfaces which stack hierarchically in the stratigraphic record. The first five migrate freely; their upstream sides are sites of erosion. The last two grow from an anchored upstream limit and contribute to the empirical data at longer scales than the transition in scaling exponents. Data from transient, migratory landforms determine the short-term exponents ($\Delta \sim T^{<0.45}$). Smaller short-term exponents characterize accumulation near the shoreface, rather than on marine shelf floors in general.

[15] Sediment accumulates much more slowly in deeper water, off-shelf environments, below the reach of surface storms. Accommodation does not limit accumulation here, yet the empirical scaling relations for the continental slopes and rises and the abyssal plain parallel the shelf data at long time spans (Figure 3). Available data are not sufficient to determine the short-term exponents of scaling relations in these deep water environments. Rippled sediment surfaces are generated by bottom currents on the continental rise [*Hollister and Nowell*, 1991] and by turbidity currents on the abyssal plain. Sparse data for the continental rise [*Sadler*, 1993] are consistent with a rollover to smaller scaling exponents in the short term.

3.2.2. Coastal Plains (Figure 4)

[16] At timescales from 10^{-3} to 10^{2} years, the scaling exponent for accumulation on alluvial plains ($\Delta \sim T^{0.17}$) is significantly smaller than that for marine shelf systems. *Pelletier and Turcotte* [1997] obtained a larger value ($\Delta \sim T^{0.24}$), by fitting one line to all our alluvial data from 10^{-6} to 10^{5} years. We chose to segment the data with regard for the full range of averaging times and to isolate the influence of alluvial channels. Channeled environments dominate the supporting data (Table 1). Plotted alone, the smaller data set from floodplain deposits generates a trend that is indistinguishable from that for marine shelf systems. Channels concentrate both erosional and depositional events. They support larger amplitude bed forms than the floodplain, but also enforce more negative feedback, alternating fill and flush intervals. The average fills would appear to be 0.2–2.0 m thick (the plateau in Δ values from 10^{-3} to 10^{2} years); scaling for events shorter than 1 day indicate steady accumulation ($\Delta \sim T^{1.0}$ at timescales from 10^{-6} to 10^{-3} years), however these intervals are too short to be adequately described by a time-averaged model and are not considered further here.

[17] At time spans longer than 10^{3} years, plots of the coastal plain, delta, and marine shelf data become indistinguishable. All three environments share the same accommodation by subsidence. Large deltas support higher mean accumulation rates, but generate the same long-term exponents.

3.3. Scaling Relations for Accommodation

[18] In the simplest models of sediment accommodation, the continental basement surface subsides steadily and the sea surface oscillates regularly. Shelf sediments fill the unsteadily increasing accommodation space between these two surfaces and the scaling relations for accumulation are modulated by scaling relations for the accommodation

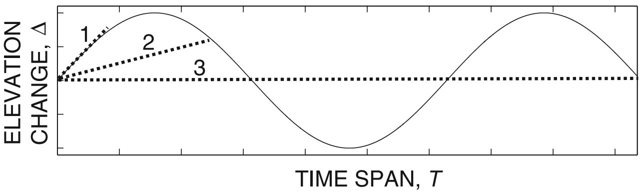

Figure 6. Example of a periodic variation in elevation change, such as sea level, and its effect on measured scaling. At short times elevation change may increase rapidly with time span (e.g., $\Delta \sim T^{1.0}$) as in, for example, measured sea level rise over the Holocene; this is line 1. At longer time spans, averaging captures some of the decrease in sea level: this is line 2. Averaging over several cycles yields little or no dependence of sea level on measured time span (e.g., $\Delta \sim T^{0}$); this is line 3.

processes [*Sadler*, 1994]. Empirical estimates of subsidence and rising sea level reveal more complexity, of course, but still substantiate two basic expectations of the simple model: the scaling relation for subsidence merges with shelf accumulation in the long term; and the scaling relation for sea level rise has a much gentler slope (Figure 5).

[19] The scaling relations for subsidence and sea level rise may be added together to estimate maximum accommodation as a function of time span. On logarithmic plots, the sum closely follows the greater of the two curves and at time spans beyond 10^4 years the sea level component becomes insignificant. At shorter timescales, changes in sea level can dominate both the marine accommodation process (Figure 6) and the position of the coastline as base level for fluvial systems. Falling sea level can eliminate the space created by subsidence up to time spans about 10^4 years [*Sadler*, 1994]. Direct estimates of mean net accommodation (Figure 5) were made by compiling the age and modern depth of ancient peritidal deposits [*Sadler*, 1994].

4. Model Scaling Relations

4.1. Diffusion Model

[20] Researchers have successfully applied a diffusion equation to model the surface of depositional landscapes. Although its form may be justified rigorously in the context of fluvial settings [*Paola et al.*, 1992], the basic one-dimensional form can be derived by assuming that sediment flux varies linearly with slope and invoking conservation of sediment mass to obtain [see *Paola*, 2000],

$$\frac{\partial \eta}{\partial t} = \nu \frac{\partial^2 \eta}{\partial x^2} - \sigma, \qquad (1)$$

where η = elevation [L], x = downstream distance [L], σ = subsidence [LT^{-1}] and ν = the diffusion coefficient [L^2 T^{-1}], which is assumed to be constant with respect to x. To model, for example, a fluvial depositional profile in the downstream (dip) direction, the input sediment flux, diffusivity, and subsidence rate must be specified, along with two boundary conditions. The resulting profile has the form of an error function, and closely approximates elevation profiles observed in many natural systems [see *Paola*, 2000]. One possible justification for applying the diffusion equation to shelf systems is that the combined long-term effect of innumerable weather-driven, wave-like events may resemble diffusive redistribution of sediment [*Niederoda et al.*, 1995].

[21] The fact that the deterministic equation (1) reproduces the morphology of depositional profiles implies that time- and space-varying fluctuations in sediment transport average out over the long term such that they can be neglected, at least for a first-order description of landscape evolution in depositional settings. The time to reach equilibrium for equation (1) may be approximated as $t_{eq} = L^2/\nu$, where L = length of the basin. This timescale is a measure of the time it takes to achieve an equilibrium slope throughout the depositional profile.

[22] The diffusion model with steady subsidence predicts that sedimentary thickness increases almost linearly with time over the entire range and therefore matches the long-term asymptote ($\Delta \sim T^1$), but cannot satisfactorily account for the observed data (Figures 1, 3, and 4) for timescales less than 10^3 years. We hypothesize that the inability of (1) to describe the time-dependent nature of the thickness of sedimentary bodies is because small-scale fluctuations have a cumulative effect on the production of stratigraphy. In other words, errors associated with using (1) average out for the instantaneous surface morphology, but accumulate in the preserved record of surfaces. While unsteadiness in deposition and erosion may be due to deterministic nonlinear dynamics [e.g., *Murray and Paola*, 1994], this autogenic behavior is not well understood in basin-scale systems. We instead explore the descriptive power of a noise term, as used in stratigraphic models.

4.2. Random Walk Models

[23] Several models of the distribution of hiatuses in stratigraphic sections represent sediment accumulation by one-dimensional random walks. The models predict scaling laws for vertical accumulation rate and for the distribution of bedding planes as indicators of gaps in the sedimentary record. *Tipper* [1983] used a discrete random walk. *Strauss and Sadler* [1989] developed the continuous version as a one-dimensional Brownian motion, i.e., the cumulative sum of a white noise, $\varsigma(t)$, having a mean value (drift), $\langle \varsigma \rangle$ [LT^{-1}], and a standard deviation, s [L]. The white noise generates random increments of deposition and erosion (the beds and bedding planes), whose variability is determined by s. They are summed to represent the elevation history of the surface of the accumulating sediment fill, at one place. Successive values of the elevation are correlated, but the sizes of successive increments are not related to one another or the current elevation; the process has no memory. The drift term has obvious physical relevance. It is the mean of the independent increments and, for $\langle \varsigma \rangle > 0$, determines the long-term accumulation rate of the sediment pile. For environments at or above wave base, accommodation rate determines this limit; below wave base, the rate of sediment supply is the limit. Unlike a diffusion term, however, the noise term lacks any physical explanatory power in terms of surface morphology. The models are one-dimensional and not intended to generate topography.

[24] The Brownian motion model successfully predicts curved scaling relations with two asymptotes (Figure 2b, curve i). The crossover time is determined by the ratio $s/\langle \varsigma \rangle$:

noisier systems roll over from noise-dominated to drift-dominated accumulation at longer timescales [*Strauss and Sadler*, 1989]. The long-term limit is steady accumulation ($\Delta \sim T^1$ and $\partial\eta/\partial t = \langle\varsigma\rangle$) and is feasible for the empirical data. The short-term limit is too steep. In Brownian motion, the standard deviation of elevation excursions grows as T^β, where β is called the growth (or Hausdorff) exponent. The preserved thickness of sediment scales linearly with the magnitude of elevation excursions, and therefore random walks generate a growth exponent of $\frac{1}{2}$ at the short-term limit of time span ($\Delta \sim T^{1/2}$). The empirical curves are less steep ($\Delta \sim T^{<0.45}$), especially in the coastal plain and shore systems. These require a model with less short-term persistence, i.e., some nonrandom negative feedback. In other words, for short timescales the increment of accumulation in one time interval does provide some information to aid prediction of the increment in the next interval and the correlation is inverse. Given rapid net accumulation in one short time interval, for example, anticipate slow accumulation or net erosion in the next interval of the same duration.

[25] Three model strategies could reduce short-term persistence: (1) replace the random walk with a fractional Brownian motion [*Sadler*, 1999; *Molchan and Turcotte*, 2002; *Huybers and Wunsch*, 2004] in which antipersistence may be set to the necessary value, (2) add regular periodic fluctuations (Figure 2b, curve iii) which can generate power law exponents as low as $\Delta \sim T^0$ [*Sadler and Strauss*, 1990], or (3) dampen the Brownian motion with a diffusion term [*Pelletier and Turcotte*, 1997]. The physical appropriateness of fractional Brownian motions is not obvious, while the second option is ad hoc, allowing infinite tunability. Again, however, the stratigraphic thinking was one dimensional and vertical. The negative feedback, or damping, was envisioned between successive increments in a vertical succession, not neighboring points on a surface. If every point on the surface executed a one dimensional Brownian motion, the surface cross section would be quite unrealistic: a white noise. For a reasonable physical model of both stratigraphy and topography, the damping must be applied laterally by sediment transport processes dominant in depositional systems. *Pelletier and Turcotte* [1997] realized that a diffusion term could provide the desired damping and add physical meaning in the form of sediment bypassing. Gaps in deposition at one stratigraphic section correspond to depositional increments in downslope neighbors.

4.3. Noisy Diffusion Model

[26] *Pelletier and Turcotte* [1997] found that a noisy diffusion model could explain the scaling of alluvial accumulation at timescales from 10^{-5} to 10^5 years, and also generate topography with realistic surface roughness statistics. They analyzed topographic sections in the lateral (strike) direction, and envisioned the noise to be associated with channel avulsions in fluvial systems. We develop the noisy diffusion model as the simplest relevant representation of transport in the longitudinal (dip) direction, where noise may be associated with storage and release events or the passage of bed and bar forms. We test the model in terms of its fit to the full range of timescales in both coastal plain and shelf systems, and use it to explore the influence of noise on equilibrium time and surface profiles.

[27] The addition to equation (1) of a noise term, $\varsigma(x, t)$ [LT^{-1}], which is uncorrelated in time and space, results in

$$\frac{\partial\eta}{\partial t} = \nu\frac{\partial^2\eta}{\partial x^2} + \varsigma(x,t) - \sigma, \qquad (2)$$

the well-known EW model [*Edwards and Wilkinson*, 1982] used to describe interfaces governed by random deposition with surface relaxation [*Barabási and Stanley*, 1995]. The only difference is the addition of the drift term, to account for subsidence. From the point of view of depositional landscapes, equation (2) models sediment transport as a diffusive process; the noise term may be thought of as representing stochastic fluctuations in transport at a scale smaller than we are explicitly modeling. These small-scale fluctuations will act to impart roughness on an otherwise smooth surface profile. From inspection of equation (2) we anticipate a change in the dominant process governing the deposition of sediment. At short times, subsidence is small compared to the instantaneous sediment transport terms associated with diffusion and noise, and therefore its contribution to sediment deposition is negligible. At long times a steady state surface profile is achieved; since the noise term has a mean of zero, its contribution becomes negligible for the long-time dynamics of the system. It is clear then that the accommodation of sediment by drift (subsidence) will govern the depositional record over long intervals.

[28] We first focus on the short-time behavior of (2), the scaling of which has been determined analytically [see *Barabási and Stanley*, 1995]. Each point on a surface governed by equation (2) undergoes a fractional Brownian walk, i.e., a random walk with some negative feedback which is provided by diffusion. As in normal Brownian motion, the standard deviation of elevation excursions grows as T^β, however for noisy diffusion the growth exponent takes on the value $\beta = \frac{1}{4}$ (i.e., $\Delta \sim T^{1/4}$). Surface roughness of a profile may be described using the interface width, w, which is the root-mean square of surface elevation measured over the domain length, L. The amplitude of roughness scales with the system size, i.e., $w \sim L^\alpha$, where $\alpha = \frac{1}{2}$ is the roughness exponent (and is directly related to the fractal exponent). The crossover time from roughness growth to steady state in a system governed by noisy diffusion scales as $t_x \sim L^\gamma$, where $\gamma = 2$ is the dynamic exponent.

[29] The dynamic exponent of the EW model is the same as that associated with the equilibrium time in deterministic diffusion ($t_{eq} \sim L^2$). This means that it takes approximately the same time to achieve a steady state surface profile, whether or not there is noise. The record of deposits, however, should be quite different for the two scenarios. The short-time scaling from equation (2) is closer to the empirically determined exponents than the random walk models. The model predicts that, beyond the scaling rollover (10^3–10^5 years for field data, Figures 3–5), the time dependence of aggradation disappears. With drift, the steady subsidence dominates at long time spans T and hence $\Delta \sim T^1$ in the asymptotic regime. This behavior models the curved form of our empirical scaling relations and approximates the exponents for the shore and coastal plain systems.

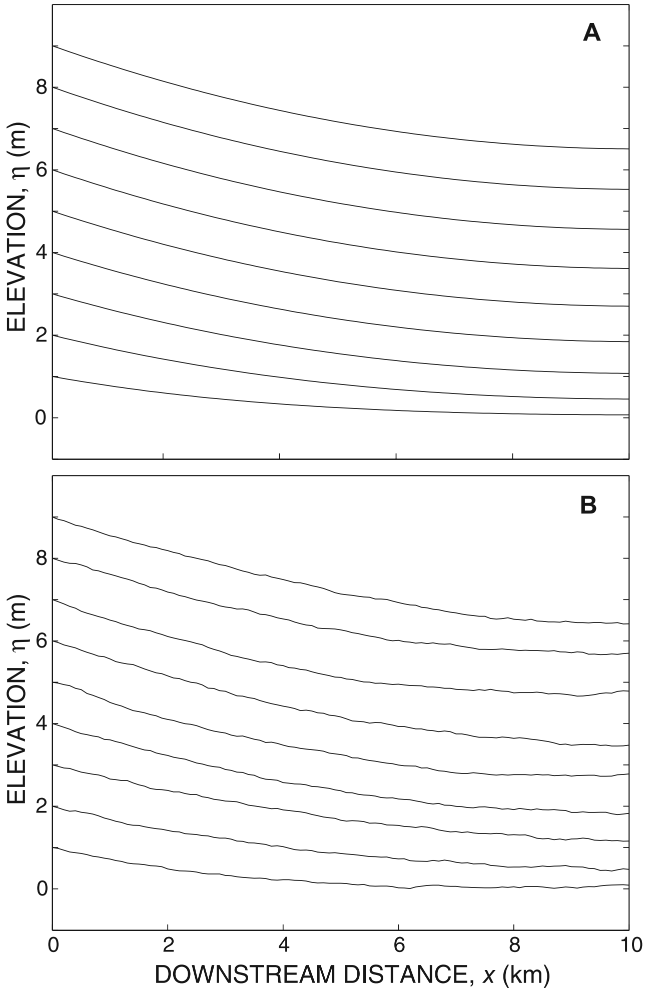

Figure 7. Numerical model results showing profiles of basin filling in a fluvial system. (a) Deterministic diffusion (no noise). (b) Noisy diffusion, in which the relative strength of noise, s/σ = 7. All other parameters and boundary conditions are base case. Profiles are shown at 200-year intervals.

[30] In order to verify the analytical results, we performed numerical simulations of (2) using base-case values for input sediment flux (10^{-4} km^2 yr^{-1}), diffusivity (0.1 km^2 yr^{-1}) and piston subsidence (5×10^{-6} km yr^{-1}) for a fluvial depositional system [see *Marr et al.*, 2000], over a 100-point domain with a grid spacing of 0.1 km. For this model system, the equilibrium diffusion time is 1000 years. Variations from this base case are noted in the text. Boundary conditions maintained a constant slope at the inlet and zero flux at the outlet (see Figure 7). The actual values for the chosen coefficients and noise amplitude affect the magnitudes of model results, but have no influence on the asymptotic scaling exponents. We solved (2) for small time steps, δt = 0.02 years ≈7 days, in order to examine the cumulative effects of short-duration fluctuations on the generation of stratigraphy. We varied the strength of noise for different model runs (as measured by the standard deviation), however noise was always uniformly distributed and uncorrelated with a mean of zero. We estimated sediment thickness over a given time interval from our model runs using the method presented by *Pelletier and Turcotte* [1997]. Results from ten runs show mean $\beta = 0.25$, with a transition to the asymptotic regime ($\beta \to 1$) at long time. The scaling behavior of Δ generated in our simulations (Figure 8) is in agreement with analytical results, and compares well to empirical data of time-dependent sedimentation on the continental shelf (Figures 1 and 3).

[31] At long time, preserved thickness should depend only on subsidence,

$$\Delta = \sigma T. \qquad (3)$$

[32] What determines the time required to achieve the equilibrium condition (3)? We focus first on the effects of varying noise strength, as measured by the standard deviation of the noise, s [L]. The crossover time from noise-dominated to drift-dominated accumulation was measured by projecting power law slopes for the respective short- and long-term asymptotic regimes, and finding the crossing point (Figure 8). Numerical simulations demonstrate that the this crossover time, t_{xA}, increases linearly with the relative strength of noise and may be approximated as

$$t_{xA} \approx \frac{s}{\sigma}, \qquad (4)$$

as found for the random walk models. The actual time to equilibrium (equation (3)) should be several times this value – from numerical experiments, it is 5–10 times larger.

[33] Note that the crossover timescale for aggradation (4) is quite different from the diffusive timescale $t_{eq} = L^2/\nu$. Although an equilibrium slope may be achieved in a depositional system in a time determined by the latter, this does not mean that vertical deposition at each point in the system has achieved a statistical equilibrium. In order for sedimentation at a point to balance subsidence on average, the accumulated sedimentary thickness must be significantly greater than the range of vertical excursions that surface undergoes owing to noisiness. In other words, the equilibrium diffusion time is a necessary but not sufficient condition to satisfy (3). The equilibrium timescale for a basin is determined by the maximum of the two timescales; for a very noisy system the fluctuations may be the limiting factor, while for a less noisy system the diffusion timescale should determine equilibrium. This is demonstrated by a numerical experiment where diffusivity was diminished by 1 order of magnitude (thus increasing t_{eq} from 1000 years to 10,000 years) while noise was held constant (Figure 8). We see that the time span associated with transient scaling ($\Delta \sim T^{1/4}$) does not change, however the long-term behavior shows that accumulation does not achieve a balance with subsidence. Increasing basin size has a similar effect to decreasing diffusivity, since $t_{eq} = L^2/\nu$.

[34] Although low-amplitude temporal fluctuations in sedimentation may not impart a large signature on the surface topography (Figure 7), the addition of noise exerts a strong influence on the depositional record. This may be understood by looking at Figure 9, which shows the fractal nature of the elevation time series and the variability of vertical thickness of depositional bodies. A synthetic stratigraphic column was generated by measuring the truncation surfaces (local minima) that bound depositional units. It is

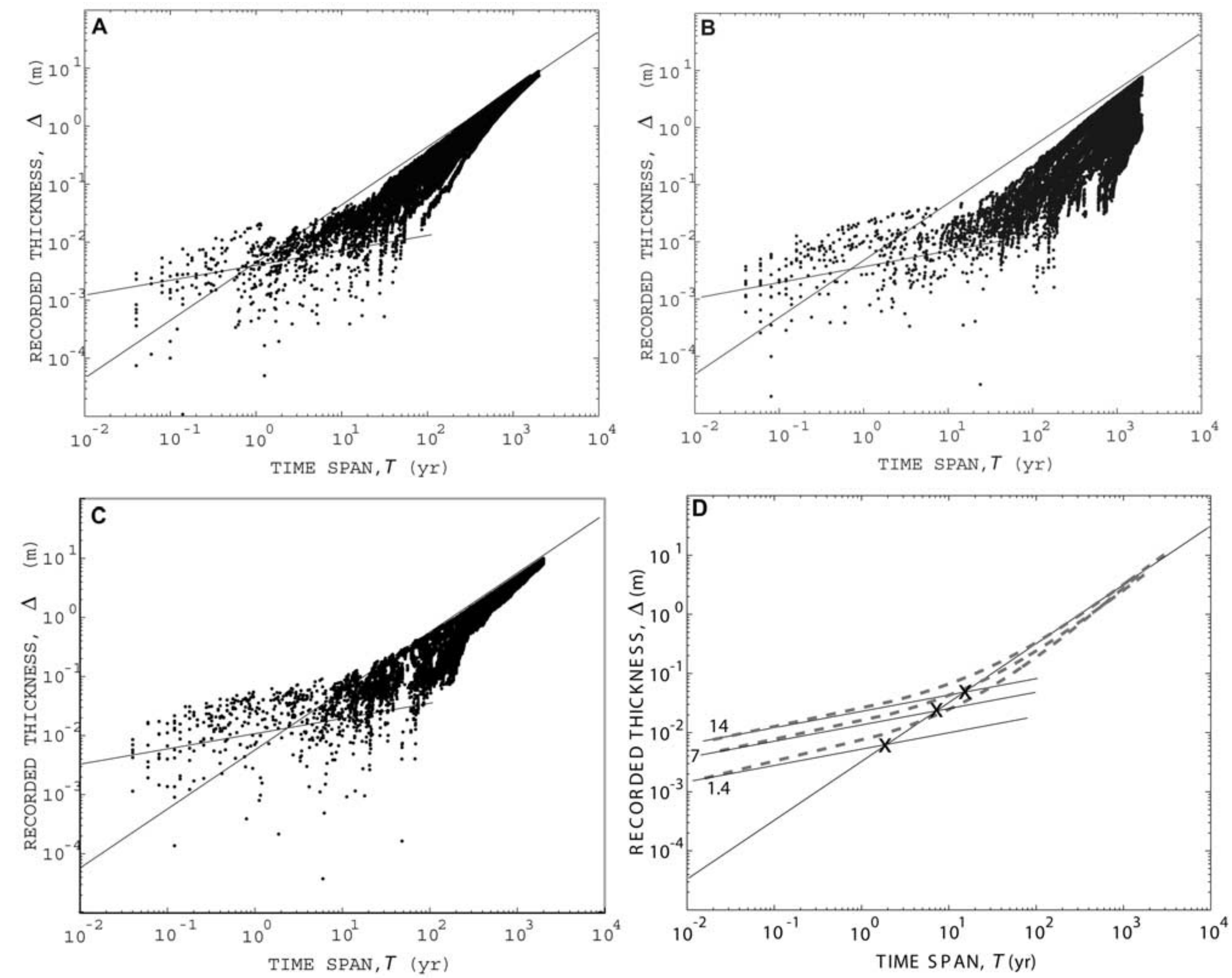

Figure 8. Noisy diffusion scaling of preserved thickness. Numerical model with (a) base case conditions and s/σ = 1.4, (b) diffusivity changed to 0.01 km^2/yr, s/σ = 1.4, and (c) base case conditions with s/σ = 7. Thin solid lines show the slopes of the asymptotic limits, where long-term limit is equation (3). Time and space scales of the model were chosen for convenience of numerical simulations, and were not calibrated to field data. (d) Summary of noisy diffusion results under base case conditions with various relative noise levels, as represented by the ratio s/σ shown next to each curve; the data have been log bin–averaged for clarity [cf. *Sadler and Strauss*, 1990, Figure 10]. Crosses show crossover times.

apparent that random fluctuations superposed on diffusive sediment transport produce hiatuses in deposition at all timescales, and their corresponding truncation surfaces reflect this behavior. The distribution of vertical thickness is exponential, as expected from theory [*Paola and Borgman*, 1991; *Pelletier and Turcotte*, 1997] and observed in some rock outcrops. The very large number of bedding surfaces is due to the fact that every interval of erosion or nondeposition is preserved in the model. In real outcrops not all bedding surfaces will be preserved, and hence very thin beds are probably underrepresented in the geologic record [*Sinclaire and Cowie*, 2003].

5. Discussion

[35] Scaling relations steepen toward the long-term limit ($\Delta \sim T^{1.0}$) for all the models that incorporate a steady drift term (Table 1). In this regard, they all fit the empirical accumulation data equally well. The models differ significantly in the short-term limit, where the field data (Figures 1 and 3–5) have scaling exponents ($\Delta \sim T^{<0.45}$) between those expected for zero net deposition ($\Delta \sim T^0$) and random walks ($\Delta \sim T^{0.5}$). As noted by *Pelletier and Turcotte* [1997], noisy diffusion ($\Delta \sim T^{0.25}$) generates a short-term scaling exponent in this range. The roughness exponent associated with noisy diffusion is also consistent with measured surface roughness scaling of along-dip seismic transects in deltaic environments [*Deshpande et al.*, 1997].

[36] We have shown that noisy diffusion with drift reproduces the salient crossover in the time-dependent scaling of accumulation (Figure 8). Falling at $T > 10^3$ years, the empirical crossover approximates the timescales at which subsidence replaces sea level as the dominant component of accommodation (Figure 5). This appears to justify equating the model drift with steady subsidence.

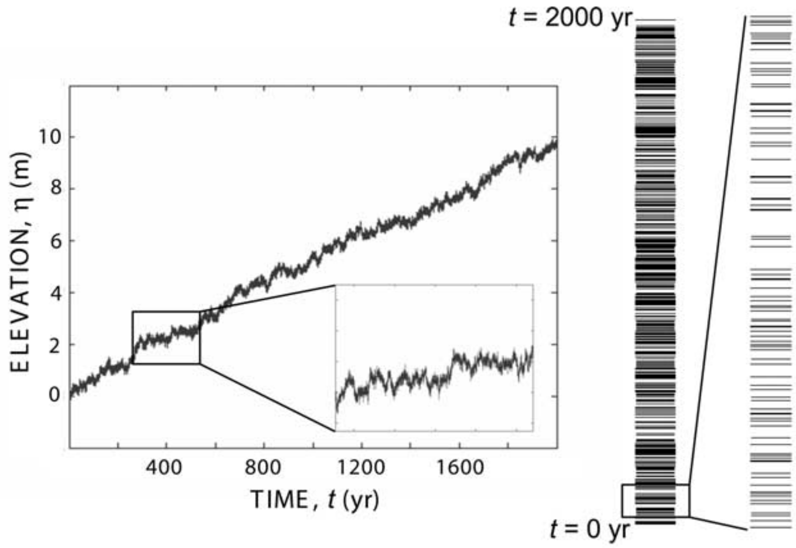

Figure 9. Noisy diffusion numerical model results for (left) the elevation history and (right) the stratigraphic distribution of beds and hiatus surfaces at a single location on the aggrading sediment fill shown in Figure 7b.

[37] For more insight we address the differences between temporal scaling exponents predicted for noisy diffusion and those fit to empirical data. First we address the long-term limit, then the short term. The discrepancies might be explained (1) by terms that are neglected in our simple model, (2) by real systems in which "noise" is not completely uncorrelated in space and time and "drift" is not steady, or (3) by the consequences of combining empirical data from many different locations in order to achieve a wide coverage of time spans.

[38] All the field data converge in the long-term limit, but none convincingly reaches the model asymptote ($\Delta \sim T^1$) for a steady state depositional profile in which accumulation balances steady subsidence. Scaling curves for depositional systems with rich long-term data sets (exclude deltas and floodplains) steepen to long-term exponents of only 0.76–0.86. In model terms, the crossover is long and it is possible that the data never truly saturate, as would be predicted if the time were smaller than that required to achieve steady state [*Barabási and Stanley*, 1995; *Dodds and Rothman*, 2000]. Certainly the boundary conditions in real depositional basins are far more complex than our idealized model. Empirical accumulation and subsidence data do converge in the long term, but subsidence itself does not reach the exponent (1) for a steady process, indicating that subsidence rates diminish with time. Minor unsteadiness may persist in subsidence to very long timescales and subsidence rates surely wane in basins of great age. Short-term clips of basin history may sample any stages of subsidence, but increasingly long-term clips must include more of the slower subsidence of late stages. Additionally, it is possible that extreme longevity in a depositional system is favored by slower subsidence; i.e., that spatial differences in subsidence are loosely translating to temporal differences in our compilation with the effect that long-term estimates are drawn preferentially from sites of low subsidence.

[39] Equating model drift with subsidence fits comfortably with the convergence of real accumulation and subsidence data, but this does not explain why the same long-term exponents characterize environments below wave base where subsidence is not required to accommodate sediment. The drift term for these environments must model sediment supply. Sea level places a lid on the shallow marine sediment trap and excess sediment leaks away to off-shelf environments. Because empirical estimates indicate a negligible role for sea level in long-term accommodation (Figure 5), the scaling connection between shelf and off-shelf systems must be sought in the sediment flux from the continent. Like subsidence, this too may inevitably wane for continental margins and their hinterland sources that persist for hundreds of millions of years. We suggest that $\Delta \sim T^{0.85}$ reflects a real long-term limit for sustainability of subsidence under the sediment trap and rejuvenation of sediment sources by uplift.

[40] Notice a critical difference between trains of ripples that represent the small-scale end of the size and life-span spectrum and a single continental shelf at the large-scale end. Sediment "leaked" from one ripple feeds the next ripple in the train. Sediment leaked from the shelf is lost to the shelf system. This is another factor that differentiates the transient and persistent landscape elements.

[41] At the short-term limit, scaling in the field data varies significantly with environment. Empirical scaling relations for floodplains and shelves are steeper than for channeled alluvium at scales from 10^{-3} to 10^2 years and the crossover to accommodation-dominated scaling is delayed until longer averaging times for channeled environments. By concentrating sediment transport into a narrow zone, channels increase both the rapidity and noisiness of accumulation. Pure floodplain environments aggrade with less noise amplitude. This difference alone could explain why they cross over from noise (transport)-dominated scaling to drift (accommodation)-dominated scaling at shorter time spans than the channeled parts of the system (Figure 8); see equation (4). The empirical data indicate, however, that differences in noise amplitude may not be sufficient to explain all the

differences in scaling between channeled and channel-free settings. From 10^{-3} to 10^2 years the channeled scaling relation is less steep than predicted by the noisy diffusion model and at averaging times shorter than 10^{-3} years it is much steeper. Scaling that deviates significantly from model predictions may indicate a regime in which slope-dependent sediment transport is not the dominant process. Certainly, modeling sediment transport as diffusive is inappropriate for very short timescales over which the assumptions of steady, uniform flow do not hold (e.g., ripples and dunes). For fluvial systems affected by channel avulsions, one must average over the timescale of several avulsions in order to approximate the system as diffusive [*Sun et al.*, 2002; *Hickson et al.*, 2005].

[42] The short-term data from nearshore environments have a scaling exponent that is intermediate between accumulation on floodplains and open shelves. It most closely matches noisy diffusion (Table 1). Short-term accumulation at the shore is uniquely sensitive to waves, tides, and storms. Although each of these processes can impose a periodic negative feedback, in aggregate they may combine to resemble diffusion [*Niederoda et al.*, 1995].

[43] In this paper we propose two timescales that describe sedimentation. One is the classical diffusion time which is based on mean transport characteristics and system size, and the other is the aggradation timescale (4) which is determined by the variance of transport in time and space. Fluctuations in sediment transport exert a first-order control on the deposits constructed from evolving surfaces. These fluctuations, however, need not strongly influence the morphology of surfaces themselves. Rather, noise generates small departures from a deterministic diffusive profile that, to first order, may be ignored when modeling the evolution of a depositional surface over geologic time (Figure 7). Our numerical results show that a fluvial system subject to low-amplitude noise achieves a steady state profile in the same amount of time as a system with no noise, i.e., $t_{eq} \approx L^2/\nu$ always.

[44] Although spatial and temporal variations may average out in the long-term evolution of depositional surfaces, they accumulate in the stacked record of surfaces. To find equilibrium in the stratigraphic record, we must also look at vertical scales that are significantly larger than the range of variability induced by transport fluctuations; the mean is not enough. This is evident in Figure 9, which shows the depositional record left by the profiles shown in Figure 7b. One might be tempted to delineate zones of regular and irregular bedding surfaces in the stratigraphic column, and to attribute these zones to different depositional environments or forcing conditions, when in fact the stratigraphic horizons at all scales were generated by the internal fluctuations of the depositional system. In an unsteady system, equilibrium is achieved when both the diffusive timescale *and* the aggradation timescale (e.g., equation (4)) are satisfied.

[45] Simplified models of sediment transport (e.g., diffusion) neglect the transient landscape evolution that dominates stratigraphy at time intervals up to tens of thousands of years. The noisy diffusion model shows that the equilibrium time associated with sediment accumulation (i.e., when deposition rate equals subsidence) depends linearly on the magnitude of noise. In order to compute the averaging time required to achieve equilibrium in a stratigraphic section, it is necessary to understand the statistics of spatial and temporal fluctuations in sediment transport. The magnitude of elevation fluctuations may be determined by the internal nonlinear dynamics of a sediment transport system, or by some environmental fluctuation like sea level. In the modeling of a fluvial profile this noise could be related to fluctuations of the bed associated with the passage of dunes and bars or to scour events; therefore, its amplitude would scale with channel depth. Applying this reasoning to a 2-m-deep river undergoing subsidence at 1 mm/yr, we expect a crossover time from noise to accommodation regimes at 2000 years: this is of the same order as crossover times observed in empirical data for alluvial plains and deltas (Figure 4). The time to achieve equilibrium should be several times larger than this crossover time (perhaps 10,000–20,000 years). Conversely, one may estimate the mean subsidence rate from the long-term accumulation curve in Figure 4 as 1 mm/yr. Estimating the crossover time to be about 1000 years, we approximate the scale of noise fluctuations to be of order meters. Applying the same crude procedure to the shelf data (Figure 3) results in a noise scale that is roughly an order of magnitude smaller, commensurate with the smaller crossover time. These estimates are only meant to qualitatively apply the principles of our model to the data that motivated it, and should not be taken any more seriously than that. In rivers, avulsions or other events may introduce a characteristic noise magnitude that is different from, and potentially larger than, the scales outlined here; the appropriateness of the suggested noise scales remains to be shown. Moreover, if the "noise" is correlated in space or in time, rather than a white noise, the exponent of the short-term scaling will change [*Sadler*, 1999; *Huybers and Wunsch*, 2004], and the application of our model is inappropriate. However, even in these cases the general concepts discussed here help to guide interpretation.

6. Conclusions and Implications

[46] By comparing our empirical data with an array of simple models, we can separate noise-dominated from drift-dominated behavior in the siliciclastic depositional landscapes of continental margins. In terms of process, these behaviors are transport-dominated (autogenic) and accommodation-dominated, respectively. The first is characterized by transient landscape elements that are freely migrating landforms, typically arrayed in trains. The second is characterized by large, persistent, anchored landscape elements that are singular, not arrayed in trains. The transition to persistent landscape forms begins at timescales of 10^2 to 10^3 years and is complete by 10^4 to 10^5 years. Noisy diffusion appears to be the most powerful simple model for these observations. It approximates all the gross features of the empirical data. In detail, differences between model and empirical scaling relations may reveal systems in which a diffusive description of sediment transport is not appropriate, or transport fluctuations that are correlated. Real continental margins probably depart slightly but consistently from the model expectations at long timescales because both subsidence and sediment supply tend to wane if the system persists for hundreds of millions of years. In any case, in

order for sedimentation to balance subsidence two conditions must be satisfied: (1) the system must achieve an equilibrium morphology (e.g., the diffusion timescale for systems where diffusion is appropriate), and (2) the thickness of deposited sediment must be significantly larger than the range of transient scour and deposition events that occur on the surface.

[47] It is a mathematical inevitability that unsteady sedimentation will produce a time dependence in the thickness of sedimentary bodies, and hence sedimentation rates [*Sadler*, 1981, 1999]. A deterministic model (1) capable of describing the instantaneous morphology of a surface profile fails to describe the cumulative record of surface morphologies because this record depends on the history of elevation fluctuations. The addition of uncorrelated noise produces hiatuses in sedimentation at many timescales, which some researchers would interpret as external forcing to the system, such as climate change.

[48] We close by posing a currently crucial question. What do these findings have to say about the chances of being able to extract that part of the stratigraphic record of successive landscapes that records climate change? The power of the noisy diffusion model is discouraging in this regard; there appears to be no need for a first-order climate-dominated segment in the scaling relations. Climate signals must be sought in departures from the predictions of noisy diffusion. At long timescales the climate influences weathering and sediment flux; it must be separated from tectonics. At short timescales there is a suggestion of cycle-dominated behavior in channeled systems. However, any cyclic environmental signals must be deconvolved from the "noise" produced by the autogenic variability of the depositional system [*Lyons*, 2004]. The timescales of glacio-eustasy, at which climate change exerts a profound influence through sea level, appear to lie in the transition from transient transport-dominated regimes to persistent accommodation-dominated regimes. These are the scales at which depositional landscapes and their stratigraphic record are sensitive to the widest array of influences and the deconvolution challenge is correspondingly hard. The challenge for geologists is to better characterize the nature of intrinsic variability in sedimentary systems. Detailed studies from the field [*Lyons*, 2004] and experiments in a laboratory-scale subsiding basin [e.g., *Hickson et al.*, 2005; *Kim et al.*, 2006] are now beginning to shed light on the dynamic behavior of depositional channels and the construction of their associated fans. Experiments can be used to study the statistical distribution of sedimentary deposits in detail, and to relate these deposits exactly to the physical processes that generated them. A more refined model in which autogenic variability is understood through nonlinear dynamics, or treated stochastically using measured distributions of fluctuations, will enhance the resolution within which we can confidently interpret the record of Earth-surface dynamics preserved in sedimentary rocks.

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