JGR Earth Surface

FEATURE ARTICLE
10.1029/2019JF005079

GRAND CHALLENGES
CENTENNIAL COLLECTION

Buffered, Incomplete, and Shredded: The Challenges of Reading an Imperfect Stratigraphic Record

Kyle M. Straub[1], Robert A. Duller[2], Brady Z. Foreman[3], and Elizabeth A. Hajek[4]

[1]Department of Earth and Environmental Sciences, Tulane University, New Orleans, LA, USA, [2]Department of Earth, Ocean, and Ecological Sciences, University of Liverpool, Liverpool, UK, [3]Department of Geology, Western Washington University, Bellingham, WA, USA, [4]Department of Geosciences, Pennsylvania State University, PA, USA

Key Points:

- We review three impediments to stratigraphic storage of environmental signals that arise from Earth surface processes
- The magnitude of these impediments is set by emergent and self-organized scales in landscapes, which can be predicted from stratigraphy
- Future work is needed to understand scales of stochasticity in a range of environments to improve our paleoenvironmental reconstructions

Correspondence to:
K. M. Straub,
kmstraub@tulane.edu

Citation:
Straub, K. M., Duller, R. A., Foreman, B. Z., & Hajek, E. A. (2020). Buffered, incomplete, and shredded: The challenges of reading an imperfect stratigraphic record. *Journal of Geophysical Research: Earth Surface*, *125*, e2019JF005079. https://doi.org/10.1029/2019JF005079

Received 8 AUG 2019
Accepted 30 JAN 2020
Accepted article online 18 FEB 2020

Abstract Climate, tectonics, and life influence the flux and caliber of sediment transported across Earth's surface. These environmental conditions can leave behind imprints in the Earth's sedimentary archive, but signals of climate, tectonic, and biologic change are not always present in the stratigraphic record. Deterministic and stochastic surface dynamics collectively act as a stratigraphic filter, impeding the burial and preservation of environmental signals in sedimentary deposits. Such impediments form a central challenge to accurately reconstructing environmental conditions through Earth's history. Emergent and self-organized length and timescales in landscapes, which are themselves influenced by regional environmental conditions, define spatial and temporal sedimentation patterns in basins and fundamentally control the likelihood of environmental signal preservation in sedimentary deposits. Properly characterizing these scales provides a key avenue for incorporating the known "imperfections" of the stratigraphic record into paleoenvironmental reconstructions. These insights are necessary for answering both basic and applied science questions, including our ability to reconstruct the Earth system response to prior episodes of climate, tectonic, or land cover change.

Plain Language Summary Reconstructing the history of Earth prior to the age of scientific instrumentation relies heavily on interpretations of layers of sedimentary rocks, collectively called the stratigraphic record. The composition, architecture, chemistry, and fossils contained in these rocks provide signals of past climate, tectonics, and biology on Earth. However, the storage of these environmental signals in stratigraphy is not straightforward. Environmental signals can be transformed by sediment transport through channels and the landscapes that surround them. This transformation continues as sediment is deposited and strata are formed. In some cases transformation of signals severely hampers paleoenvironmental reconstruction. Recent theoretical developments allow us to model environmental signal propagation through landscapes and to estimate signal distortion or destruction during the burial process. This aids estimation of uncertainties in our paleoenvironmental reconstructions. Further improvements in our ability to quantify these uncertainties will require more detailed descriptions of the statistics and underlying physics of sediment transport and deposition. Improvements in theory, which could aid our ability to predict the statistics of sediment transport and deposition, will need to be tested against laboratory and field observations.

Two quotes highlight both the opportunities and challenges associated with reading the stratigraphic record:

"The sediments are a sort of epic poem of the earth. When we are wise enough, perhaps we can read in them all of past history." (Carson, 1951)

"The stratigraphical record is a lot of holes tied together with sediment." (Ager, 1973)

Our best record of past conditions on Earth, for most of its history, comes from strata in sedimentary basins. Clues to this history are housed in the composition, spatial organization, chemistry, and fossils of strata. The record allows us to interpret the Earth system response to past episodes of climate change. These interpretations are not only important for characterizing Earth's past but also aid our ability to predict responses to ongoing climate change. To read Earth's epic poem, though, we must solve one of the most complicated inverse problems known to science. Ager's quote, referenced above, was made to highlight that the stratigraphic record, at any one site, is incredibly incomplete. Harking back to twentieth century technology, some have described the stratigraphic record as a tape recorder of Earth's history. However, the tape recorder is only on when sediment is being deposited. Gaps in the record, resulting from periods of inactivity and/or erosion, have at least been known since James Hutton published his *Theory of the Earth* in 1788, in which he discussed the significance of an unconformity in Scotland (Hutton, 1788). Even when sediment is deposited, interpreting strata for paleoenvironmental conditions is difficult due to complex Earth surface responses to forcings (Schumm, 1973). Depending on the type of forcing, signals of the deterministic response found in landscapes and stratigraphy can sometimes be amplified or buffered. Further complicating the problem is that extreme sensitivity to preexisting conditions results in a component of Earth surface response to environmental forcings that is best described stochastically. The time and length scales of stochastic Earth surface processes, relative to environmental forcings, impact our ability to separate signal from noise in landscapes and strata.

Ongoing work in the fields of geomorphology and stratigraphy is focused on quantifying the magnitude and source of impediments to environmental signal storage in landscapes and strata. Here we focus on defining three critical impediments to environmental signal storage, with particular emphasis on the physics that sets the upper limits on the time and length scales of these impediments. We note that impediment scales are set by emergent and self-organized length and time scales in a landscape, which are set by regional environmental conditions. As such, not all landscapes and strata have the same signal storage potential.

We start by exploring the tendency for landscapes to buffer their response to environmental forcings. This response can be approximated with a diffusion equation, which results from a statement of conservation of mass coupled to a slope-dependent sediment flux term (Paola et al., 1992). A scaling argument and rearrangement of this equation yields a prediction for the time required to regrade topography to a new steady state following a change in forcing. This time is referred to as a basin's equilibrium time, T_{eq}, and is often on the order of 10^4–10^6 yr. The diffusional response over such long timescales means that many landscapes are unable to fully respond to environmental change and that their strata might have muted signals relative to the true magnitude of the forcing change (Allen, 2008).

The transport of water and sediment in channels leaves wide swaths of a landscape inactive for long time spans. This inactivity, coupled with periods of erosion, produces hiatuses in stratigraphic records. This (in)completeness warps the representation of environmental records in the spatial domain that, given our current suite of geochronomenters, can be challenging to correct for when constructing age models. A critical timescale, constructed with knowledge of landscape roughness and long-term deposition rates, can be used to estimate the completeness of stratigraphic records and thus record fidelity. This "compensation" timescale is commonly of the same order as the period or duration of many environmental forcings, which challenges our ability to reconstruct Earth history from 1-D stratigraphic sections. New techniques are starting to be developed which may allow us to recover signals as we widen our field of view to include 2-D and 3-D observations.

Reconstructing some environmental forcings is challenging even with temporally complete stratigraphic records due to the nature of sediment transport. Sediment transport and landscape evolution are analogous to fluid turbulence, in that the chaotic transport of sediment can dissociate environmental signals across space and time to the point where they can no longer be reconstructed (Jerolmack & Paola, 2010). Similar to fluid turbulence, we are starting to develop mechanisms to quantify the scales of morphodynamic turbulence and compare them to environmental forcings. Much of the recent theory constructed to predict signal storage thresholds in light of morphodynamic turbulence has also used the compensation timescale.

An exciting attribute of current research into storage thresholds for signals in landscapes and stratigraphy is that they can be plausibly estimated from measurable parameters in field-scale systems, for example, estimates of paleo-channel depths, deposition rates, and sediment fluxes.

Theory and field methods can be applied to studies of societally relevant topics that involve the sedimentary record, for example, reconstructions of the Paleocene-Eocene Thermal Maximum (PETM) climate

change. This example highlights how constraining practical uncertainties ultimately entail comparing the timescale of environmental change to the scale of the transport system relevant for a given study.

Grand challenges that the community must address to further improve our ability to read the stratigraphic record share several common themes. These include the need for (1) defining the morphodynamic roots of landscape stochasticity across depositional environments, (2) the surface process and stratigraphy communities to engage more fully with one another, (3) stratigraphers to embrace hypothesis testing and the quantification of uncertainty in our interpretations, and (4) the next generation of stratigraphers to be trained in both quantitative theory and the application of theory to specific problems using available field techniques.

1. Goals and Scope of the Paper

Much of what we presently know about deep-time Earth history, including the history of climate, tectonics, eustasy, and life, comes from analyzing the stratigraphic record. Accurately reading this history allows us to address many pressing questions about how landscapes and ecosystems respond to climate change (Clift et al., 2008; Davies & Gibling, 2010; Fan et al., 2018; Foreman et al., 2012; Knight & Harrison, 2014; Turner, 2018) and expands our ability to explore environmental conditions and Earth states far beyond what is observable from recent instrumental or geologic records. Information contained in this record also defines the relatively unique conditions associated with the rise of our species (Gani & Gani, 2011; Villmoare et al., 2015) and the delicate balance we must maintain for our future success. Outside the confines of our home planet, interpreting sedimentary deposits from other planetary bodies broadens our understanding of the evolution of our solar system and the origins of life (Goudge et al., 2018; Grotzinger et al., 2005; Grotzinger et al., 2014).

While the stratigraphic record is our best window into Earth's deep past, it is not without flaws. Derek Ager famously wrote that the "stratigraphical record is a lot of holes tied together with sediment" [1973]. These holes are stratigraphic hiatuses or gaps in the record, resulting from periods of nondeposition or erosion. Many of these holes arise from complex sediment transport dynamics over Earth's surface (Paola et al., 2018), which have fundamental implications for the stratigraphic storage of environmental signals. Here we synthesize the origins and consequences of three primary impediments that arise from sediment transport dynamics and limit how environmental signals are stored in and can be recovered from the stratigraphic record:

1. The buffering of signals as they propagate from sediment sources to sinks, as modeled by diffusion.
2. The (in)completeness of the stratigraphic record.
3. The shredding of environmental signals by stochastic sediment transport processes.

We review each impediment, highlighting critical thresholds that determine whether and how environmental signals are encoded in the stratigraphic record. We then show how this theory can be connected to field-scale systems and used to constrain uncertainties associated with paleoenvironmental reconstructions. This largely centers on techniques to estimate critical paleolandscape length and timescales relative to an environmental signal of interest. We then present how theory and field methods can be applied in studies of societally relevant topics that involve the sedimentary record, focusing on reconstructions of the PETM climate change as an example. This example highlights how constraining practical uncertainties ultimately entails comparing the timescale of environmental change to the scale of the transport system relevant for a given study.

Finally, we identify Grand Challenges that the community must address to further improve our ability to read the stratigraphic record and thus unlock quantitative information about the past, which will aid our ability to forecast the future. These challenges share several common themes which include the need for (1) the surface process and stratigraphy communities to engage more fully with one another, (2) stratigraphers to embrace hypothesis testing and the quantification of uncertainty in our interpretations, and (3) the next generation of stratigraphers to be trained in both quantitative theory and the application of theory to specific problems using available field techniques.

Table 1
Key Terms With Associated Definitions

Autogenic: Patterns, variability, or dynamics that arise solely as a consequence of interacting components within a particular system.
Buffered landscapes: Those that respond to perturbations over timescales that are greater than a periodicity of an environmental forcing. Resulting deterministic landscape responses (i.e., environmental signals) are muted relative to the magnitude of the forcing.
Deterministic: A process that involves no randomness in the development of future states of the system.
Emergent: A property of a macroscopic system (e.g., a channel on a landscape) that is not present at a microscopic scale (e.g., the scale of individual sediment grains) despite the fact that the macroscopic scale is a large ensemble of the microscopic systems.
Environmental forcing: The large-scale external factors that ultimately control the amount of sediment available and the amount of space to store sediment on Earth's surface (e.g., parameters related to climate, tectonic setting, or biological conditions).
Environmental signal: Attributes of a landscape's structure, sediment transport capacity and/or stratigraphic characteristics that can be linked directly to the environmental forcings.
Self-organized: An ordered or patterned outcome of the internal dynamics of a system. The order arises from local interactions between parts of a system that can initiate as a disordered system.
Signal shredding: The process of stochastically transporting, depositing, and eroding (reworking) sediment such that the signal of a changing environment forcing is chaotically disassociated across space and time to the point where it can no longer be reconstructed.
Stochastic: A process that can be defined by a random probability distribution which is best analyzed statistically.
Stratigraphic (In)completeness: The concept that a stratigraphic record, for example, a 1-D section of strata, contains sediment that imperfectly samples the time between the start and end of the construction of the section. Thus, when a spatial series of strata is converted to the time domain, discretized with a defined time interval, the record is either incomplete (i.e., not all time steps are represented by some preserved sediment) or complete (i.e., all time steps are represented by preserved sediment).

2. Essential Definitions and Background Information

2.1. What Is an Environmental Signal?

Extracting environmental signals (here referred to generally as “signals” of climate, tectonic, eustatic, or land cover change) from stratigraphic records largely relies on detailed observations of strata and inductive reasoning for the development of conceptual or semiquantitative models (Allen, 1963; Beerbower, 1961; Bull, 1991; Catuneanu et al., 1998; Gilbert, 1890; Vail et al., 1977; Vendettuoli et al., 2019). While careful descriptions continue to form the backbone of our subdiscipline, a grand challenge for the geoscience community is to develop quantitative theory that provides a basis for generating field testable hypotheses to understand how environmental signals are stored in strata. Given the vast diversity of depositional environments found on Earth and those that existed in the past (Anderton, 1985; Boyd et al., 1992; Miall, 1977; Van Wagoner et al., 1988), this critical task is nontrivial. Channelized clastic systems are characterized by measurable and predictable morphodynamic relationships (Colombera et al., 2017; Paola et al., 2006; Rodriguez-Iturbe et al., 1992) and offer a model for how we can understand and predict the link between Earth surface processes and stratigraphic products.

A sediment routing system (SRS) is sensitive to a host of environmental conditions that operate over a wide range of space and timescales (Allen, 2017). The channelized portions of SRSs take advantage of gravity to move material from regions of sediment production (sources) to sites of permanent stratigraphic storage (basins or “sinks”) and are consequently particularly sensitive to topographic gradients, discharge availability, and conditions that influence the mobility of channel networks. Here we differentiate environmental *forcings* from environmental *signals*. Environmental forcings are the large-scale external factors (e.g., climate, tectonic, eustatic, or biological conditions) that ultimately control the amount of sediment produced and transported across Earth's surface and the amount of space available to store sedimentary deposits. We focus on environmental signals that are attributes of a landscape's structure or sediment transport capacity that can be linked directly to the environmental forcings (Table 1). Such attributes, including sediment production, erosion, transport, and deposition rates and the caliber of sediment in flux, can impart stratigraphic signatures identifiable through, for example, the spatial arrangement of lithofacies or physical and chemical patterns in deposited sediments.

In this review we focus on quantifying the potential for stratigraphic signal storage of environmental forcings that operate over mesotimescales. We follow a definition of mesotimescales promoted by Sheets et al. (2002): A timescale that “lies between a ‘short’ timescale on which individual channels or channel segments behave coherently and deterministically—the timescale of most engineering models—and a ‘long’ timescale on which autocyclic variability sums to produce the average behavior represented in large-scale stratigraphic models." In field-scale systems these mesoscale dynamics and products usually span 10^1–10^5 yr (Foreman & Straub, 2017; Sheets et al., 2002; Straub & Wang, 2013). Recent research suggests that mesotimescale environmental forcings are associated with the greatest difficulty in separation of environmental signals from the products of stochastic sediment transport dynamics ('noise') (Armitage et al., 2011; Covault et al., 2013; Fernandes et al., 2016; Foreman & Straub, 2017; Forzoni et al., 2014; Hajek et al., 2010; Jerolmack & Paola, 2010; Li et al., 2016; Toby et al., 2019; Trower et al., 2018). While high-precision records of recent events like floods and hurricanes can be captured in some settings (e.g., Aalto et al., 2003; Aalto & Nittrouer, 2012; Donnelly & Woodruff, 2007), dynamic hydraulic and morphodynamic processes inherent in sediment transport systems over mesotimescales can rework these snapshots making it unlikely that they will find their way into the long-term sedimentary record.

A combination of deterministic and stochastic processes describes the transport of sediment and the evolution of the Earth's surface along a SRS. Understanding the processes and scaling associated with these conditions provides a practical approach to understanding how and when landscape dynamics impede the stratigraphic storage of environmental signals.

2.2. Deterministic Processes in SRSs

Deterministic models (Table 1) that explore the surface dynamics and stratigraphy of channelized SRSs over mesotimescales typically predict the diffusion and transformation of signals over time and propagation distance (Figure 1a) (Allen, 2008; Paola et al., 1992). This approach receives significant attention for several reasons. First, it enables a first-order prediction of landscape adjustment to environmental signals and therefore capacity to buffer or transmit environmental signals (Allen, 2008; Romans et al., 2016; Wickert & Schildgen, 2019). Second, many of the physical parameters incorporated into their derivations can be estimated for field systems and so deterministic forward stratigraphic models offer testable results (Allen, 2017; Armitage et al., 2013; Métivier & Gaudemer, 1999; Paola, 2000). Third, the very nature of deterministic formulations means that they give exact and repeatable predictions that can be tested in the field without worrying about the variability introduced by stochastic processes (Duller et al., 2010; Watkins et al., 2018).

The deterministic response of sediment transport systems to environmental forcings, however, can be quite complex. This has been recognized at least since the 1970s, when Shumnn and Parker used physical experiments conducted in the Rainfall Erosion Facility at Colorado State University to investigate the response of drainage systems and stratigraphy to simple changes in forcing conditions (Schumm, 1973; Schumm & Parker, 1973). For example, they proposed deterministic models to explain an observation that a single step lowering of base level could drive multiple episodes of channel incision, with periods of deposition between. The complex deterministic response results from the spatial propagation of erosional waves over a landscape, which increases sediment flux downstream of the incision, and because of complex adjustments in channel width generally not captured in 2-D models (Finnegan et al., 2005; Pelletier & DeLong, 2004; Tofelde et al., 2019).

2.3. Stochastic Processes in SRSs

In addition to complex deterministic responses, many numerical formulations do not generate the rich structure of strata that is required to explore limits of environmental signal recovery because they do not account for the stochastic (Table 1) variability of processes that contribute to the construction of strata. Below, we explore two ways in which stochastic processes influence environmental signal storage in stratigraphy. First, elevation fluctuations, best characterized by probabilistic distributions, result in stratigraphic hiatuses due to periods of stasis (nondeposition) between fluctuations and due to erosion (e.g., Ganti et al., 2011; Schumer & Jerolmack, 2009; Strauss & Sadler, 1989) (Figure 1b). Second, stochastic episodes of sediment storage and release in landscapes rework signals of input forcings, a process recently termed signal shredding, which can sometimes obliterate a signal prior to stratigraphic storage (Jerolmack & Paola, 2010; Li et al., 2016; Toby et al., 2019) (Figure 1c).

Stochastic dynamics in SRSs arise for two reasons: environmental stochasticity and autogenics. Environmental stochasticity refers to events like earthquakes, storms, and floods which are difficult to predict given their chaotic response to initial conditions but can impart significant variability to sediment flux and accommodation for sediment storage (Ashton et al., 2001; Goldfinger et al., 2012; Hajek & Straub, 2017; Peters & Loss, 2012). For example, landscape thresholds mean that some depositional environments,

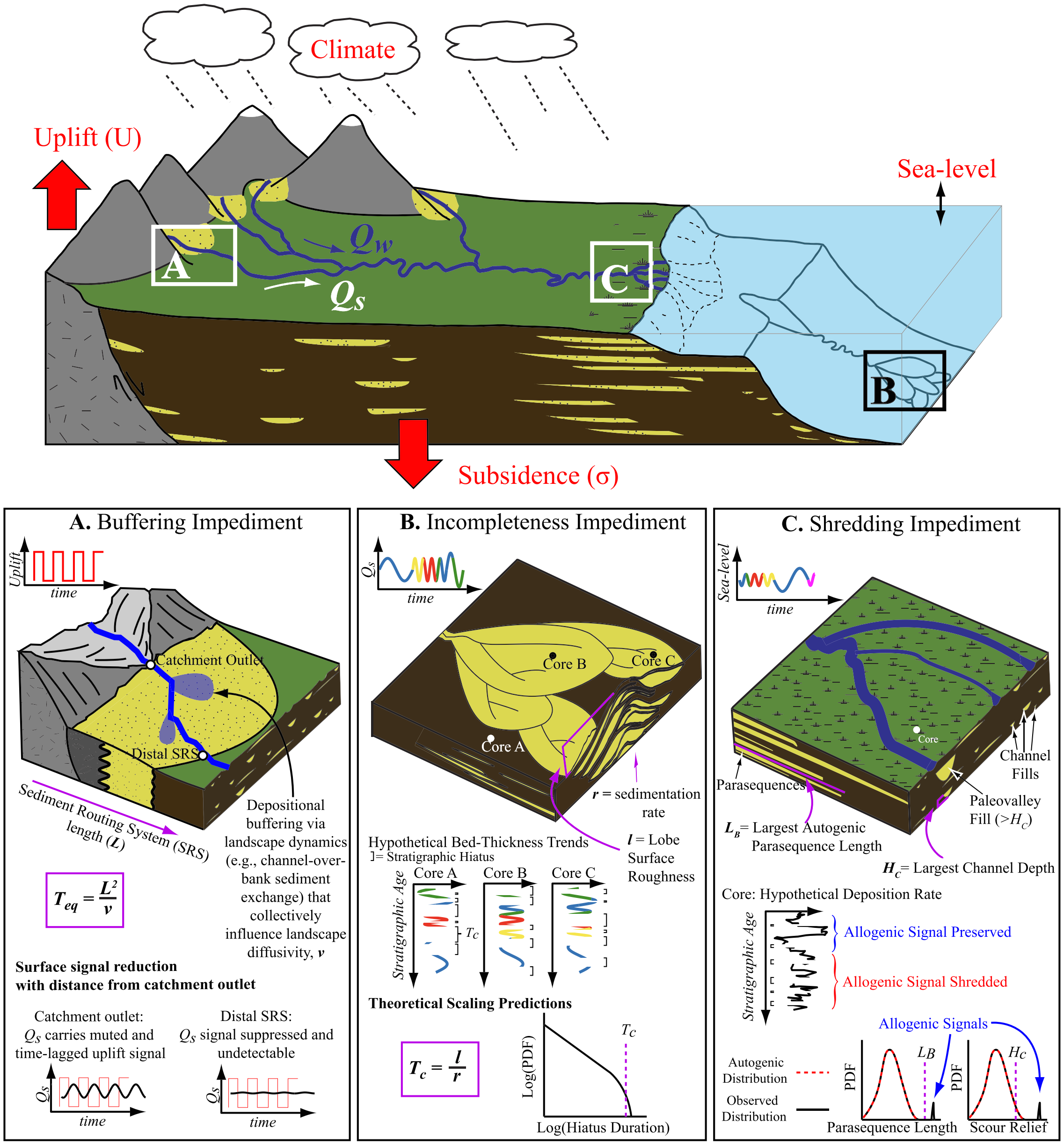

Figure 1. Conceptual diagram of a sediment routing system (SRS) and the major impediments to stratigraphic storage of environmental signals. SRSs transport sediment from erosional sources to depositional sinks and are sensitive to environmental forcings (climate, tectonics, and sea level). Propagation of signals of these forcings across the Earth surface and into the stratigraphic record is influenced by three primary impediments shown in schematic form in subpanels. Each impediment can affect any section of a SRS, but for simplicity we highlight just one impediment per transport system segment. (a) The propagation of environmental signals is buffered via the interaction of channels with landscapes. For example, step changes in tectonic uplift experienced in catchments produce muted sediment flux signals with time lags at catchment outlets. These sediment flux signals become more muted with additional transport through fluvial SRSs. The amount of buffering can be predicted by comparing the duration or period of an environmental forcing with a system's equilibrium timescale (T_{eq}). (b) The stochastic relocation of SRSs, for example, through depositional lobe avulsions in deepwater systems, can convert a continuous input sediment supply signal into a discrete and discontinuous record with time preserved in sedimentary beds separated by gaps or hiatuses of nondeposition. The duration of these hiatuses is limited by the system's compensation timescale (T_c). (c) Cut-and-fill processes in net depositional settings can shred environmental signals prior to stratigraphic storage by smearing input signals though space and time. When environmental forcings, for example, changing sea level, have magnitudes and periods less than autogenic timescales, they are prone to shredding. As such, stratigraphic scales, for example, deposition rates, parasequences or scours associated with the forcing, might be indistinguishable from autogenic stratigraphic scales.

or parts of depositional environments, are only activated during events of a given magnitude. For instance, during floods the near bed shear stresses act to increase rates of in-channel sediment transport and overbanking flow activates floodplains (Aalto & Nittrouer, 2012; Adams et al., 2004). The frequency at which a

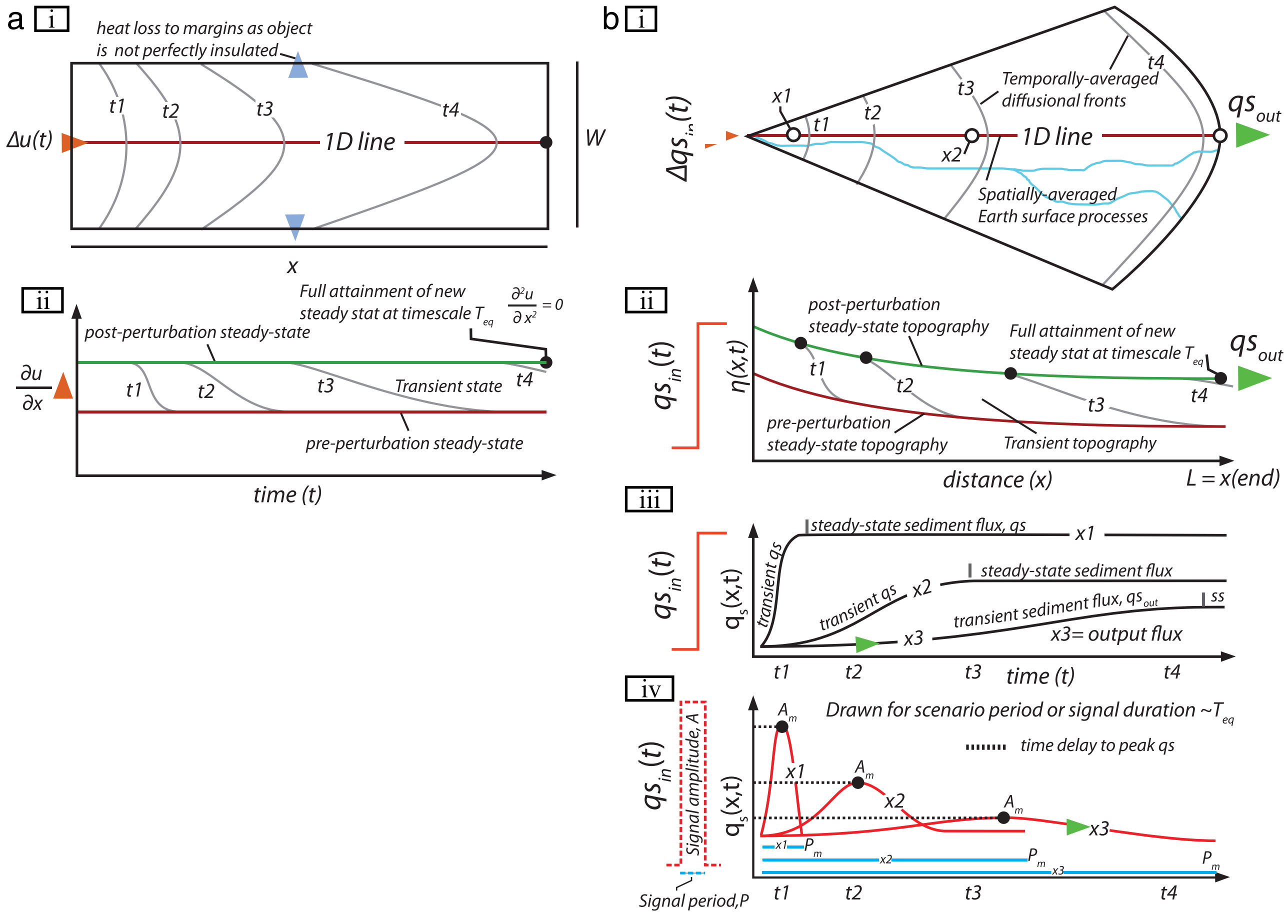

Figure 2. Schematic of processes and timescales important in diffusion of signals. (a)i The 2-D view of a homogenous and isotropic solid medium illustrating the diffusion of heat following a step change increase in the rate of heat input. ii. The time-dependent response of the system from the initial or preperturbation steady-state condition (red line) to the post perturbation steady-state condition (green line). Note that the system is still in transient response or transient state until the new steady state has been achieved. The time taken to reach this new steady state, or the response time T_{eq}, is related to the material properties, amplitude of perturbation and the dimensions of the solid medium. (b) i, ii. The 2-D Map view of a spatially and temporally averaged (so homogenous and isotropic) diffusional fluvial landscape showing the position of diffusional fronts and time-dependent response of the system following a step change in sediment flux, q_{sin}. iii. Time-dependent response, to this step change in terms of q_s, at three positions (x1, x2, and x3; see Bi) as a function of time. iv. Time-dependent response of the system, in terms of q_s, shown at three positions (x1, x2, and x3 see Bi) to a single spiked perturbation of q_{sin} of duration ~T_{eq}. Note the reduction in magnitude of the signal and the increase in duration of the signal as it travels downsystem. A_m, modified amplitude, P_m, modified period.

landscape threshold is surpassed (i.e., magnitude-frequency relationships) helps describe the stochastic environmental parameter and the discontinuous nature of erosion and deposition in time and space.

The stratigraphic recording process is further complicated by patterns and dynamics in SRSs that arise solely as a consequence of the interaction of the components within a system (Greenberg, 2016; Hajek & Straub, 2017; Muto et al., 2007; Paola, 2016). These autogenic (Table 1) dynamics are often discussed in comparison to allogenic processes (i.e., environmental forcings). Examples of autogenics abound. On the pattern end, they include the self-organization (Table 1) of bed and barforms (Ganti et al., 2013; Jerolmack & Mohrig, 2005; Myrow et al., 2018; Southard, 1991), tributary (Dodds & Rothman, 2000; Hasbargen & Paola, 2000; Rodríguez-Iturbe et al., 1992) and distributary (Coffey & Shaw, 2017; Edmonds & Slingerland, 2010; Jerolmack & Swenson, 2007) channel networks, and shoreline features like spits (Ashton et al., 2001) and barrier islands (Ciarletta et al., 2019). Autogenics have even recently been linked to the development of some waterfalls (Scheingross et al., 2019). Their dynamics include the cutoff of river bends (Howard & Knutson, 1984), river (Mohrig et al., 2000) and delta lobe (Jerolmack, 2009; Slingerland & Smith, 1998) avulsions, barrier island migration (Ciarletta et al., 2019), and formation of lakes (Kim & Paola, 2007). Together, these patterns and their dynamics produce stratigraphic products like ripple and dune trough cross stratification (Ganti et al., 2013; Paola & Borgman, 1991; Rubin & Hunter, 1983), channel clustering (Hajek et al., 2010), construction of some parasequences (Straub et al., 2015; Van Wagoner et al., 1990), and compensational stacking of geobodies (Straub et al., 2009).

Autogenics result in stochastic dynamics in sedimentary basins primarily because they promote the configuration of transport systems into narrow channelized corridors. This means that wide swaths of the landscape are in stasis at any one time, which generates stratigraphic hiatuses in the form of paraconformities (Sadler, 1981; Straub & Foreman, 2018; Tipper, 2015); previously deposited sediments can be eroded by the confined flow in channels, and episodes of channel aggradation and avulsion distribute sediment to inactive parts of the landscape. At mesotimescales, the spatially and temporally variable patterns of deposition and erosion rates, and the duration of resultant stratigraphic hiatuses, follow probabilistic distributions (Ganti et al., 2011; Martin et al., 2009).

3. Impediments to Environmental Signal Storage

In the following section we focus on the quantitative theory underpinning three impediments to environmental signal storage in stratigraphy. For each impediment we examine the numerical, experimental, and field findings that motivated and vetted theory development. Specific field application of theory then follows in sections 4 and 5.

3.1. Signal Buffering by Deterministic Surface Processes

We start our exploration of impediments to environmental signal storage in stratigraphy at the largest scale by characterizing deterministic processes that dictate how environmental signals propagate through SRSs across the Earth's surface and ultimately into the stratigraphic record.

3.1.1. Heat Diffusion as an Analogy for Landscape Evolution

A natural starting point is the use of the diffusion equation to describe the response and evolution of surface topography to a change in boundary conditions that influence the flux of sediment provided to a basin (Begin et al., 1981; Muto & Swenson, 2005; Paola et al., 1992). The diffusion equation describes how heat is distributed through a medium over time as it diffuses from locations of, for example, high temperature to locations of low temperature. The 1-D equation has the form:

$$\frac{\partial u}{\partial t} = \nu \frac{\partial^2 u}{\partial x^2} \tag{1}$$

where t is time, u is temperature, x is distance from the heat source, and ν is the diffusion coefficient. In the case of heat, the diffusion coefficient is defined as $\nu = \lambda/\rho c$, (λ, conductivity; ρ, material density; c, specific heat capacity) and describes the ability of the medium to conduct thermal energy relative to its ability to store it, which dictates the rate at which heat can spread through a solid medium. This equation describes the tendency of a system to distribute and smooth out any external temperature disturbances and internal temperature anomalies to attain thermal steady state (Figure 2a).

The diffusion equation is applied to a diverse range of systems where a property is conserved in one dimension and flows down a gradient according to a flux-gradient relationship (Slingerland & Kump, 2011). Solutions to the diffusion equation have been applied successfully to model alluvial fans, prograding deltas, eroding fault scarps, coastlines, hillslopes, and river long profiles (Flemings & Jordan, 1989; Paola, 2000 and references therein). The main difference among these applications is the definition of the diffusion coefficient, or transport coefficient, each of which requires a different set of parameters to capture specific properties of the system.

To capture the diffusive behavior of river profiles and the generation of key facies fronts and stratigraphic surfaces, the Exner equation for mass conservation:

$$\sigma + \frac{\partial \langle\eta\rangle}{\partial t} = -\frac{\partial q_s}{\partial x} \tag{2}$$

(where σ is subsidence rate, η is bed surface elevation, q_s is the mean sediment flux per unit width, and t is time; $\langle\ \rangle$ denotes time averaging) is combined with an algorithm for slope-dependent sediment flux,

$q_s = -\nu(\partial\eta/\partial x)$ (where ν is the transport coefficient) to generate the familiar-looking diffusion equation (Flemings & Jordan, 1989; Paola et al., 1992):

$$\sigma + \frac{\partial <\eta>}{\partial t} = \nu \frac{\partial^2 <\eta>}{\partial x^2} \tag{3}$$

The equation states that the rate of surface elevation change is described by the downsystem rate of change in the topographic gradient and the transport coefficient: This is analogous to heat diffusion being described by temperature gradients, thermal conductivity, and heat capacity. Topographic steady state then refers to the condition where elevation does not change as a function of time (Figure 2b).

For channelized river environments under transport-limited conditions, fluvial diffusion and the transport coefficient can be derived from mass (water and sediment) and momentum conservation under a set of physical transport laws (Marr et al., 2000; Paola, 2000; Paola et al., 1992). Fluvial diffusion spatially and temporally averages the effect of all stochastic fluctuations (e.g., variable water and sediment discharge) on bulk sediment transport and thus describes the average evolution of topography in a basin (Paola et al., 1999). The evolution of fluvial landscapes modeled this way captures the aggregate mass balance behavior of landscape diffusion and does not reflect in any real sense the diffusion of individual sediment particles and individual sediment transport events (see Paola, 2000).

Although primarily developed for longer-term geological problems (~10^6 yr), fluvial diffusion is utilized to conceptualize the response of recent alluvial plains and associated strata (Castelltort & Van Driessche, 2003; Métivier, 1999). These authors reasonably advocate that the transport coefficient of an alluvial plain approaching topographic steady state will scale with a constant, time-averaged output sediment flux $<Q_{st}>$, floodplain width W, and mean topographic slope $<(\partial\eta/\partial x)>$ (i.e., when in steady state; Figure 2b):

$$\nu = \frac{<Q_{st}>}{W<\frac{\partial\eta}{\partial x}>} \tag{4}$$

Metivier and Gaudemar (1999) intriguingly showed that values of $<Q_{st}>$, although representative of average output sediment flux values over longer timescales, are very similar to modern Q_{st} values for a number of large river systems, raising the possibility that large river systems may act as buffer zones, inhibiting the propagation of sediment flux disturbances across the alluvial plain and to the sea.

3.1.2. Buffering in SRSs

Changes in the supply of sediment at the upstream margin of a system, or changes in the downstream boundary elevation, will result in the downstream or upstream propagation of spatially decaying diffusional waves (Armitage et al., 2018; Paola, 2000). These waves act to regrade topography to a new topographic steady state in-line with the new boundary conditions (Figure 2). The time required to completely regrade topography to a new steady state is known as the basin "response time," "equilibrium time," or "basin filling time," T_{eq} (Paola, 2000; Paola et al., 1992) and scales as:

$$T_{eq} = \frac{L^2}{\nu} \tag{5}$$

Combining equations (4) and (5), Metivier and Gaudemer (1999) defined the response timescale as follows:

$$T_{eq} = \frac{L^2 W<\frac{\partial\eta}{\partial x}>}{<Q_{st}>} = \frac{LWH_{max}}{<Q_{st}>} \tag{6}$$

where L is the length of the transport system and H_{max} is the elevation of the alluvial plain at $x = 0$ (i.e., catchment-alluvial plain transition). With knowledge of key parameters, natural river systems yield values of T_{eq} ~ 10^5–10^6 yr (Paola et al., 1992; Dade & Friend, 1998; Métivier & Gaudemer, 1999; Castelltort & Van den Driessche, 2003).

For natural river systems, the implication of the reported range of T_{eq} values is that they are unlikely to attain a new topographic steady state (i.e., complete topographic regrading) when perturbed by

10.1029/2019JF005079

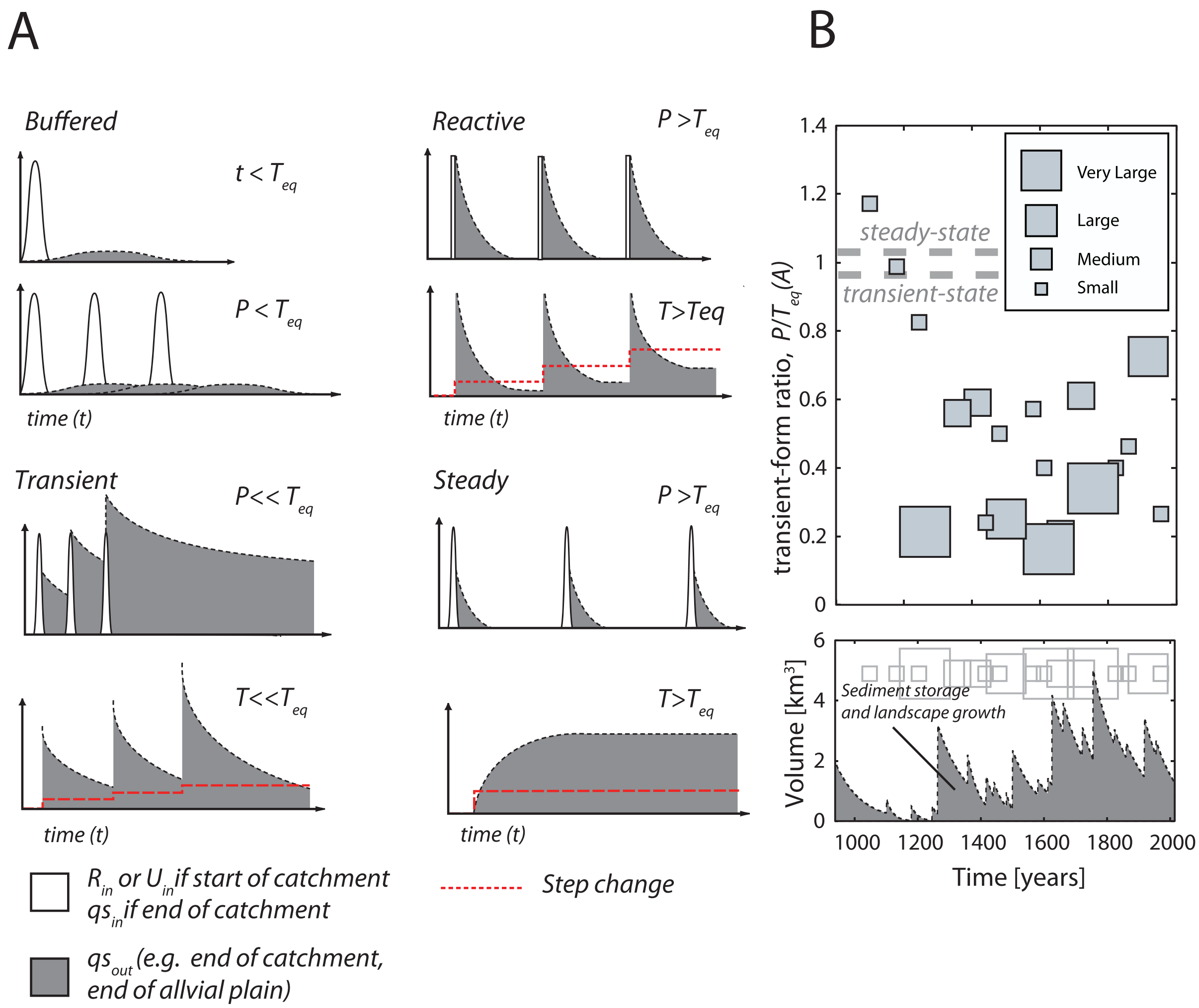

Figure 3. Terminology for defining landscape response to changes in boundary conditions. (a) In each case the equilibrium timescale, T_{eq}, is used as the fundamental comparator. P, period of perturbation; t, duration of perturbation; and T, time since last perturbation. For each case sediment flux at an outlet, either the catchment outlet or the alluvial plain outlet. Red dashed lines represent a step change in perturbation. Note that while buffered systems show a delayed and reduced response, reactive system shows an immediate response and an exponential reduction in amplitude. (b) Transient-form ratio (Brunsden & Thornes, 1979) for a landscape that has received massive amounts of sediments from glacier outburst events over the past 1,000 yr, occurring at a frequency of two events per 100 yr. Using the known response time, T_{eq}, of the landscape to events of different magnitudes, the transient growth of the alluvial plain could be predicted, which had implications not only for landscape response and recovery but also for sediment export offshore. Filled gray region in the lower figure represents the volume of sediment stored above mean sea level, which demonstrates a long-term transient growth of the landscape (proglacial outwash plain). This landscape is both highly reactive and transient. T, time since last event. From Duller et al. (2014).

environmental forcing of period (or duration), $P < T_{eq}$. These landscapes therefore likely act as buffers (Table 1) to environmental signal propagation across the Earth's surface (Métivier, 1999) and into the stratigraphic record (Castelltort & Van den Driessche, 2003; Allen, 2008). We note that if the time period of forcing, P, or the duration of forcing is greater than a landscape response timescale, the landscape is described as *reactive* (Figure 3) (Allen, 2008; Covault et al., 2013). Recalling that diffusion does not model sediment particles or landscapes in any real physical sense, what then is the potential cause of signal buffering in real systems? The buffering agent of a sediment flux signal in real systems is the composite redistribution of sediment mass across the Earth's surface, integrated over a range of spatiotemporal scales that include the net effects of intermittent sediment transport and temporary or permanent deposition of sediment within distinct landforms (e.g., river bars, floodplains, alluvial fans, and catchment valleys). It is the cumulative effect of these processes, particularly the temporary storage and subsequent remobilization (or reworking) of sediment that enables landscapes to buffer environmental signals such as

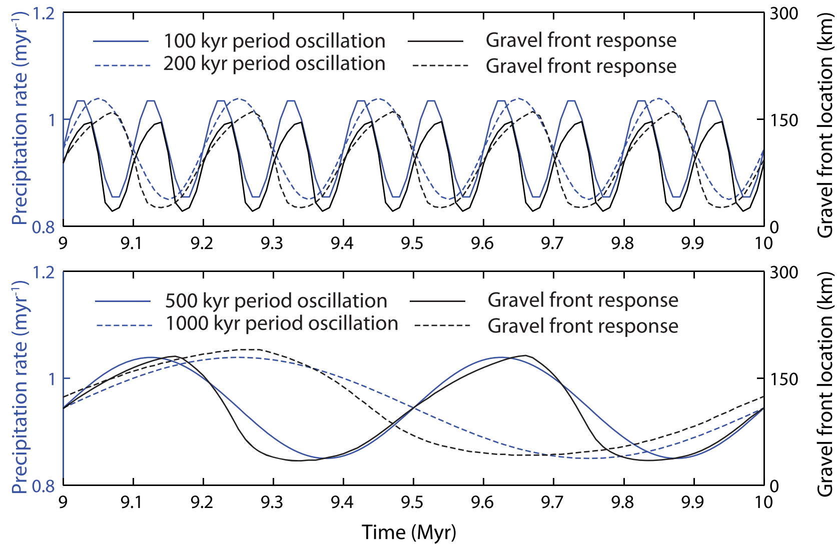

Figure 4. Response of the gravel front (black lines) to oscillation in precipitation (blue lines) with periods of 100, 200, 500, and 1,000 kyr. Note a phase shift between the period of forcing and the periodic gravel front response, which appears to be longer for shorter period oscillations in precipitation rate. After Armitage et al. (2018).

changes in sediment supply (e.g., Jonell et al., 2018; Métivier, 1999). Revisiting the diffusion analogy, buffering of a heat input signal as it travels through a solid medium is dependent on the ability of the solid medium to transmit heat (conductivity) and the ability of a solid medium to store it (specific heat capacity). Similarly, the efficiency with which sediment can be distributed across a landscape (e.g., increasing efficiency with increased water supply) and the potential for sediment storage in a landscape (e.g., larger landforms provide more temporary storage) influence sediment flux buffering in SRSs.

Landscape buffering processes have the effect of reducing a signal magnitude, that is, damping, at a site of recording (Figure 2b), relative to either the input magnitude or to an expected response for a system that is able to achieve a new steady state (Allen, 2008; Romans et al., 2016). In addition to the reduction of signal magnitude, buffering processes tend to increase the timescale over which a response is observed in a record, relative to the actual timescale of a perturbation (Figure 2b). A signal like a change in the erosional sediment flux with amplitude, A, and period, P, that is transmitted at the upstream end of a river system will reach the outlet of the system (e.g., the sea) as a transformed signal with modified amplitude $A_m \sim A(P/T_{eq})$ and modified period $P_m \sim T_{eq}$ (Métivier & Gaudemer, 1999). Under a scenario of perfect basin mass balance, the downstream reduction of input signal amplitude is proportional to the spatial distribution of deposition, becoming zero at the end of the system, and the modified duration or period must be equal to T_{eq}. So diffusion has the effect of smoothing out the transmitted signal, as the landscape is regraded via erosion or deposition (Figure 2b).

A complementary response timescale for erosional, uplifting catchments, T_{RT}, can be defined through the use of the stream power model for detachment limited incision (Armitage et al., 2018; Whipple, 2001; Whittaker & Boulton, 2012), which describes the time necessary for signals of changing rock uplift rates or climate to propagate fully through a catchment (Densmore et al., 2007; Gasparini et al., 2007; Tucker & Bras, 1998; Whittaker et al., 2007).

The determination of T_{eq} or T_{RT} is important as it represents the fundamental landscape measure, from which we are able to predict longer-term landscape behavior and response to single or periodic perturbations in boundary conditions. This is accomplished, quite simply, through comparison of T_{eq} or T_{RT} with different classes of perturbation timescales [Figure 3, Allen, 2008; Duller et al., 2014].

3.1.3. Transient Response of Buffered Systems

While the framework provided above suggests long response times (10^5–10^6 yr) of natural systems, there is sedimentological evidence of landscape response to shorter environmental forcing (Blum et al., 2013;

Goodbred, 2003; Romans et al., 2009; Watkins et al., 2018). The reason for this, as shown through numerical modeling by Snow and Slingerland (1990), is that while full adjustment to a new steady state (response) is well estimated by T_{eq}, fluvial adjustment begins immediately [sensu Bull, 1991; Duller et al., 2014] after a perturbation in forcing conditions (Figures 2 and 3). As topography approaches a new steady state, it does so asymptotically (Allen, 2008; Duller et al., 2014) suggesting that this transient response has the potential to be stored in strata (e.g., Shen et al., 2012), if operative over long enough timescales, even if full adjustment has not been reached.

The idea that parts of the Earth's surface will respond instantaneously and transiently (e.g., Anderson & Konrad, 2019) to a perturbation is not new but is somewhat overlooked by the field stratigraphic community when using system scale T_{eq} estimations to assess the likelihood of signal retention in ancient successions. As an example, a decrease in the sediment flux from the catchment outlet will cause a river to incise into the surrounding floodplains, whereas an increase will cause aggradation of floodplains (Métivier, 1999; Paola, 2000). Each scenario has a demonstrable local effect on the proximal depositional landscape, and therefore on stratigraphy, but in each case, somewhere downstream, a constant output sediment flux is maintained and so the output sediment flux signal is buffered (Métivier & Gaudemer, 1999; Castelltort & Van den Driessche, 2003). This was demonstrated earlier by the diffusional stratigraphic models of Paola et al. (1992) and Marr et al. (2000), who show that rapid (i.e., $P < T_{eq}$; "buffered") changes in sediment flux and water flux will produce demonstrable stratigraphic signatures.

Interestingly, Armitage et al. (2018) used a diffusion model to show that periodically changing precipitation might influence progradation of the gravel-sand transition (a region of abrupt grain size reduction common in fluvial SRSs and identifiable as a facies transition in the sedimentary record). Armitage et al. showed that, while there is a delay between maximum precipitation rate and the maximum progradation of the gravel front, which is not unexpected from a diffusional system, high-frequency oscillations in precipitation rate are recorded in the movement of the gravel-sand transition with the appropriate periodicity (Figure 4). Therefore, the response of the gravel front to high-frequency changes in precipitation rate is a stratigraphic record out of phase but not buffered.

A diffusional framework provides a basis for estimating the theoretical downstream distance a periodic input signal can propagate through a SRS and be incorporated into the stratigraphic record. The downstream distance at which the amplitude of a periodic input signal can affect the Earth's surface can be described as $x < (P\nu)^{0.5}$ (Paola, 2000), and the buffer distance, which describes the distance from a source at which the amplitude of a periodic input signal is reduced by one third of its initial value, can be estimated as $B_d = (\nu P/\pi)^{0.5}$ (Castelltort & Van den Driessche, 2003). Whether a system is described as buffered or reactive depends not just on the properties of a particular landscape but also on the amplitude and duration or period of the environmental signal (e.g., Allen, 2008).

A comparable timescale (T_{FS}) to T_{eq} described above was developed by Muto and colleagues (Muto & Steel, 2004; Muto & Swenson, 2006) to predict when a nonequilibrium, autostratigraphic response in a delta system will dominate:

$$T_{FS} = \frac{D^2}{\nu} = \alpha \frac{q_s}{{\eta_{sl}}^2} \tag{7}$$

In equation (7), D is a characteristic length scale equal to q_s/η (q_s, constant sediment supply rate; η, constant rate of sea level change), ν is the fluvial transport coefficient, and α is a characteristic slope of the fluvial surface equal to q_s/ν. This work demonstrates that a fluvio-deltaic system will produce an unsteady "autostratigraphic response" even under conditions of steady allogenic forcing. For a duration of sea level rise or fall, T, a fluvio-deltaic system will have a nonequilibrium response if $T >> T_{FS}$ and will have an equilibrium response, that is, steady forcing is closely tied to stratigraphic architecture, when $T << T_{FS}$. Without awareness of T_{FS}, stratigraphic analyses of marginal marine successions could easily overestimate or underestimate the timing and magnitude of interpreted environmental forcings (Muto et al., 2007).

3.1.4. Signal Propagation Through Catchments

Similar to the autostratigraphic concepts at play on the distal end of SRSs, at the proximal end catchments tend to respond to environmental forcing in complex ways (e.g., Allen, 1974; Coulthard & Van de Wiel, 2013;

Schumm, 1979; Walling, 1983). In the absence of an observable catchment or evidence to the contrary, an implicit assumption made by field investigators [i.e., Duller et al., 2010; Duller et al., 2012; Ventra & Nichols, 2014; Armitage et al., 2015; Watkins et al., 2018] is that catchments are reactive. This assumes that environmental forcing information is perfectly transcribed into environmental sediment flux information. This assumption might be construed as a legitimate and unavoidable one but still remains incorrect. However, given that catchments also have an intrinsic response timescale (T_{RT}), sediment flux out of catchments affected by periodic environmental forcing is not straightforward (e.g., Li et al., 2018). Clearly, this is an important consideration because if stratigraphers hope to reconstruct a sediment flux signal related to a single step change in forcing or periodic forcing, then this sediment flux signal must first make it out of the catchment. We note that the reconstruction of a single sediment flux, let alone time-varying values, from ancient catchments using strata is a formidable task (Allen et al., 2013).

Recently, numerical models and physical experiments have been used to investigate the response of catchments to stepwise forcing and periodic forcing of climate and tectonics, in terms of a time series of sediment flux. These studies highlight that catchments respond transiently to forcing conditions with response time of the order of $0.5–1 \times 10^6$ yr, which is reflected in the stratigraphic signature of the adjacent basin. When the forcing period is less than the response time of the catchment (i.e., $P < T_{RT}$), the interaction of forcing conditions with transient catchment topography induces a complex and nonuniform feedback, which dampens or buffers sediment flux signals exiting catchments (Figures 5 and 6). This means that when $P < T_{RT}$, it would be difficult to extract forcing signals from deposit volumes and grain size trends of strata in adjacent sedimentary basins (Armitage et al., 2011; Armitage et al., 2013; Armitage et al., 2018; Densmore et al., 2007).

3.1.5. Summary and Implications of Diffusion and Buffering Impediments

Parameterizing channelized fluvial systems and catchments as diffusive systems provides a framework from which key time and length scales can be compared and used to explore how a sediment mass flux signal will propagate through a SRS. Given a specified environmental signal, this helps us predict (1) the capacity of a SRS to store sediment at locations along its length and therefore how much sediment is bypassed to locations downsystem; (2) the expected amplitude of the environmental signal and expected duration over which locations along a SRS will be affected; and (3) the propagation distance of these disturbances through a SRS (including catchments) and the capacity of a SRS to buffer against step changes and periodic environmental signals.

The representation of landscapes as diffusive does not explicitly incorporate the specific processes and landscape dynamics that give rise to the diffusion of sediment through SRSs (Hajek & Straub, 2017; Hajek & Wolinsky, 2012; Malmon et al., 2003; Métivier, 1999; Paola, 2016; Parker, 1978; Phillips & Jerolmack, 2016; Simpson & Castelltort, 2012). Therefore, diffusional representations of SRSs over mesotime scales implicitly assume (an unspecified degree of) spatiotemporal averaging in response to an environmental signal. Adding an element of stochastic noise into the standard diffusion model ("noisy diffusion") can mimic the inherent stochastic variability of natural system-scale processes and successfully reproduces a number of natural sedimentary system behaviors, such as the temporal scaling of surface elevation change, the distribution of hiatuses and bed thicknesses, and the temporal scaling of stratigraphic completeness (Jerolmack & Sadler, 2007; Pelletier & Turcotte, 1996, 1997). Noisy diffusion bridges the interaction between deterministic and stochastic processes in natural landscapes and highlights their importance in mediating signal propagation and preservation in SRSs.

3.2. Stratigraphic (In)completeness

Signals of environmental change that propagate to a depocenter without severe damping still face hurdles for stratigraphic storage. Stratigraphers have long known that at some scale all stratigraphic sections are incomplete: riddled with time gaps or hiatuses (Ager, 1973; Barrell, 1917; Miller, 1965). Unfortunately, the sedimentary expression of missing time is subtle and can be difficult to identify (Boulesteix et al., 2019; Trabucho-Alexandre, 2015). Further, quantifying missing time, stratigraphic (in)completeness (Table 1), at mesotimescales in specific sections is challenging, if not impossible in deep-time successions, due to resolution limitations in geochronometers. Even without erosion, we know that all records must have some temporal gaps. Sediment particles, which have finite granularity, arrive at sites of deposition instantaneously and thus have infinite deposition rates at that moment. These flashes of deposition must be balanced by long periods of stasis, or periods in which the sediment interface is inactive and thus neither aggrading or

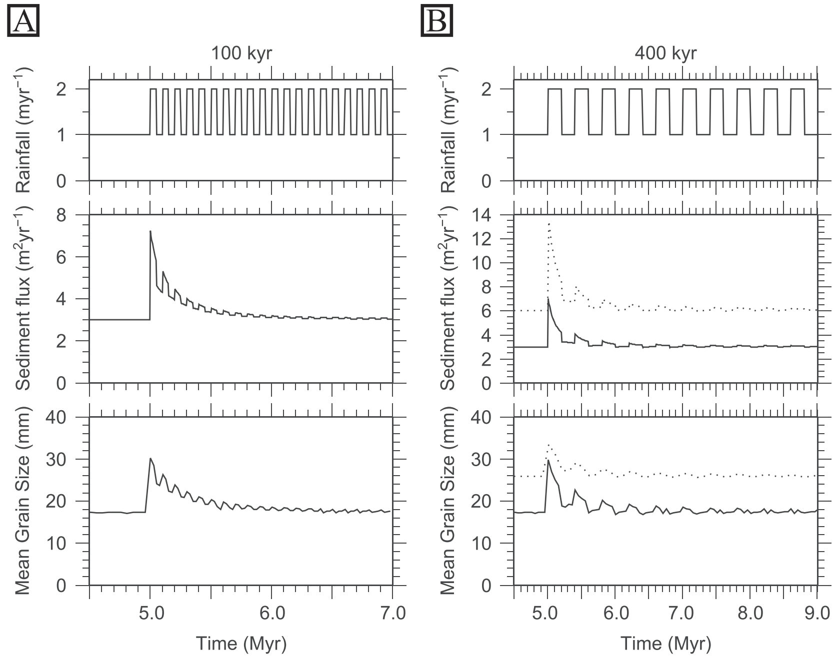

Figure 5. Diffusional response of catchments. Response in terms of sediment flux and mean grain size to oscillations in precipitation rate with periods (a) 100 kyr and (b) 400 kyr (dotted lines show sediment flux and mean grain size for a catchment and basin twice as long). After Armitage et al. (2013).

degrading, to give us the deposition rates we commonly measure (McElroy et al., 2018). At longer timescales sediment transport dynamics (e.g., bedform migration and channel avulsion) cause further pulses of deposition and stasis (and therefore (in)completeness) throughout SRSs. Such flashes of deposition are best described probabilistically (Einstein, 1950; Furbish et al., 2012).

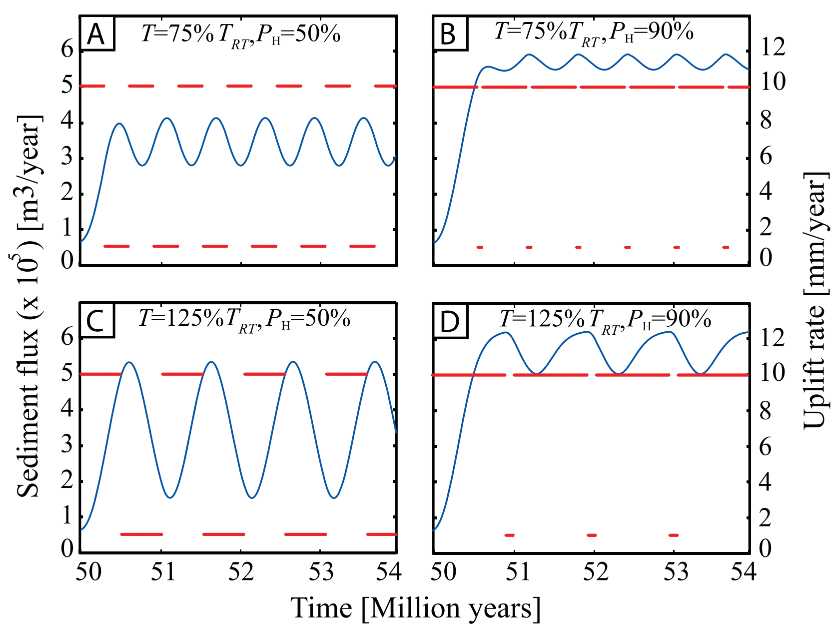

Figure 6. (a–d) The advective response of catchments, in terms of a time series of sediment flux to oscillations in uplift, scaled to a catchment response time, T_{RT}. Red lines represent the time series of uplift, and blue lines represent sediment flux at the catchment outlet; the catchment response proxy. P_H, percentage of the period during which uplift rate is high. After Li et al. (2018).

10.1029/2019JF005079

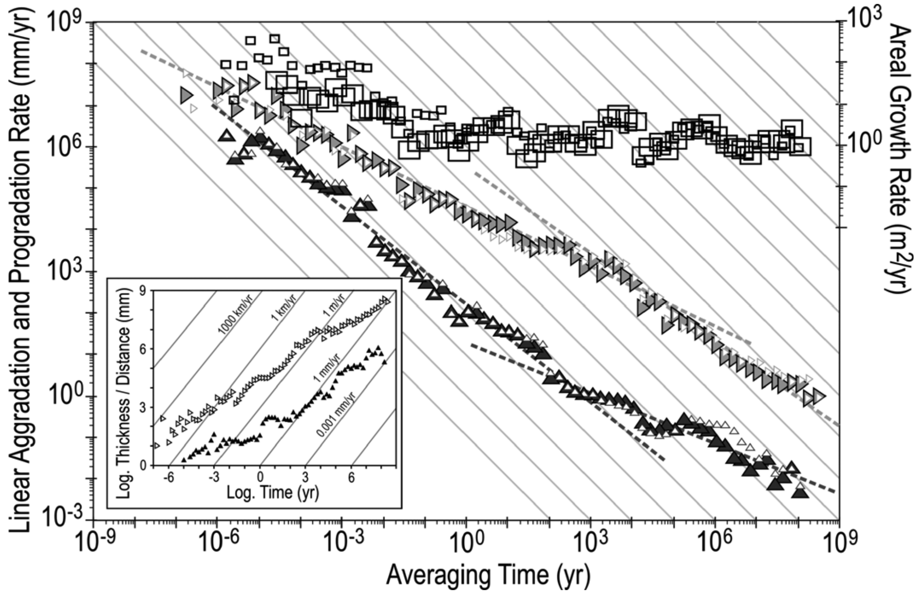

Figure 7. Compilation of 1-D aggradation and progradation rates and expected sediment flux as a function of measurement duration. Reproduced from Sadler and Jerolmack (2015) with permission of authors. Upward-pointing triangles plot aggradation rates, while forward-pointing triangles plot progradation rates. Expected sediment flux, plotted as squares, expressed as volume discharge per unit width of the transport system does not vary systematically from timescales of months to 100 million years. These growth rates are not empirical field estimates; they are products of expected aggradation and progradation components determined separately from mean empirical values. Inset: Unsteady increase of expected progradation distance and aggradation thickness as a function of time span.

The first study to quantify stratigraphic (in)completeness from field observations was conducted by Peter Sadler (1981). Given the difficulty in identifying and measuring all hiatuses directly, Sadler measured (in)completeness through an indirect method that took advantage of a global database of deposition rates (Figure 7). Analysis of this database showed that aggradation rates decrease as a power law function of the measurement timescale and Sadler linked this to stratigraphic (in)completeness. He noted that deposition rates measured over short durations, for example, direct observations over hours, were less likely to contain hiatuses compared to rates calculated from dated horizons separated by millions of years. Further, results from numerical (Schumer & Jerolmack, 2009) and physical (Ganti et al., 2011) experiments suggest that distributions which describe the durations of stratigraphic hiatuses are heavy tailed and thus the chance for inclusion of an exceptionally long hiatus increases as the duration of observation increases. Qualitatively, Miall (2015 and references therein) linked a component of the timescale of measurement rate dependence to a hierarchy of depositional processes. For example, bedforms are short-lived features but have relatively high deposition rates in comparison to the channels they reside in; the coexistence of these morphodynamic features inherently imparts multiple superposed sources of (in)completeness in a SRS.

The issue of stratigraphic (in)completeness has obvious ramifications for the detection of environmental signals in stratigraphy. Put simply, if sediment of a given age is not present in a section, it is difficult to infer environmental conditions for that time. This is particularly important for the construction of stratigraphic age models, in which sediment age is often assigned by linear interpolation between sparsely dated horizons or biozones (Abels et al., 2010; Vázquez et al., 2017). This challenge is especially vexing in one-dimensional (1-D) sections of strata where observations of lateral changes are often difficult to make. We focus first on theory and quantification of stratigraphic (in)completeness in 1-D as many records come from vertical sections, whether they originate from analysis of cores or measured outcrop sections (e.g., Aswasereelert et al., 2013; Aziz, Hilgen, et al., 2008; Flower et al., 2004; Shen et al., 2012).

Because of limits in geochronology, it is difficult to directly measure stratigraphic completeness at mesotimescales and shorter; consequently, we have little theory, benchmarked by observations, to predict its magnitude in different sedimentary systems. This is starting to change as numerical models and field observations are shedding light on the link between morphodynamics and the storage of time in stratigraphy (e.g., Bhattacharya et al., 2019; Davies et al., 2019; Durkin et al., 2018; Xu et al., 2016). Intuitively, however, it makes sense that the more strongly channelized or intermittent a SRS is, the more opportunity there is for long-term hiatuses to form on inactive parts of a landscape.

3.2.1. Exploration of Signal Storage in Physical Experiments

Laboratory experiments are particularly useful for capturing self-organization and complex stochastic behavior that is difficult, if not impossible, to decipher in modern landscapes. Key advantages of laboratory experiments for stratigraphic studies are threefold. First, landscapes generated in laboratories are small, (<10 m^2 in planview) and so it is possible to monitor them comprehensively and dissect their stratigraphy in 3-D with high precision. Second, as landscapes shrink in size, the timescales of key processes also generally shrink. This allows stratigraphic packages, which are many channel depths in thickness, to be constructed in days to months. Third, bringing SRSs into laboratories allows forcing conditions to be independently controlled, thus allowing targeted experimental campaigns that can isolate the importance of single variables.

A thorough review of the philosophy and methodology of upscaling these results to typical field scales is beyond the scope of this paper but can be found in several publications (Kleinhans et al., 2014; Malverti et al., 2008; Paola et al., 2009). Here we use results of physical experiments to highlight how many aspects of SRSs are scale independent, including the process of channelization, which allow experiments to be "unreasonably effective" at capturing dynamics important for the construction of the stratigraphic record (Paola et al., 2009). In addition, laboratory experiments aid the development of intuition on the processes associated with the self-organization of channel networks that migrate and avulse since they can be observed in real time.

A suite of stratigraphic experiments performed in the Tulane University Delta Basin (TDB) provides a good baseline for evaluating how fluvial-deltaic landscapes self-organize under different boundary conditions. These experiments were performed with identical forcing parameters (input water and sediment supply history, input sediment characteristics, sea level rise history, etc.) with the exception of one variable that was altered per experiment. This allows the isolation of cause and effect to varying a single parameter. In each experiment, water and sediment are fed into a basin with a standing body of water. The resulting self-channelized fan deltas experience a background constant sea level rise that mimics a spatially uniform subsidence pattern and promotes the development of thick stratigraphic sections.

3.2.2. Linking the Sadler Effect to Stratigraphic (In)completeness in Experiments

We demonstrate a "Sadler effect" with a fan delta that freely evolved under its own internal physics as a result of constant inputs of water and sediment and uniform generation of accommodation through steady base level rise (similar to steady and spatially uniform subsidence) (Figure 8). The surface of the fan delta over time was monitored with high-resolution imaging and elevation scans. Experimental morphodynamics, similar to geomorphic processes operating in field-scale systems, produce elevation time series with inherent correlation resulting from the movement of coherent landforms. As a result, periods of stasis, deposition, and erosion are the direct result of the experimental morphodynamics (Figure 9).

From the 3-D experimental data set we extract a strike-oriented cross section (Figures 8c and 8d) and calculate all preserved deposition rates (rate > 0) as a function of timescale of measurement, δt. We generate a dimensionless timescale of measurement using the compensation timescale, T_c, sometimes referred to as the integral or saturation timescale. T_c represents an estimate of the maximum timescale of autogenic organization in stratigraphy (Sheets et al., 2002; Wang et al., 2011) and is estimated as

$$T_c = \frac{l}{r} \tag{8}$$

where l is the maximum autogenic vertical roughness scale on a landscape, often equal to a maximum channel depth, H_c, and r equals the long-term aggradation rate. We will return to the importance of this timescale later on, but here we note that T_c approximates the maximum time necessary to bury a particle to a depth below the reworking zone, such that it is no longer susceptible to erosion from autogenic processes (Ganti et al., 2011; Schumer & Jerolmack, 2009; Straub & Esposito, 2013; Straub & Foreman, 2018). We started by calculating deposition rates with $\delta t/T_c$ set to the temporal resolution of data collection, which is the

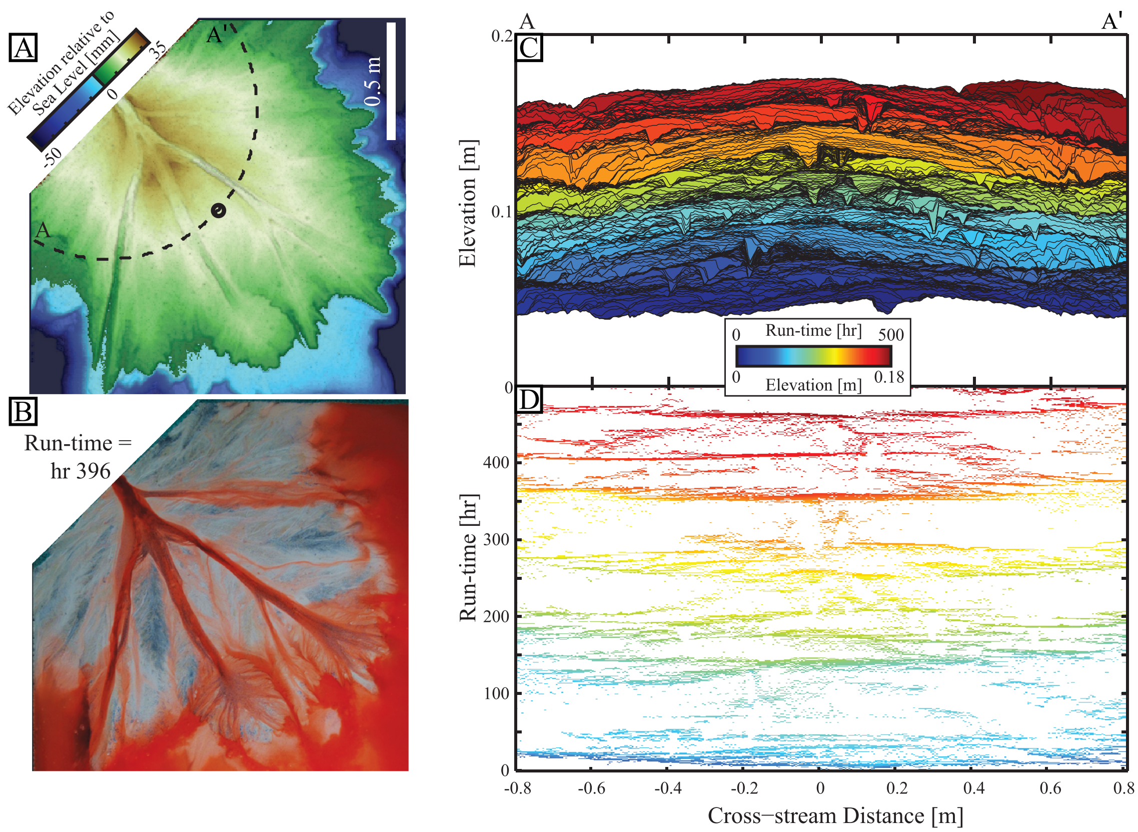

Figure 8. Experimental data used to explore stratigraphic (in)completeness. The experimental deposit, referred to as TDB-13-S2 (Li & Straub, 2017), evolved in response to a constant supply of water and sediment and a constant rate of sea level rise. Dynamics were monitored at high spatial and temporal scales relative to morphodynamics with a laser scanner that both captures elevations of the delta and coregistered images of the active surface with flow dyed for visualization. The total run time of the experiment was 500 hr, and topography was measured every 1 hr. (a) Digital elevation model (DEM) and (b) image of the experimental surface at run hour 396. Dashed line A-A′ indicates location of transect extracted for analysis. (c) Cross section of synthetic stratigraphy generated from stacked DEMs, clipped for erosion. Deposits are colored by time of deposition. Solid black lines are contours of constant time. (d) Time-space map of preserved elevation for the cross section presented in panel c. White regions in the map represent time-space pairs where either stasis or erosion resulted in a lack of preserved time. A time-lapse movie of the evolution of the experimental surface and flow field can be found online (https://www.youtube.com/watch?v=E3CDyeBahYg&t=43s).

finest temporal resolution we can explore. We then systematically coarsened the temporal discretization to a final resolution equal to $10T_c$. Deposition rates are normalized by an expected rate, equal to the imposed base level rise rate, which sets the accommodation production. Up to a timescale equal to the compensation timescale, we observe a similar power law decay in deposition rates with measurement duration as observed in Sadler's (1981) compilation (Figure 10a). However, deposition rates approach the expected rate at T_c, where they remain for longer measurement durations. No hard saturation scale exists in Sadler's data set as (1) it is a global compilation of data from different sites and thus incorporates the complete span of saturation timescales determined by the physical scales of the systems and (2) field systems in Sadler's compilation were subjected to both autogenic and allogenic hiatuses, of which the latter can have very long timescales. However, Jerolmack and Sadler (2007) did note a transition from transient to more persistent rates of sedimentation that occurs around typical field system compensation timescales.

We compare our "Sadler effect" plot to a similarly constructed plot of stratigraphic completeness (Figure 10b). The time of deposition for strata in our experiment is known to the temporal resolution of our data collection, which allows us to identify all stratigraphic hiatuses with durations equal to or longer than our sample interval. We quantify stratigraphic completeness, f_c, as the fraction of time intervals in a section, with a given temporal discretization (δt), that preserves sediment over the total duration of the 1-D section. Mathematically, this is expressed as follows:

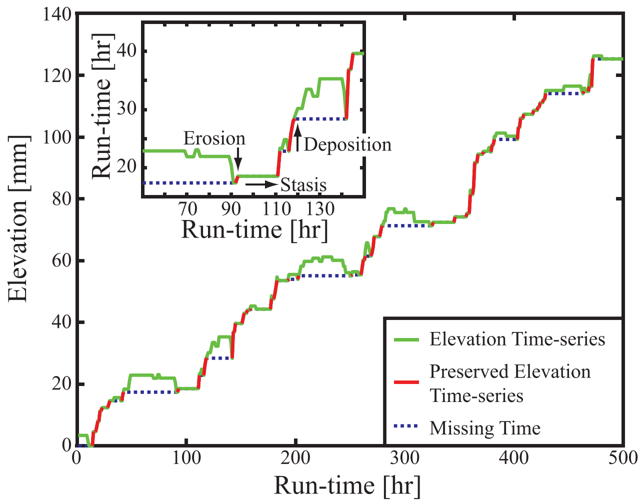

Figure 9. Data defining the 1-D evolution of topography and resulting stratigraphy for a system that exhibits long-term net deposition but over shorter periods experiences discrete episodes of deposition, erosion and stasis. The 1-D transect is extracted from midpoint of topographic data set presented in Figure 8c.

$$f_C = \frac{n\delta t}{T} \tag{9}$$

where n is the number of time intervals within a 1-D stratigraphic section discretized at δt, which leave a record in the form of preserved sediment over the length of a section that has a total time, T. From this one can see, as Ager (1973) and Sadler and Strauss (1990) noted, that f_c depends on the temporal discretization that one seeks to achieve. Specifically, f_c increases as a power law function of the dimensionless timescale of discretization, until saturating at 100% when $\delta t/T_c \geq 1$.

This analysis shows that the first-order control on the (in)completeness of a stratigraphic record is related to the timescale at which the record is discretized; it will always be more difficult to know what happened during every second of a basin's evolution compared to knowing some information about what happened during 1,000 yr increments. However, when analyzing strata from a particular basin, measured deposition rates and stratigraphic completeness both saturate when discretization timescales exceed the basin's maximum autogenic timescale. This maximum autogenic timescale is an emergent (Table 1) value set by how a system configures its surface roughness and deposition rates. This underscores the importance of emergent length and timescales within a SRS in setting the fidelity of the stratigraphic record, rather than any absolute spatiotemporal scales.

Estimates of maximum autogenic timescales, and thus the timescale of discretization necessary to obtain a complete stratigraphic record, have been made with the compensation timescale. Straub and Wang (2013) estimated compensation timescales for 13 river deltas around the world using reported maximum channel depths (or mean depths when maximum depths could not be found) and long-term deposition rates. Estimated values of T_c vary between 6 and 278 kyr (Figure 11). We note the overlap of the estimated autogenic timescales with the timescales of many mesotimescale environmental forcings (e.g., Milankovitch-forced climate cycles). Basins with T_c estimates in excess of an environmental forcing timescale of interest are then expected to contain incomplete forcing records.

3.2.3. Stratigraphic (In)completeness and Our Ability to Estimate Time

The uneven preservation of time, resulting from intrinsic depositional variability and episodes of stasis, warps the representation of environmental signals in 1-D sections. Several recent studies leverage 1-D random walk models with drift of surface elevation to explore the ability of stratigraphic records to capture climatic events (Kemp & Sexton, 2014; Trampush & Hajek, 2017). Similar 1-D random walk models have a long history in sedimentology and have been used to explore the preservation of time in strata and controls on bed thickness distributions (Kolmogorov, 1951; Schumer et al., 2011; Straub et al., 2012; Tipper, 1983). Kemp and Sexton (2014) used a random walk model, parameterized by geological data from deep marine records, to explore the preservation of abrupt events in the stratigraphic record. These events, classified as millennial or less in duration, are rarely temporally constrained in deep-time records given the limits of our geochronometers. Kemp and Sexton (2014) explored deposition rates and found that estimating the duration of millennial events through simple linear partitioning of time in their models produced significant errors, even in astronomically tuned and unbioturbated successions.

We explore inherent limits on our ability to constrain the timing of events between dated horizons by comparing the preservation of time in two fan delta experiments that varied strongly in their type of surface dynamics and their rate of evolution. The first experiment is the same used to characterize stratigraphic completeness above. We analyze each 1-D location along the strike transect and identify the age of the basal and topmost deposit and the total section thickness. These values are used to generate a linear age model for the total span of the section, similar to what might be constructed from sparsely dated horizons in the field, for example, dated biozones. As we know the exact age of all strata to the temporal resolution of our topographic scans, we quantify the error between the actual deposit age and the linear age model, T_{AE}, for each 1 mm as we move up section, which is used to generate a distribution of age errors (Figure 12).

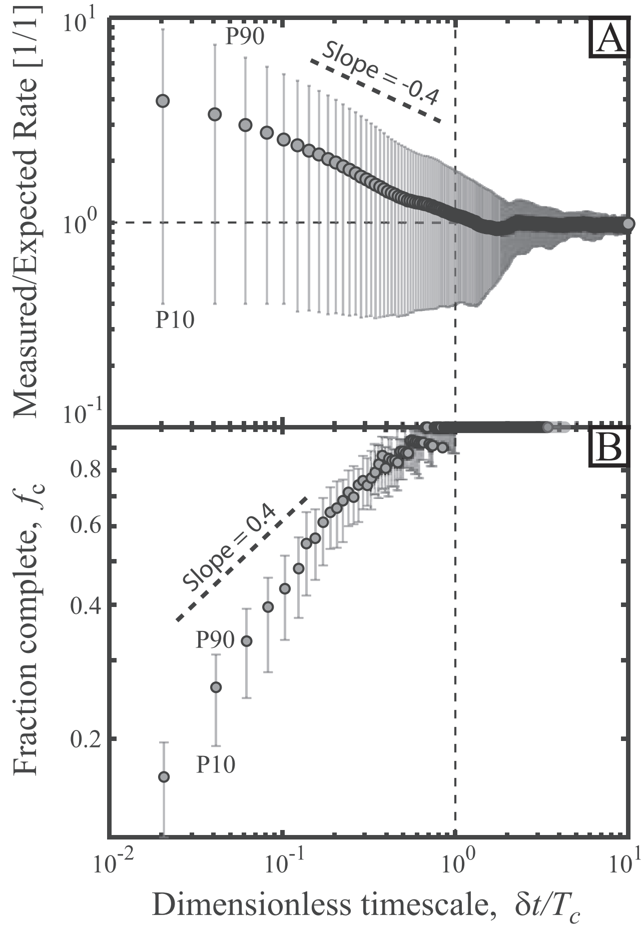

Figure 10. Data defining the Sadler effect and stratigraphic (in)completeness measured from experimental stratigraphy. (a) Rate of deposition as a function of duration of observation measured from synthetic stratigraphic data set shown in Figure 8c. Measured rates are normalized by an expected rate which equals the rate of accommodation production facilitated by long-term base level rise. Duration of observation is normalized by the compensation timescale, T_c. Symbols represent mean value of all positive rates measured at a given timescale, while vertical bars represent ± 1 standard deviation of distribution that defines the variability in the rate at that timescale. (b) Completeness of stratigraphic data as a function of discretization timescale. Symbols represent mean completeness value at a given timescale from all 1-D sections in data set, while vertical bars represent ± 1 standard deviation of distribution that defines the variability in completeness from all 1-D sections. A movie that documents the generation of strata along the strike section shown in Figure 8c colored by deposit age and corresponding Wheeler diagram that denotes when and where time is stored in the section can be found online (https://www.youtube.com/watch?v=4uaVo_EdaB0&t=6s).

The second experiment also evolved under constant forcings and a constant rate of accommodation production (TDB-10-1). However, in this experiment the accommodation production was 20 times higher than the first experiment and the sediment feed rate was also significantly higher, resulting in a braided channel configuration. As the long-term aggradation rates are known and the emergent surface roughness in each experiment can be measured, we can normalize age errors by T_c ($T^*_{AE} = T_{AE}/T_c$). Importantly, we find similar distribution shapes with maximum age error characterized well by twice the standard deviation of each distribution, which is approximately $0.8T_c$ (Figure 12). We suggest that this error is characteristic of any strata constructed by surface dynamics with laterally migrating roughness features, be they channels, bedforms, or features in between. We thus recommend including this uncertainty in any formal error analysis of the timing of signals recovered from channelized strata, particularly for sections in which signals are first identified in the spatial domain (e.g., identification of signal period in meters per cycle)and then converted to the temporal domain based on estimated long-term deposition rates from dated ashbeds or biochrons (Aziz, Di Stefano, et al., 2008; Abels et al., 2013). This error is also likely a minimum estimate as systems with

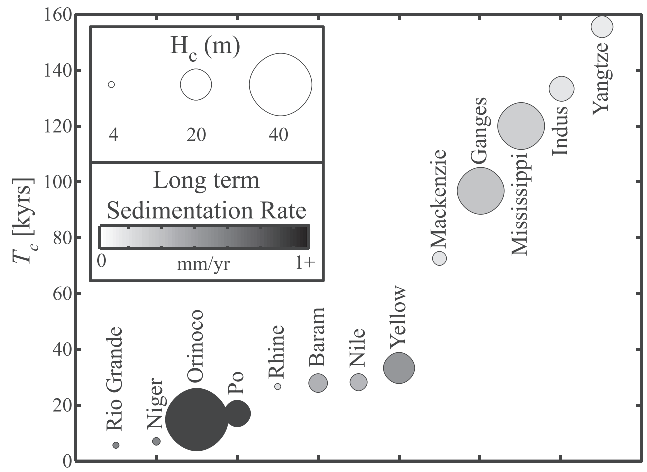

Figure 11. Compensation timescale estimates for compilation of 13 modern delta systems. Estimates calculated with published data on channel depth and long-term deposition rate (measurement interval in excess of 100 kyr). Size of circle scales with the depth of a system's channels while grayscale value of filled circles scales with the long-term sedimentation rate of a basin. Modified from Straub and Wang (2013).

intermittency in system activity, for example, deepwater fans constructed by infrequent turbidity current activity (Pirmez & Imran, 2003), will have greater age errors. Constraining and understanding how spatial and temporal intermittency in SRSs vary among different environments is an important outstanding challenge.

3.2.4. Detection of Signals From Incomplete Proxy Records

Trampush and Hajek (2017) demonstrated that a relatively simple input signal to a sedimentary environment, in this case a geochemical signal associated with the PETM, can be substantially altered by stochastic sedimentation in different environments. The PETM was a mesotimescale climate change event in which global temperatures increased between 5–8 °C due to a rapid increase in atmospheric CO_2 (Kennett & Stott, 1991; McInerney & Wing, 2011; Zachos et al., 2001; Zachos et al., 2003; Zachos et al., 2006). The vertical character of the associated geochemical proxy signal from preserved strata provides an important baseline for how Earth's surface might respond to, and recover from, current global warming. Trampush and Hajek convolved a generic PETM-like proxy signal with 1-D synthetic cores produced with a stochastic sedimentation model that mimics variation in sedimentation of fluvial or shallow marine environments. The apparent duration and magnitude of the PETM-like event preserved in the synthetic sections produced by Trampush and Hajek differed from the input signal. Model runs with high stochastic variability relative

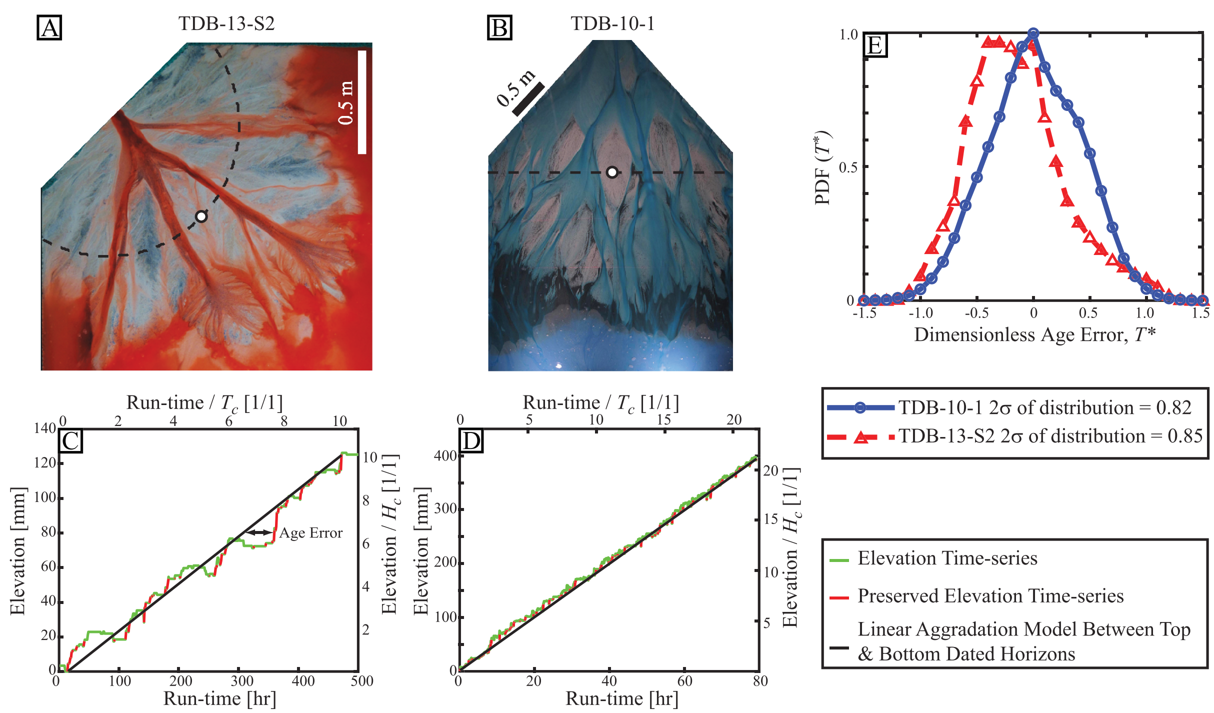

Figure 12. Measurements of deposit age and linear age model estimated ages from two experiments. (a, b) Images of TDB-13-S2 (Li & Straub, 2017) and TDB-10-1 (Wang & Straub, 2017) experimental surfaces with flow field dyed for visualization. Both experiments evolved in response to constant supplies of water and sediment and constant rates of sea level rise. However, the magnitudes of these forcings varied in the two experiments by over an order of magnitude, while other properties such as the ratio of water to sediment supply and sediment cohesion varied between experiments. As such, the rates and styles of surface processes varied. Dashed lines indicate location of transects extracted for analysis. Open circles show location of 1-D time series of surface elevation and stratigraphically preserved elevations in the two experiments shown in panels c and d, as well as an age model for experimental stratigraphy constructed with knowledge of the basal and topmost deposit age and assuming linear aggradation between. (e) Distributions defining dimensionless age error between measured and estimated ages in two experiments from all 1-D transects that comprise cross sections extracted for analysis. Dimensionless age error is generated by normalizing measured age error by compensation timescale.

10.1029/2019JF005079

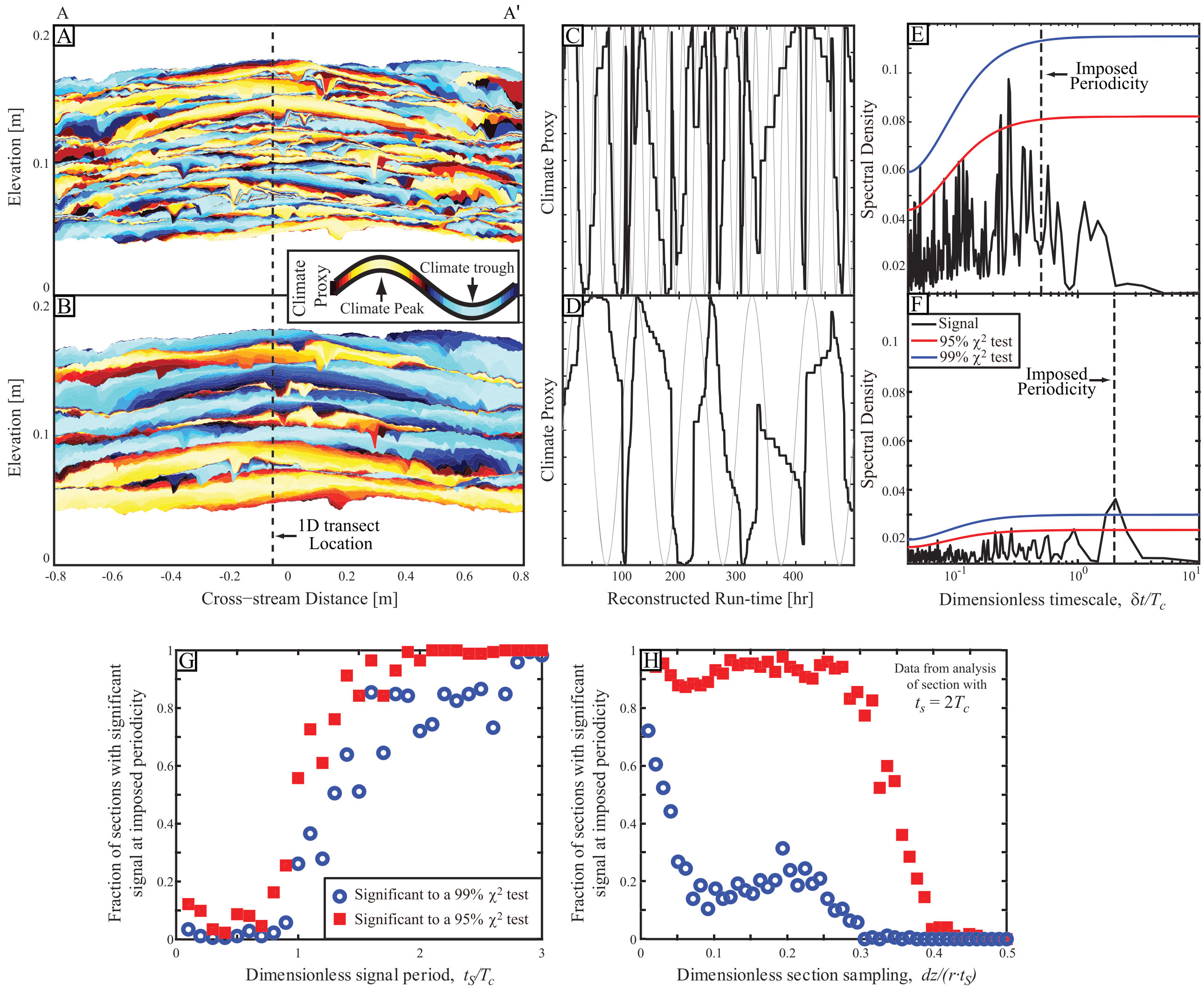

Figure 13. Results illustrating stratigraphic storage thresholds of environmental proxies in stratigraphy. Analysis utilizes synthetic stratigraphic panel shown in Figure 8c. Each 1-D topographic sample location along strike section samples an environmental signal, which for simplicity follows a sinusoidal cycle. The amplitude of this cycle is held constant, but the period is altered systematically. Stasis, erosional, and depositional events of varying duration and magnitude convert continuous environmental signal into a proxy record discretized into stratigraphic beds. Strike sections of stratigraphy colored by magnitude of proxy for environmental signals with periods equal to $0.5T_c$ (a) and $2T_c$ (b). Movie documenting evolution of strata in time and sampling of proxy can be found online (https://www.youtube.com/watch?v=-l94NTIZT8A). The 1-D preserved climate proxy records (c, d) show mismatch between reconstructed time series of proxy (solid black line) and imposed environmental signal (light gray line). Reconstructed time is generated with age of basal and top most deposits and linear age model between. Power spectra of reconstructed proxy time series and χ^2 confidence bands show storage of $2T_c$ signal (e) but no significant signal in $0.5T_c$ stratigraphy (f). Analysis of environmental signals with a range of periodicities (g) demonstrate reliable storage of environmental signals in proxy records when period of signal is greater than or equal to $2T_c$. For strata that reliably house environmental signals in proxies, vertical sample spacing must be less than a dimensionless sample spacing of 0.25 to recover signal (H). Sample spacing is made dimensionless by dividing by the product of the long-term aggradation rate and the period of the signal to be recovered.

to the long-term sedimentation rate resulted in the most altered records, including many runs that failed to record any evidence of the imposed >100 kyr proxy excursion. Although this result is sobering, the model results also demonstrate that, when averaged together—even crudely with very simple age models—an aggregate synthetic section can accurately reflect the input climate signal. This underscores how local variation in sedimentation rate (due to stochastic forcing and the autogenic reorganization of a SRS) can lead to significant incompleteness in any given section or core. To improve uncertainty or overcome this impediment, aggregate or spatially averaged data sets must be considered [see also Burgess et al., 2019].

Significant signal warping, due to unsteady deposition rates, is not just a problem in high-energy terrestrial settings as once thought. Marine mudrocks are often assumed to have high stratigraphic

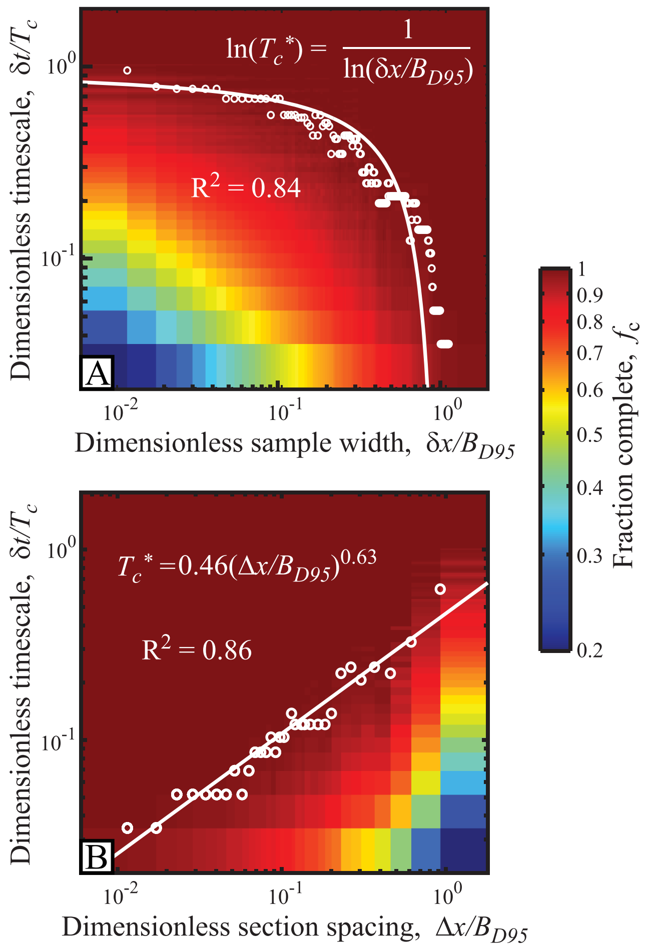

Figure 14. Results illustrating how stratigraphic completeness is a function of both timescale of discretization and lateral field of view. (a) Phase space defining how stratigraphic completeness changes as a function of dimensionless timescale of discretization and dimensionless sample width. Open circles reflect timescale necessary to reach 100% completeness for a given width of observation. Solid white line is theoretical trend which is defined by equation (9), while R^2 value characterizes fit of equation (9) to the data. (b) Phase space defining how stratigraphic completeness changes as a function of dimensionless timescale of discretization and dimensionless 1-D section spacing. Open circles reflect timescale necessary to reach 100% completeness for a given section density. Solid white line is empirical fit to data, which is defined by equation (10), while R^2 value characterizes quality of regression. Both phase spaces are generated with data set shown in Figure 8c.

completeness resulting from a continuous "rain" of sediment from above in low-energy settings. However, many recent observational and modeling studies are challenging this assumption by suggesting that advective and high-energy processes, like turbidity currents, can dominate the construction of thick muddy sections in the deep marine environments (Boulesteix et al., 2019; Kemp et al., 2018; Trabucho-Alexandre, 2015).

Motivated by this recent work, Foreman and Straub (2017) set out to quantify the signal duration necessary for confident extraction from proxy records. Rather than focusing on absolute timescales, the goal was to generate a flexible procedure for analysis of any environment where surface roughness features (e.g., channels) freely migrate over a region of long-term sediment accumulation. This was done by exploiting the topographic time series and the resulting synthetic stratigraphy from a physical laboratory experiment. Here we recreate the analysis of Foreman and Straub, using the same experimental data set used to demonstrate the "Sadler effect" and its link to stratigraphic (in)completeness (Figure 8). Deposition that occurs between two time steps samples, through a geochemical proxy, a climate parameter that follows a sinusoidal cycle with defined period (Figure 13). This method assumes no variability in how sediments are geochemically altered or measurement error in sampling the proxy from stratigraphy and is thus a best case scenario. Similar to the models of Trampush and Hajek (2017), the stratigraphic records are converted back from space to time by assuming a constant deposition rate, estimated from the section thickness and ages at the beginning and end of a section. Foreman and Straub proposed that the key autogenic timescale for faithful signal transfer is the compensation timescale, T_c. When the stratigraphy samples a climate signal with period or duration less than T_c, recovery from individual sections is fraught with error and spurious signals. Not until the climate signal period or duration is greater than $2T_c$ is the true signal faithfully and consistently recovered from strata (Figure 13g). We suggest that this timescale also influences the minimum sample spacing necessary for identification of environmental signals in proxy records (Figure 13h). We find that a dimensionless section spacing given by $dz/(rt_s)$ must be less than 0.25 for signal recovery in our data, where dz is the vertical distance between sample points and t_s is the period of an environmental signal.

3.2.5. Stratigraphic (In)completeness in Higher Dimensions

Given the challenges and imperfect transfer of environmental signals to stratigraphic storage in 1-D sections, new efforts are being directed toward quantifying the completeness of strata in higher dimensions (e.g., Mahon et al., 2015). This effort again started with the indirect approach of quantifying sediment accumulation rates as a function of measurement duration but now focusing on 2-D accumulation rates. Sadler and Jerolmack (2015) constructed a new database of literature-reported progradation rates (Figure 7). Multiplication of previously reported aggradation rates by these progradation rates removes the accumulation rate dependence on the timescale of measurement (Figure 7). This suggests that if a complete source-to-sink dip section can be constructed, sediment should be deposited somewhere along the transect for all moments in time. Sadler and Jerolmack attributed this to mass conservation, inferring that erosion at proximal sites likely leads to progradation of a landform at more distal sites. The globally averaged nature of the Sadler and Jerolmack (2015) data set, while encouraging for signal detection, has several caveats. These include a requirement of complete coverage of all depositional sinks in the dip transect and that measurement duration is long enough such that transport in the third (lateral) dimension can be ignored.

Partially to explore these caveats, Straub and Foreman (2018) quantified the influence of lateral field of view on stratigraphic completeness using the previously discussed control experiment (Figure 8). This analysis started by measuring f_c in 1-D sections with $\delta t/T_c$ set to the resolution at which data were collected. Straub and Foreman then systematically widened the field of view in steps equivalent to the lateral DEM grid spacing, δx, and asked if preserved deposition occurred in any grid cell in the field of observation: If yes, that time step is considered preserved. Similar to the normalization of δt by T_c, they sought a dimensionless sample width that could be constructed with an autogenic length scale. This was accomplished by dividing δx by B_{50}, the half-width of a sedimentary basin. B_{50} was found in the experimental data to approximate the widest swath of the geomorphic surface in stasis at any time in the experiment, which scales with the widest stratigraphic hiatuses. They found that f_c increases with dimensionless sample width and saturates at 100% when $\delta x/B_{50} \geq 1$. The f_c dependence on both the dimensionless sample width and the dimensionless timescale of discretization suggests a 2-D completeness phase space (Figure 14a). Straub and Foreman identified a formulation to estimate the minimum record discretization to achieve 100% f_c, T^*_C, as a function of the width of observation:

$$ln\left(T^*_C\right) = \frac{1}{ln(\delta x/B_{50})} \tag{10}$$

A similar analysis can be performed to characterize the density of 1-D sections across a basin necessary to achieve 100% stratigraphic completeness, for questions that are more feasible to tackle with collection of numerous cores or measured sections. We generate a 2-D phase space of f_c as a function of $\delta t/T_c$ and a dimensionless section spacing constructed as $\Delta x/B_{50}$, where Δx is the distance between 1-D sections (Figure 14b). In this phase space the timescale of discretization necessary for 100% completeness is empirically approximated as follows:

$$T^*_C = 0.46(\Delta x/B_{50})^{0.63} \tag{11}$$

Use of either equations (10) or (11) provides a path to recovery of signals that might be present in sedimentary basins but laterally dispersed due to the dynamics of the transport system. A similar analysis performed with dip-oriented sections might also yield a formal justification for use of the Sadler and Jerolmack (2015) framework for estimating sediment accumulation rates from measured aggradation and progradation rates.

3.3. Signal Shredding

In addition to smearing of signals over space, highlighted in section 3.2, environmental signals can be smeared in time due to the temporary sediment storage and later release through erosion of sediment in a variety of landforms, including dunes, bars, and floodplains (Pizzuto et al., 2017; Van De Wiel & Coulthard, 2010). As a result, abrupt perturbations to environmental forcings, which produce sediment-supply signals in hinterland regions, can be spread out over a wide band of time, in addition to space, prior to final and permanent stratigraphic storage.

Motivated by an analogy to turbulence and more specifically the modulation of input signals in fluid systems by turbulence, Jerolmack and Paola (2010) explored the capacity of autogenic processes (i.e., sediment storage and release) to alter, or in some cases shred, sediment flux signals during their propagation. Jerolmack and Paola define shredding (Table 1) as the smearing of an input signal over a range of space and timescales by stochastic processes such that an input signal is not detectable at the outlet of a system. This shredding was hypothesized to occur if the magnitude of the morphodynamic turbulence is strong relative to the magnitude and period of the signal. Jerolmack and Paola's definition represents an important distinction from signal loss due to stratigraphic incompleteness and is worth emphasizing: *A signal that is shredded is not recoverable from stratigraphy regardless of the ability to date deposits, how wide the field of view is, or how many 1-D sections are averaged.*

An analogy to describe the difference between a signal that is lost due to completeness versus shredding is as follows: Landscape dynamics that generate stratigraphic incompleteness is akin to cutting up and scattering pages of a journal article through an office; although it would take time to reassemble the pieces, the complete article could theoretically be recovered if we found enough remnants of the pages. In contrast, the effect of morphodynamic turbulence and resulting stratigraphic shredding is closer to that of burning pages of the journal article; in this case, the information contained in the pages (like a signal propagating through a SRS) is chaotically disassociated across space and time to the point where it can no longer be reconstructed. In this situation, even if the spatial-temporal distribution of sediment packages is completely known for a basin, the turbulent pathway of spatial/temporal smearing is irreproducible and cannot be known, so reconstruction is untenable (analogy of S. Toby, *pers. com.*). In the analogy from above, an example of landscape dynamics that generate incomplete but not shredded signals of environmental change would include the evolution of submarine fans by some turbidity currents, specifically, the evolution from flows that were spatially restricted, such that only part of the fan was active at any one time, and flows that were purely depositional (e.g., Burgess et al., 2019). An example of signal shredding would be a sediment flux signal exiting a catchment that decayed with transport length to the point where it could not be identified at the terminus of a SRS. The decay of the signal would be associated with temporary storage and then later redistribution of sediment in landforms such as dunes, bars, and floodplains (e.g., Jerolmack & Paola, 2010).

Jerolmack and Paola (2010) focused on transport of signals across Earth's surface and demonstrated with a suite of numerical models that the maximum scales of a system's autogenic processes likely control the degree of signal alteration by morphodynamics. They defined a timescale, T_x, that is expected to scale with the largest sediment flux perturbations, q', generated by autogenic processes:

$$T_x = \frac{L^2}{q_0} \tag{12}$$

where q_0 is the input sediment flux to a system and L is the system length. When input sediment flux cycles at periodicities greater than T_x, they pass through a transport system, but cycles with periodicities less than T_x are expected to be shredded prior to transfer to the record. Jerolmack and Paola also noted that signal magnitude is important. If the size of an input signal is greater than that of the maximum potential autogenic sediment release event, they hypothesized that it should propagate through a transport system, even when the input sediment flux cycles at periodicities less than T_x. They defined this maximum autogenic release event or autogenic magnitude threshold as follows:

$$M = L^2 S_c \tag{13}$$

where S_c is a critical slope for a transport system.

The Jerolmack and Paola (2010) theory provides a framework for assessing the shredding of signals across the Earth's surface but does not include signal loss due to the vertical cut and fill processes associated with sediment burial beneath the autogenic reworking depth. For example, an environmental signal might make it through the *surface process signal shredder* that controls temporary storage and intermittent transfer of sediment, but if net sediment-accumulation rates are small relative to landscape surface kinematics, the signal will ultimately be shredded by processes that rework previously deposited sediments. A suite of experiments performed in the Tulane University Delta Basin were conducted to extend the Jerolmack and Paola concepts to the transport of signals through the Earth's surface and the reworking zone to the stratigraphic record, thus developing a *stratigraphic signal shredder* framework.

Development of this stratigraphic framework began with storage of relative sea level (RSL) signals. As these signals interact with sediment transport at shorelines, examination of deltaic and marginal marine deposits allows one to isolate the stratigraphic shredder with minimal alteration due to signal propagation over the Earth's surface. Li et al. (2016) defined two dimensionless numbers that compare the upper spatial and temporal scales of deltaic autogenic processes to the magnitude and periodicity of RSL cycles:

$$H^* = \frac{R_{RSL}}{H_C} \tag{14}$$

$$T^* = \frac{T_{RSL}}{T_C} \tag{15}$$

where R_{RSL} is the difference in sea level elevation from cycle peak to trough and T_{RSL} is the period of a cycle. RSL cycles with either H^* or T^* values in excess of 1 were successfully stored in the stratigraphy of deltaic

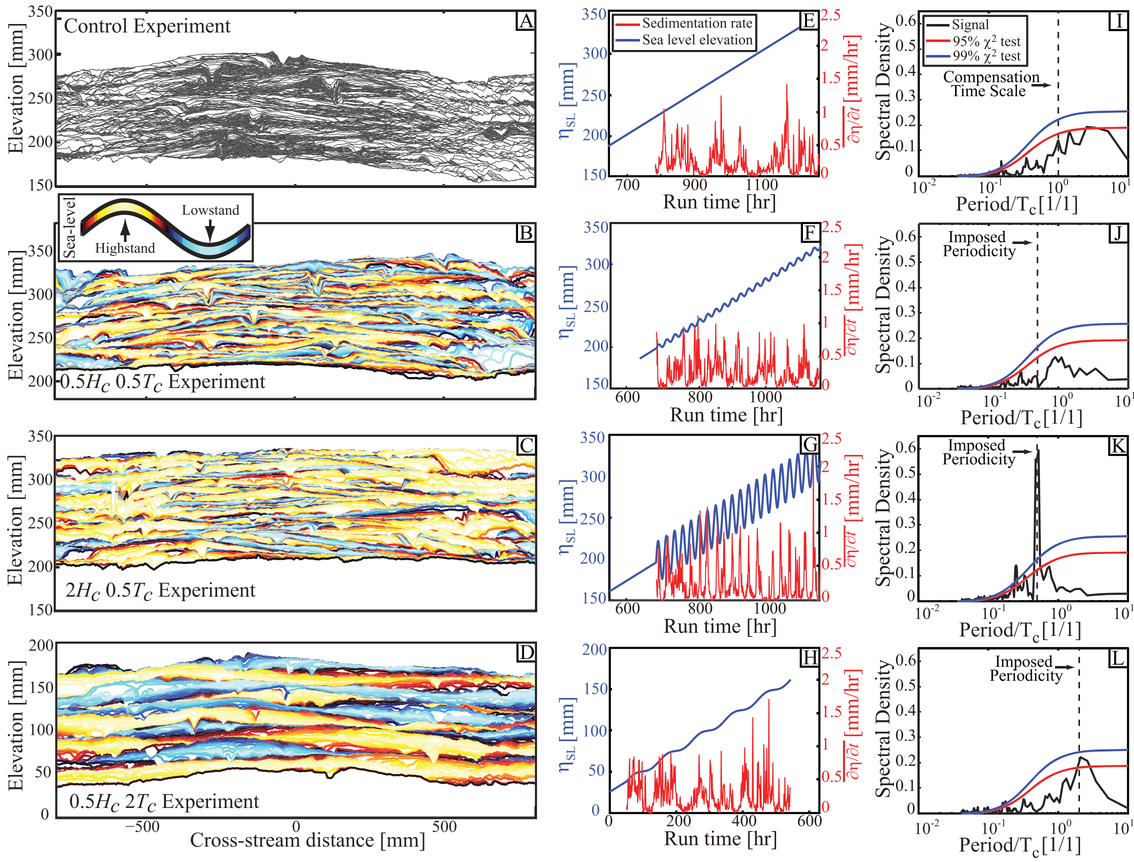

Figure 15. Results illustrating stratigraphic storage/shredding thresholds for relative sea level (RSL) cycles from Li et al. (2016). Analysis centers on time series of mean deposition rates measured from preserved experimental stratigraphy along a strike transect located approximately halfway between the basin entrance and mean shoreline. Data from four experiments are presented that share identical forcing conditions, with the exception of the period and magnitude of RSL cycles. The experiments include (a) a control experiment with no RSL cycles, (b) an experiment with cycles defined by a range (R_{RSL}) that is half of the largest channel H_c and a period (T_{RSL}) that is half of the compensation timescale T_c, (c) an experiment with cycles defined by a R_{RSL} that is twice H_c and a T_{RSL} that is half T_c, and (d) an experiment with cycles defined by a R_{RSL} that is half H_c and a T_{RSL} that is twice T_c. (a–d) Synthetic stratigraphy colored by time of deposition relative to location in RSL cycle. (e–h) Sea level (η_{SL}) and mean deposition rate ($\delta\eta/\delta t$) time series. (i–l) Power spectra of mean deposition rate time series and χ^2 confidence limits. Figure modified from Hajek and Straub (2017).

laboratory experiments, but no evidence could be found for signal storage of cycles with H^* and T^* values less than l, supporting their theory (Figure 15). These experiments had nearly identical forcing conditions as our previously discussed autogenic experiment (Figure 8), with the exception of the RSL cycles. Parameters that carried signals of RSL when H^* and/or $T^* > 1$, but lacked signals when H^* and $T^* < 1$, included accumulation rates calculated from synthetic stratigraphy (both in 1-D and 3-D), the variability in these rates, the completeness of the final record, the vertical and lateral extent of parasequences, the dimensions of channel bodies and paleovalleys, and distribution of sandy deposits (Li et al., 2016; Yu et al., 2017). Results from a linked shelf edge slope experiment suggest that this holds not just for terrestrial signals but also continental slope deposits (Straub, 2019).

An important attribute of the Li et al. (2016) framework is the use of variables that can plausibly be measured from or estimated for field-scale systems. For example, we can use the reported values for H_c and T_c in Figure 11 to test the susceptibility of these deltaic systems to Quaternary-scale eccentricity-driven (R_{RSL} ~ 100 m, T_{RSL} ~ 100 kyr) and Late Miocene-scale obliquity-driven (R_{RSL} ~ 15–30 m, T_{RSL} ~ 40 kyr) sea level cycles (Figure 16). Li et al. suggested that storage of Quaternary-scale RSL cycles is likely in all systems examined, primarily due to their large R_{RSL}. While most small- to medium-scale systems are also likely to store signals of Late Miocene-scale cycles, large deltaic systems like the Mississippi and Ganges might not due to their large autogenic scales.

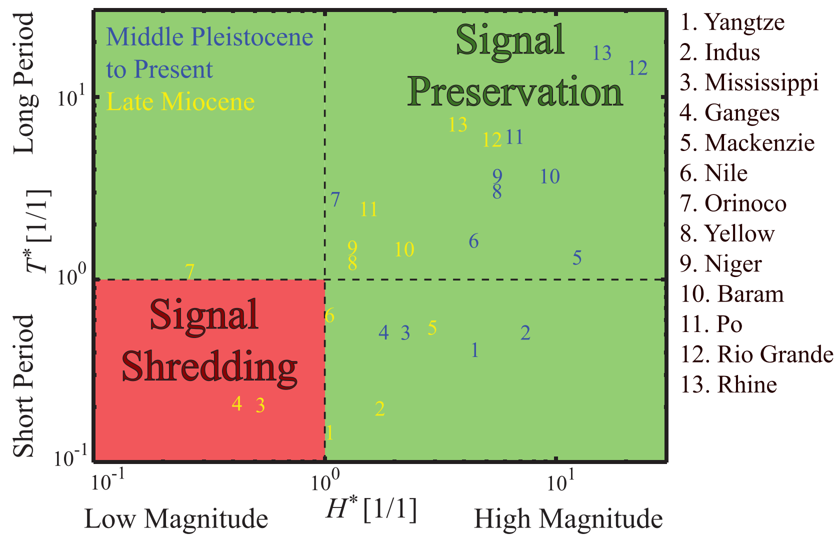

Figure 16. Predictions of magnitude and period of RSL cycles and their stratigraphic storage potential when normalized by autogenic length and timescales for 13 major river systems. Data originally published by Li et al. (2016). Predictions are for two time periods, the middle Pleistocene to the present (blue) and the late Miocene (yellow) and are based on modern channel depths and long-term (time window of measurement >100 kyr) sedimentation rates.

Many of the systems in Li et al.'s compilation lie close to the predicted storage thresholds for Late Miocene-scale RSL cycles, suggesting that the stratigraphic records of these deltas can preserve signals of periodic RSL change, but they may be difficult to identify given practical limits to collection of field data. Two typical scales extracted from field data sets to explore paleo-sea level history are the planform length of parasequences and the relief of erosional surfaces. However, recent work highlights how autogenic parasequences and scour scales might be larger than previously thought and linked to the backwater dynamics of coastal rivers. Straub et al. (2015) observed in physical experiments that the maximum proximal to distal length of parasequences scales with a system's backwater length, L_B. This length approximates the distance upstream of the shoreline where channels start to lose sediment transport capacity as their water surface slopes approach zero to match those of the receiving basin (Chow, 1959; Paola & Mohrig, 1996) and scales as follows:

$$L_B \approx \frac{H_N}{S_N} \tag{16}$$

where H_N and S_N are the normal flow depth and slope, respectively. The reduction in transport capacity at the upstream extent of the backwater zone drives channel aggradation and nodal avulsion sites (Chatanantavet et al., 2012; Jerolmack, 2009). In the Tulane experiments the scales of parasequences generated in the presence of RSL cycles, with characteristics that would place them in the shredding domain, could not be differentiated from the scales of parasequences generated by a purely autogenic experiment. In the Mississippi River Delta this length scale is on the order 500 km, suggesting the possibility of large autogenic length scales for parasequences in some systems.

Whereas backwater hydrodynamics drive channel aggradation, a spatial acceleration of river flow as it approaches a shoreline during seasonal floods can cause channel bed erosion (Lamb et al., 2012; Nittrouer et al., 2012). New theory and observations suggest that flood-induced spatial flow acceleration can produce erosional surfaces with relief that scales to the flood regime and observationally equal to 0.5–3.0 times the mean channel depth (Ganti et al., 2019; Trower et al., 2018). This suggests a similar storage threshold as the H^* threshold discussed above since the H^* formulation uses a maximum channel depth and that flood-induced scours could exceed the range of many RSL cycles in medium to large deltaic channels, making them difficult to differentiate in outcrop. Trower et al. even noted that this challenges our ability to identify

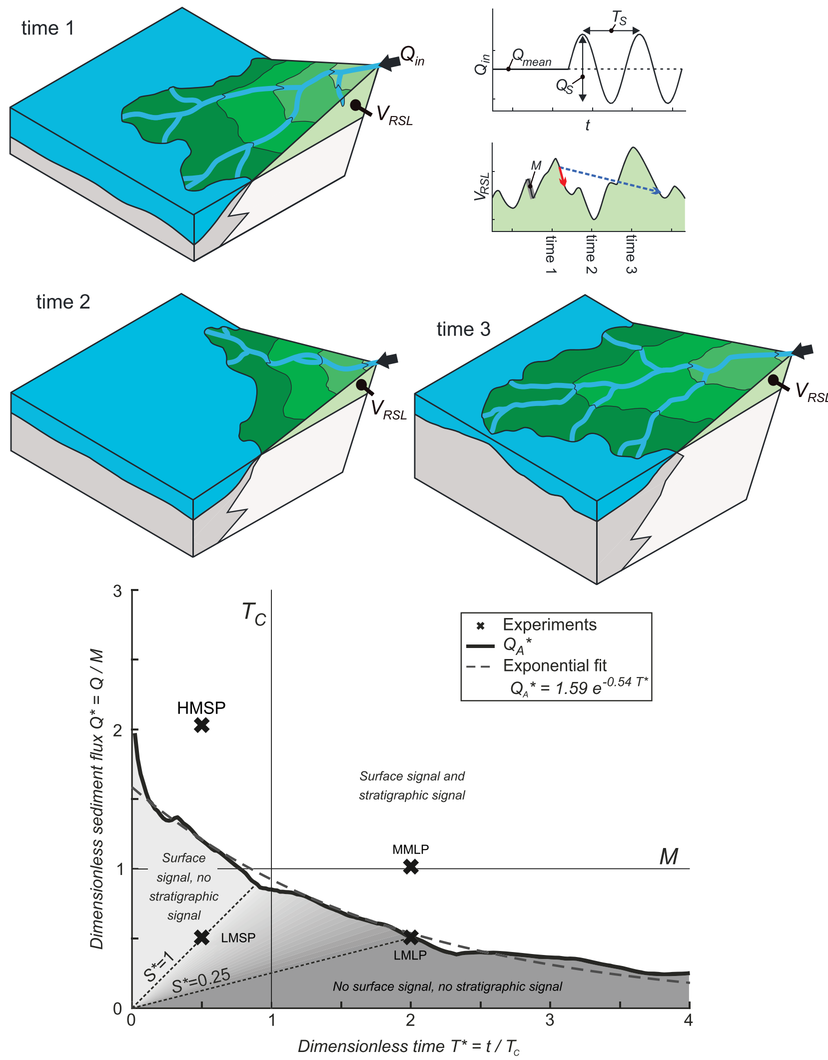

Figure 17. Processes responsible for stratigraphic shredding of sediment flux signals and their quantification. Set of schematic diagrams illustrate an aggradational delta at three different time steps. Even under constant sediment supply rate (Q_{in}), the volume stored above sea level (V_{RSL}) continually adjusts in response to autogenic sediment storage bypass and release (SSBR). Schematic graphs show a cyclic sediment supply signal and expected autogenic changes in V_{RSL}. The rate of V_{RSL} change generally decreases with the time window measured over (e.g., red vs. dashed blue arrow) and approaches zero over long timescales. The magnitude and period of these autogenic SSBR events set a signal storage threshold which is dependent on the timescale of a signal. Plot at base defines the stratigraphic storage threshold of sediment flux signals for an experimental system and is based off measurement of autogenic change in a delta that evolved in the absence of sediment flux signals. Crosses mark combinations of periodicity and peak-to-peak amplitude for cyclic experimental stages. Analysis of stratigraphy supports the signal threshold theory, with experiments plotting above the threshold storing signals in their stratigraphy and experiments that plotted below the threshold lacking any stratigraphic signals of periodic supply signals. Modified from Toby et al. (2019).

RSL signals in outcrops from the Upper Cretaceous Castlegate Sandstone of the Book Cliffs, Utah, USA, where many sequence stratigraphy techniques to identify sea level cycle signals were developed (Van Wagoner, 1995; Van Wagoner et al., 1990).

Coming back to environmental information transmitted by the amount of sediment in flux, Toby et al. (2019) built on the Jerolmack and Paola (2010) theory by defining a threshold for signal storage of sediment supply cycles in stratigraphy. In contrast to the independent magnitude and period thresholds proposed in earlier shredding studies, they hypothesized a time-dependent magnitude threshold. This threshold is set by the maximum scale of autogenic storage, bypass, and release for a measurement window of interest. This was defined in their study by a change in the volume of terrestrial deposits as a function of measurement duration, which conveniently has units of a volumetric flux (Figure 17). A dimensionless version of this threshold can be constructed by normalizing the duration of interest by T_c and the magnitude of the threshold by a maximum rate of change in sediment volume stored in an environment of interest over a full period of sustained autogenic volume growth or loss, M. This threshold was supported by results from an identical set of experiments as those performed by Li et al. (2016), except here it was the rate of sediment supplied that was varied rather than RSL.

A noted parallel exists, between the signal buffering mechanism proposed by Metivier and Gaudemer (1999) and the experimental signal shredding work of Toby et al. (2019). Metivier and Gaudemer proposed that a deterministic exchange of sediment between channels and their floodplains could buffer the influence of a change in Q_s supplied to a basin when measured at the outlet of a river. While described in a deterministic framework, this exchange must occur through stochastic processes like channel migrations and relocation through avulsions. These processes occur in sedimentary basins even with constant forcings but might accelerate or decelerate depending on the change in flux of sediment provided to a basin (Powell et al., 2012; Wickert et al., 2013). The work of Toby et al. suggests that stochastic sediment storage, bypass, and release alter the volume of sediment stored in overbank environments. If a change in sediment flux to a basin is not of sufficient magnitude or duration, then the stochastic sediment exchange between channels and their overbanks buffer the change to such an extent so as to render it unidentifiable at the terminus of a river or in the strata produced downstream of the river terminus. However, Q_s changes that exceed the shredding thresholds will still be buffered to some degree by the autogenic processes unless the period or duration of these changes exceeds T_{eq}.

An initial exploration of signal storage of climatic and tectonically produced sediment supply signals suggests that many, but certainly not all (Blum et al., 2018), commonly discussed signals (i.e., Milankovitch climate or punctuated uplift signals) are either prone to shredding or likely fall very close to the proposed threshold, making extraction of signals with common field exposure and methods challenging (Toby et al., 2019).

While the framework of Toby et al. (2019) suggests that a Milankovitch-scale change to mean supply of sediment is prone to shredding, stratigraphic signals might still result from changes in discharge variability. For example, a recent study highlighted the difference in stratigraphic architecture of two experimental deltas constructed with the same long-term water and sediment supply rates but different flood hydrographs (Esposito et al., 2018). The system constructed with larger floods had lower preservation of floodplain strata and was enriched in channel deposits, similar to results from long-standing stratigraphic architecture models (Allen, 1978; Bridge & Leeder, 1979; Leeder, 1978). Changes in discharge variability have been linked to changing channel patterns and stratigraphic products, specifically in the Mississippi River system, which experienced changes in channel pattern and avulsion rates due to large glacial outburst floods in the late Quaternary (Bentley et al., 2016; Knox, 1996). Even small changes in discharge variability, though, can influence autogenic processes. For example, many river avulsions occur during a trigger event that follows a long-term morphological avulsion setup (Mohrig et al., 2000; Slingerland & Smith, 1998). Floods are common trigger events, and thus, changing their magnitude-frequency distribution will influence the style and frequency of channel relocation events (Ganti et al., 2019). This highlights a need for an expansion of the theory of Toby et al. (2019) to describe shredding thresholds that might predict the magnitude and duration of a change in flood environment necessary to produce stratigraphic signals.

Finally, we note that the stratigraphic signal-shredding frameworks developed for RSL and sediment supply cycles only work for depocenters that are long-term sinks of sediment and thus associated with positive T_c values. These frameworks cannot yet be used to assess the fidelity of river terrace deposits, bounding channels that are in the long-term incisional (e.g., Bridgland & Westaway, 2008; Hancock & Anderson, 2002), to store signals of environmental variability. Development of theory to predict fidelity of records in net incisional settings is needed and will likely also have to take into account autogenic activity that occurs in net incisional corridors (Finnegan & Dietrich, 2011; Tofelde et al., 2019).

4. Estimating Key Scales for Signal Preservation in Field Systems

Although the presence of key length scales and timescales in SRSs is clearly important for the preservation of environmental signals in the stratigraphic records of physical and numerical experiments, a major challenge moving forward for the community is integrating and testing these in the field. This will allow the community to develop null hypotheses and quantify the confidence in reconstructing past eustatic, climatic, and tectonic events. While much of the focus in this section is on theoretical and empirical methods for paleo-scale reconstructions from natural systems, we note that bounds on some of these scales can be inferred simply by knowing the general tectonic context in which strata were deposited. For instance, bounds on the length of a SRS or the relief and spacing of catchments can be made with knowledge of the basin type, that is, rift versus foreland versus passive margin (Colombera et al., 2017; Hovius, 1996; Talling et al., 1997; Whipple & Trayler, 1996). For some problems, these first-order scale bounds alone could be useful in estimating the signal storage potential of a basin.

4.1. T_{eq} and T_c Equivalencies

Our synthesis of impediments to environmental signal storage in stratigraphy highlighted two emergent timescales in landscapes: the equilibrium timescale (T_{eq}) and the compensation timescale (T_c). These timescales both describe the amount of time necessary for systems to respond to a change in forcing conditions such that either the landscape structure returns to steady state at timescales $>>T_{eq}$ or the structure of strata is fully set by regional allogenic forcings at timescales $>>T_c$. T_{eq} fundamentally describes the intrinsic timescale of regrading the land surface, and T_c describes the intrinsic timescale of constructing strata whose depositional geometry is set by forcing conditions in a basin. Given that both the downstream and lateral transport of sediment across basins are mediated by the morphodynamics of depositional transport systems, it holds that T_{eq} and T_c should be linked. This link appears in the physical experiment used throughout the section on stratigraphic completeness (Figure 8). We estimated T_{eq} and T_c using equations (6) and (8) with values from known forcing conditions and experimental measurements. We note the independence of these calculations as no single parameter can be found in both equations. We find that T_{eq} and T_c are equal within a factor of 2 ($T_{eq} = 1.8T_c$).

The link between T_{eq} and T_c can also be gleaned from the T_{eq} formulation put forward by Paola et al. (1992) and further explored by Paola et al. [1999]. Their description of landscape diffusion was constructed for the dynamics of sedimentation at basin filling length and time scales. This resulted in a set of equations that allow diffusional modeling of landscapes through a 1-D approach by averaging lateral variability in basin dynamics; such lateral averaging accounts for the timescales necessary for the products of stochastic dynamics (i.e., topographic roughness across a basin) to average out in the resulting strata.

Estimation of T_{eq} in depositional basins is commonly achieved through use of a transport coefficient, or diffusivity parameter, (ν) estimated from mean transport conditions in a basin (Métivier & Gaudemer, 1999; Paola et al., 1992). This simplifying assumption has been parameterized in different ways for different studies; for example, estimates of ν typically rely on knowledge of either a characteristic sediment flux, Q_s, (equation (4)) (Métivier & Gaudemer, 1999) or water flux, Q_w (Jerolmack & Paola, 2007; Paola, 2000). While reasonable estimates of diffusivity can be obtained in extant systems, where, for example, Q_s or Q_w can be approximated with reasonable assumptions (e.g., BQART model of Syvitski and Milliman [2007]), identifying an appropriate value of ν for deep-time deposits necessitates a multitude of assumptions, each with high uncertainties and limited opportunities for validation.

A potentially powerful approach to reduce uncertainty associated with T_{eq} estimates may be to more comprehensively integrate T_{eq} reconstructions with efforts to determine exhumation rates from the thermochronologic ages of detrital minerals within the sedimentary record. Several studies compare the cooling age of various minerals (e.g., apatite, zircons) with the depositional age of the strata that host the minerals [*Bernet et al.*, 2006; *Carrapa*, 2009; *Painter et al.*, 2014]. The difference between the two ages represents a "lag time" that includes both the rate of catchment exhumation (when the geothermal gradient in the catchment can be estimated) and the time it takes for the sediment to be transported to a given location in the sedimentary basin (Garver et al., 1999). Comparing lag times of sequential sedimentary strata allows researchers to determine if the catchment is in steady state or if exhumation is increasing/decreasing through the history of the coupled catchment basin system (Garver et al., 1999; Carrapa, 2009; Whitchurch et al., 2011). In current approaches the transmission time across the basin is underconstrained but should be directly related to T_{eq}. Estimating T_{eq} should provide a greater resolution on the rates of exhumation, which in turn allow sedimentologists an independent constraint on an important boundary condition (tectonic variation).

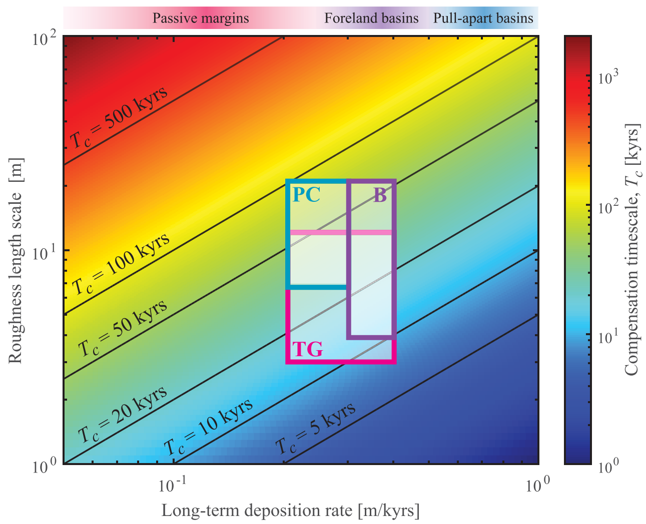

Figure 18. Contoured T_c values for reasonable ranges of landscape roughness and long-term sedimentation rates. Larger roughness length scales and slower sedimentation rates yield longer T_c values. Estimated T_c ranges for the Bighorn Basin, Wyoming, USA (B), Piceance Creek Basin, Colorado, USA (PC), and Tremp-Graus Basin, Spain (TG) (Table 2) indicated in transparent boxes. Approximate characteristic sedimentation rate ranges for different basin types shown above plot [after subsidence rates of Xie & Heller, 2009]. Landscape roughness length scales range from relatively small rivers that might fill, for example, proximal basins (e.g., 1 m deep flows) to large relief scales (100 m) that would be associated with very large terrestrial rivers, large aggradational deepwater channels, or potentially large-scale relief associated with fluvial or alluvial fans.

In contrast, T_c is directly measurable from stratigraphic deposits and is relatively easy to constrain with first-order assumptions, relying on estimates of the maximum morphodynamic vertical roughness scale across a basin and long-term sedimentation rates. Future work may help reduce uncertainties associated with estimating T_{eq} directly; however, at present, using T_c to constrain key scales associated with buffering, (in)completeness, and shredding provides a practical way to evaluate the potential for signal preservation in the sedimentary record.

4.2. Field Methods for Estimating T_c

Central to estimating T_c for field systems is characterizing the largest vertical topographic roughness scale, l, in a depocenter. Many of the studies that characterized T_c in laboratory or field data sets utilized the largest autogenically generated channel depths as estimates of l. However, we note that the maximum roughness scale that is important for estimating T_c is a measure of the largest roughness scale that can occur across a basin. The imprint of T_c can be directly measured from stratigraphy (e.g., Pisel et al., 2018; Straub & Pyles, 2012; Trampush et al., 2017; Wang et al., 2011); such studies have shown that landscape roughness associated with compensation can significantly exceed maximum channel depths in fluvial and deltaic systems and in nonchannelized settings might equate to maximum mounding from lobe deposition. In channelized systems, paleoflow-depth measurements (e.g., from scour surfaces or preserved bar-clinoform heights) are a sensible starting point for estimating l and reconstructing a minimum estimate of T_c. To explore upper-limit estimates of T_c, however, maximum estimates of channel belt sand body thickness or even the thickness of stacked channel belt clusters should be used to account for the potential influence of mounding from alluvial ridge or fluvial fan deposition (Hajek & Straub, 2017). Characterizing T_c also necessitates an estimate of long-term aggradation rates, r. Quantifying this rate relies on available geochronometers. Calculations of T_c should be based off a long-term rate and thus more straightforwardly measured in deeper geologic time than short-term rates. Here we define a long-term rate as one measured at least over a time span necessary to aggrade everywhere in a basin by, on average, one l. This corresponds to the timescale where deposition rates switch from being transient and dependent on the timescale of measurement to persistent and independent of measurement duration, set by the generation of accommodation. Typically, this timescale will be on the order of 100 kyr (Jerolmack & Sadler, 2007). In the absence of available geochronometers, simple estimates of sedimentation rates appropriate to a particular basin type (e.g., Xie & Heller, 2009) place useful constraints on the value of r.

5. Leveraging Sedimentary Scales to Estimate Signal Preservation in the Field

By understanding the key scales at which signal buffering, (in)completeness, and shredding occur in different landscapes, it is possible to identify an appropriate depositional record, physical scale of inquiry, or sampling strategy necessary to answer a particular science question. Estimating the scales at which intrinsic physical sediment transport dynamics might dominate sedimentary patterns helps establish a null hypothesis about whether stratigraphic observations in a particular record may reflect landscape morphodynamics or external signals of tectonic, climate, or sea level change.

To a first order, the key timescales (T_c and T_{eq}) can be inferred from basic assumptions about climatic and tectonic conditions in a basin. Figure 18 shows contour plots of T_c values for reasonable ranges of input values characteristic of sedimentary basins on Earth. In some systems, input values might be measurable, and in systems where one or both of these variables is unconstrained, a range of estimates can provide useful insight into potential uncertainties about key scales. For T_c, the vertical relief parameter (l) can be estimated directly from field measurements, and, in cases where measurements are unavailable, a range of reasonable l values could be assumed based on analogous depositional systems in similar geological settings. Similarly, long-term sedimentation rates can be estimated in specific basins with geochronology, biozones, or astrochronology, but reasonable uncertainties on long-term sedimentation rates could be assumed for different tectonic settings (e.g., Xie & Heller, 2009).

Applying and testing these theoretical concepts to field cases is an actively evolving area of research (e.g., Covault et al., 2010; Trampush et al., 2017; Watkins et al., 2018). Ideally first-order field estimates of T_c (or potentially T_{eq}, in the future) will become routine metrics for determining whether a particular record is likely to bear a signal of paleoenvironmental change. For example, such efforts would be particularly important for reconstructing the frequency-magnitude distribution of earthquake or flood events from the stratigraphic record, where T_c and T_{eq} estimates would provide insight into the likelihood that a particular deposit faithfully records events of a given size and recurrence interval. In another example application, because T_c and T_{eq} determine the fundamental length scales over which stratigraphic correlation can be confidently accomplished, these scales can provide unique insight into subsurface prediction and correlation. From a sequence-stratigraphic perspective, we should only expect regional, mass balance correlations to hold for scales above T_c and T_{eq}; below this, the depositional record should be dominated by local, stochastic variability, and hence sediment packages that are limited in extent. This insight can be useful for correlating between isolated outcrops or well logs, for predicting the subsurface distribution and connectivity of geofluid flow units, and for populating reservoir models.

Important avenues for future progress include estimating T_c and T_{eq} for a range of systems with previously identified climatic, tectonic, and sea level signals and evaluating the effects of buffering, (in)completeness, and shredding. Furthermore, comparison studies of basins with different characteristic T_c and T_{eq} values and how they responded to global change provide opportunities to test the theory put forth by experiential results. To highlight this potential, we present interpretations of landscape response to the PETM as a well-developed example of how these connections can be made.

5.1. Impediments on the Stratigraphic Record of the PETM

The quality of paleoclimate time series extracted from sedimentary deposits will be impacted by (in)completeness, buffering, and geomorphic shredding processes. For climate changes that far exceed system equilibrium response times, these impediments are expected to have a limited effect on the quality of a record; however, climate events that occur over mesotimescales (10^3–10^5 yr) may directly overlap with the timescales of impediments in a given depositional setting.

The PETM was a large-magnitude, mesotimescale (~1.8×10^5 yr long) global climate change event that occurred ~56 Ma and involved increases in global temperatures between 5 and 8 °C (Kennett & Stott, 1991; McInerney & Wing, 2011; Zachos et al., 2001; Zachos et al., 2003; Zachos et al., 2006). PETM warming is associated with a significant, negative stable carbon isotope excursion in a variety of organic and inorganic proxies that are linked to a massive release of exogenic carbon into Earth's atmosphere and oceans (McInerney & Wing, 2011). It has particular societal relevance because it provides a case example of a rapid onset (<10 kyr) CO_2-forced global warming event, which preserves not only the perturbed climatic state during carbon release, stabilization, and sequestration but also baseline climate states before and after the event. Many researchers hope to exploit this event to constrain Earth system models of climate response to a CO_2 forcing. Since its discovery, it has been evident that the PETM had severe consequences for marine and terrestrial systems (Crouch et al., 2003; Currano et al., 2008; Foreman et al., 2012; Gingerich, 2006; Koch et al., 1992; Koch et al., 1995; Kraus et al., 2015; McInerney & Wing, 2011; Sluijs et al., 2007; Thomas, 2003; Thomas & Monechi, 2007; Wing et al., 2005). Many of these paleo-observations are consistent with predictions and observations of modern day Earth systems in response to anthropogenic climate change (Carmichael et al., 2017; McInerney & Wing, 2011; Zeebe et al., 2016).

At present, there remain several outstanding questions that prohibit a comprehensive and robust comparison between Anthropocene climate change and the PETM. To answer these questions, the effects of signal buffering, (in)completeness, and signal shredding in SRSs must be considered. We will deal with each of these in turn, first establishing estimates for T_c in three basins containing sediments spanning the PETM: the alluvial Piceance Creek and Bighorn basins of North America (Colorado and Wyoming, respectively) and the Tremp-Graus Basin (northern Spain). Using previously measured long-term average sedimentation rates and estimates of maximum relief on paleolandscapes, from maximum river flow depths and maximum

Table 2
Estimates of Compensation Timescales for PETM Basins With Key Scales Necessary for Calculation

		l (m)	References	r (cm/kyr)	References	T_c (kyr)
Bighorn Basin	Channel belt sandbody thickness	20	Foreman (2014)	30	Clyde et al. (2007)	67
	Bar clinoform relief	4		40		10
Piceance Creek Basin	Channel belt sandbody thickness	20	Foreman et al. (2012)	20	Johnson (1992), Foreman & Rasmussen (2017)	100
	Bar clinoform relief	6.5		30		22
Tremp-Graus Basin	Channel belt sandbody thickness	11	Chen et al. (2019) Columbera et al. (2017)	20	Duller et al. (2019)	55
	Bar clinoform relief	3		40		8

fluvial sandbody thickness, we estimate T_c for the Piceance Creek Basin as 22–100 kyr, the Bighorn Basin as 10–67 kyr, and the Tremp-Graus Basin as 8–55 kyr, respectively (Table 2 and Figures 18 and 19) (Clyde et al., 2007; Colombera et al., 2017; Duller et al., 2019; Foreman et al., 2012; Foreman & Straub, 2017; Johnson, 1992; Schmitz & Pujalte, 2007). These estimates suggest that, as a whole, the PETM can be considered a "long-term" climatic event in these basins, but shorter-duration segments of the PETM (e.g., the onset and peak excursion time periods) overlap differently with timescales that might be affected by buffering, (in)completeness, and shredding. As such, we would predict different quality records and environmental signals in each basin.

One of the outstanding questions surrounding the PETM is the character of geomorphic response in the non-marine basins listed above. The stratigraphic response is represented by anomalously thick and laterally extensive fluvial sand bodies associated with floodplain strata composed of better drained paleosols, related to seasonal drying (Foreman, 2014; Foreman et al., 2012; Kraus et al., 2015; Schmitz & Pujalte, 2007). A time lag of ~15 kyr has been found to exist between the onset of carbon isotope excursion and the onset of fluvial deposition and increased rates of siliciclastic input to deep marine systems (Duller et al., 2019). The cause of this time lag is related to buffering processes within the sediment-routing system that act to retard the propagation of the initial sediment flux signal (Duller et al., 2019).

A key difference among these three basins is the magnitude and duration of fluvial response to the PETM. In both the Tremp-Graus and Bighorn basins a single, anomalous fluvial unit is closely associated with the onset and early portion of the peak PETM excursion, whereas in the Piceance Creek Basin, there are multiple, thick, interconnected fluvial sand bodies that appear coincident with the earliest carbon isotopic excursion but persist after the PETM isotopic excursion ends (Foreman, 2014; Foreman et al., 2012; Schmitz & Pujalte, 2007). This difference may be related to differences in fundamental autogenic time scales.

The cross-basin persistence of fluvial response in each basin is consistent with the short T_c estimates for each basin relative to the duration of the PETM. The allogenic change was sufficiently long that the river system was able to visit much of the basin in its perturbed condition. The shorter T_{eq} of the Tremp-Graus and Bighorn basins would allow the PETM environmental signal to propagate more rapidly from source to sink and potentially regrade the landscape. Indeed, recent work suggests proximal and distal alluvial shifts in the two basins and minimal changes in fluvial gradients spanning the PETM in the Tremp-Graus Basin (Foreman, 2014; Kraus et al., 2015).

Importantly, the resolution of proxy records (e.g., geochemical and fossil proxies) is subject to the (in)completeness of the stratigraphic record that hosts them. In order to capture rates and magnitudes of past climatic changes accurately, proxies must be well sampled in the time domain, which is only possible when a "complete" stratigraphic section can be obtained. The consequences of stratigraphic incompleteness appear to be most detrimental when the timescale of climatic change is less than twice that of key autogenic timescales (Foreman & Straub, 2017). Thus, it is likely that a reasonably complete representation of PETM time is present in the Tremp-Graus and Bighorn basins, since the upper limit of T_c estimates is ~60 kyr less than half of the ~180 kyr duration of the PETM. T_c estimates for the Piceance Creek Basin extend into the ~100 kyr range, suggesting that the PETM timescale may be on the edge of preservation in this system. Furthermore, we

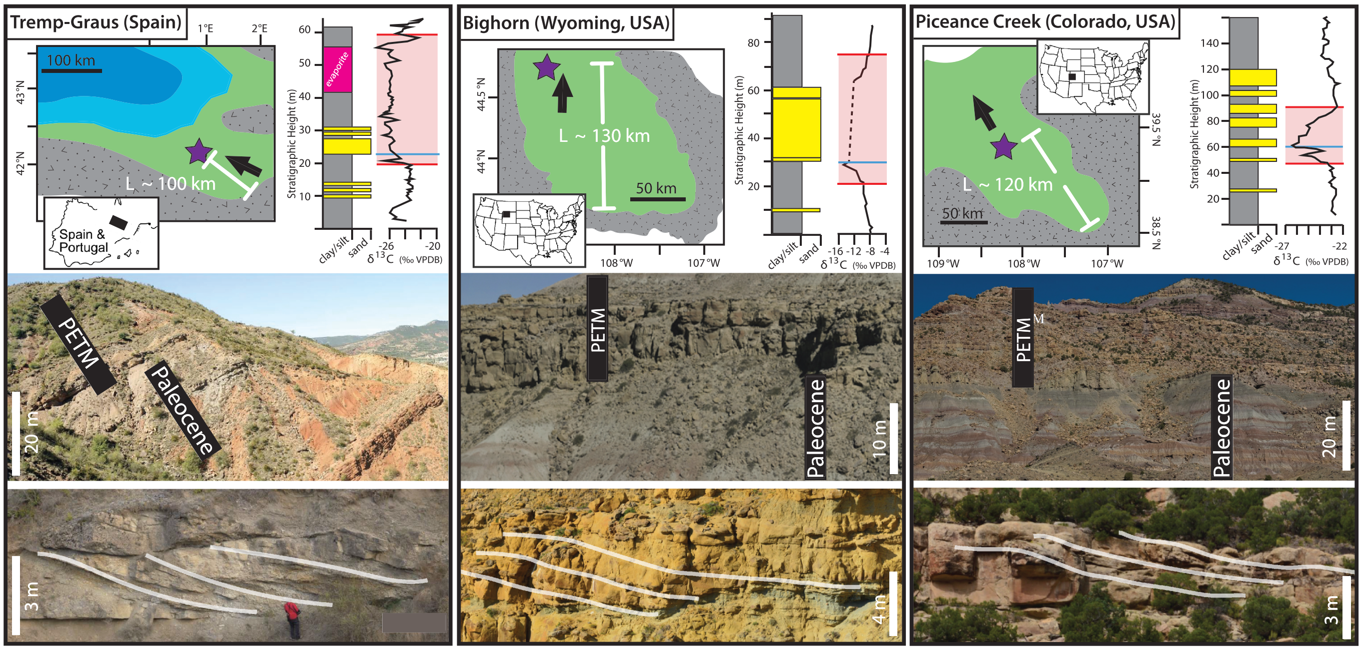

Figure 19. Example field localities of fluvial deposits spanning the Paleocene-Eocene Thermal Maximum (PETM) in the Tremp-Graus, Bighorn, and Piceance Creek basins. In each panel appears a basin map (top left), stratigraphic sections (top right), a field photo showing large-scale stratigraphic architecture (upper photo) and a close-up photo showing a channel belt sandstone body (lower photo). Basin maps: Context for constraining key scales associated with signal preservation, for example, depositional system length (L), can be estimated from basin dimensions, shown here as the distance from uplifted basement source to the study locality (star) along the path of paleoflow direction (black arrows). Maps and section data after Duller et al. (2019) (Tremp-Graus Basin, Claret Conglomerate section near Tremp, Spain), Foreman (2014) (Bighorn Basin, Willwood Formation near Powell, Wyoming), and Foreman et al. (2012) (Picenance Creek Basin, Wasatch Formation near De Beque, Colorado); paleogeography shown as uplifted source areas in gray, fluvial environments in green, shallow and deep marine in light blue and dark blue. Stratigraphic logs show architecture through PETM intervals in each basin; onset of coarse-grained deposition noted with blue line. Carbon isotope sections show moving averages of sampled data through the intervals represented by the stratigraphic logs; onset and termination of the PETM-defining carbon isotope excursion indicated by red lines. This comparison shows the apparent lag between the onset of the PETM and the stratigraphic response. This lag time was quantified by Duller et al. (2019) and provides insight into landscape-response times. Overview photos: Each basin shows a marked coarsening of overall stratigraphic architecture in the vicinity of the PETM (Baisden, 2014; Greenberg, 2016). Close-up photos: Example channel belt sandstone bodies from each basin with preserved bar-clinoform deposits (highlighted with white lines) that can be used to approximate paleoflow depths.

know that all records are incomplete to some extent; the larger the range of depositional variability relative to the long-term sedimentation rate, the more missing time is likely in any individual record (Trampush & Hajek, 2017). This can cause particular challenges in estimating rates of change (i.e., inverting stratigraphic spatial series into a time series), for example, the rate of onset of the carbon isotope excursion or the rate of recovery (Bowen et al., 2015).

Importantly, solving this “rates” issue is critical if we are to reasonably compare the PETM and other global warming events to anthropogenic changes accurately (Gingerich, 2019). One way forward would be to identify basins with extremely short T_c and T_{eq} to maximize the potential resolution of proxy records. Furthermore, some proxy systems (e.g., those that develop within soil horizons) exploit the “gaps” in the record, so leveraging different proxy systems in combination may help overcome some issues of temporal incompleteness. Moving forward, it will be useful to find basins with significantly longer autogenic timescales such that autogenic processes would shred the PETM signals and/or target shorter Eocene hyperthermal events in these same basins, which may be shredded. Framing both proxy records and extracting geomorphic information in the context of these autogenic scales is necessary to establish time series of change rigorously as well as quantifying autogenic variability to distinguish it from allogenic signals.

6. Future Directions

The core of this synthesis lies in defining processes that reduce the fidelity of environmental signals stored in stratigraphy. A central, take-home message generated from an analysis of the timescales associated with these signal storage impediments is that they overlap with many mesotimescale environmental forcings. While in some sense this is unfortunate, we take an optimistic view. We highlight that an appropriate appreciation of these impediments, grounded in their quantification, improves our ability to identify basins or parts of basins with high signal storage potential. In general, we find that smaller systems with higher aggradation rates are better at recording environmental signals due to their smaller and shorter autogenic scales, but the storage capacity of any basin can be estimated given measurements of emergent landscape scales. A key future direction thus will be implementing this theory to predict signal storage capacity when designing data collection and interpretation campaigns. However, our ability to predict with high precision the storage capacity of stratigraphic records is still limited and requires focused research on several fronts.

While those sedimentary basins that are characterized by large and long autogenic scales might be poor record keepers of environmental change, the plus side is that the stratigraphy in these basins can be used to address other questions. For example, the stratigraphy of large basins with slow accumulation rates is likely biased toward the products of autogenic processes, which could be used to characterize surface process scales and rates and improve our understanding of autogenic morphodynamics, particularly those resulting from longer-term processes that are difficult to observe in modern system (e.g., channel avulsion).

Moving forward, we define four main avenues of opportunity for the community, each of which will improve our capacity to reconstruct paleoenvironmental conditions from stratigraphy or to perform stratigraphic prediction.

6.1. Defining the Morphodynamic Roots of Landscape Stochasticity Across Depositional Environments

We are at a point where constructing more complicated deterministic models to aid our stratigraphic interpretations will be less effective than efforts to further understand the causes and scales of stochastic processes. In many depositional environments the upper limits on this stochasticity can be formulated from knowledge of the maximum timescale of autogenic processes. For example, in the last decade an expansion in our appreciation of backwater hydrodynamics and their implication for depositional mechanics (Lamb et al., 2012; Nittrouer et al., 2012) enhanced our appreciation of key autogenic marginal marine stratigraphic scales (i.e., autogenic parasequence and erosional scour scales) (Fernandes et al., 2016; Ganti et al., 2019; Straub et al., 2015; Trower et al., 2018). While every depositional environment must have an upper bound on autogenic scales, we lack morphodynamic descriptions that predict stochastic processes and stratigraphic products for many environments. This is particularly true for unchannelized strata, for example, some deep marine records, where our understanding of the source and magnitude of stochastic surface fluctuations relative to background deposition rates is in its infancy. It also holds for nonclastic depositional settings, for example, carbonate systems, where biogeochemical processes also have a stochastic component that results in stratigraphic hiatuses and autogenic fluctuations in aggradation rates (Kemp & Van Manen, 2019; Kim et al., 2012; Purkis et al., 2016).

Defining the causes and scales of morphodynamic stochasticity will be critical to applying signal transfer thresholds to field-scale systems. For example, Toby et al. (2019) recognized that the scale of autogenic volume fluctuations that defines transfer thresholds likely varies as a function of the mean and stochastic components of a system's forcing. Work to define the magnitude of autogenic fluctuations as functions of forcing conditions is starting to accelerate, with studies quantifying autogenic scales as functions of the ratio of sediment to water supply (Powell et al., 2012; Straub & Wang, 2013), sediment grain size (Caldwell & Edmonds, 2014) and cohesion (Edmonds & Slingerland, 2010; Hoyal & Sheets, 2009; Li et al., 2017), vegetation (Lauzon & Murray, 2018; Piliouras et al., 2017), flashiness of system hydrographs (Esposito et al., 2018; Ganti et al., 2019; Miller et al., 2019), basin water depth (Carlson et al., 2018), and wave (Ratliff et al., 2018) and tidal climate (Kleinhans et al., 2015; Lentsch et al., 2018).

We recognize a need also to characterize the response of systems to time-varying forcing. We simply do not have enough data to characterize how systems reconfigure themselves as they respond transiently, and this applies to relatively well understood fluvial landscapes where we still struggle to predict how a given river will adjust in response to, for example, an increase in sediment discharge. These reconfigurations result in stratigraphic products defined by scales that may be difficult to differentiate from scales of autogenic reconfigurations and so need to be studied in the context of the buffered-transient, buffered-steady framework.

In addition to new theory and measurements from physical experiments, we need more field data that captures the stochastic scales in depositional environments. Currently, we rely strongly on the Sadler database of deposition rates (Sadler, 1981; Sadler & Jerolmack, 2015) and their measurement span to define scaling relationships for stratigraphic completeness. However, some depositional environments (e.g., alluvial plains and continental shelves) are better sampled than others (e.g., terrestrial floodplains, continental slopes, and abyssal plains), mainly due to biases in where previous field campaigns have sampled. Further campaigns, similar to those of Vendettuoli et al. (2019) that seek to capture how time gets stored in stratigraphy through direct measurements, and continued expansion of databases like Sadler's, will help us close these loops.

6.2. Resolving the Influence of Morphodynamics on the Production and Resetting Timescales of Stratigraphic Proxies

In this review we have taken a generic view on sedimentary signals. The morphodynamic processes that drive buffering, incompleteness, and signal shredding primarily influence local sediment flux and intermittency between depositional and erosional events. There is significant opportunity to explore these dynamics more thoroughly and perhaps overcome certain aspects of signal degradation by leveraging different types of proxies that respond differently to sedimentation rate, exposure time, or sediment mixing. For example, soils form during periods of nondeposition and therefore mark temporal gaps in sedimentary records, and geochemical proxies found within these surfaces likely average over long (soil forming) periods of time. In contrast, organic matter preservation is facilitated by high sedimentation rates; consequently, preserved organic material may record specific environmental conditions from short windows of time.

Reexhumation of previously deposited material in a SRS can also have a significant impact on proxy records. Bulk organic carbon records, for example, sample across several types of organic carbon each of which has a different sensitivity to degradation upon exposure to oxygen. Over the length of a SRS, biomass that is sequentially buried and exhumed will degrade quickly relative to soil organic carbon, and petrogenic or "fossil" organic carbon may not experience significant degradation at all (Blair & Aller, 2012). Changes in organic carbon pools and preservation viewed through a lens of landscape dynamics may provide deeper insight into the nature of landscape response to external forcing and can help interpret whether confusing proxy records may be impacted by sediment reworking and landscape dynamics (e.g., Baczynski et al., 2016; Lyons et al., 2019).

6.3. Moving Toward Comprehensive Characterization of the Lithostratigraphic Signature of Environmental Signals

We recognize that much of the theory outlining the impediments environmental signals face for stratigraphic storage focuses on fluctuation in the elevation of the Earth's surface and therefore the thickness of depositional units. Other measurable characteristics provide important information. For example, deposit grain size (both mean and sorting) can inform on paleoflow conditions and petrology of source material, while closure age of individual minerals (e.g., detrital zircons) found in strata can be used to reconstruct the paleogeography of SRSs and understand how different sediment sources contributed to and were mixed in sediment transport networks. At a larger scale the spatial arrangement of facies and facies associations, commonly referred to as stratigraphic architecture, can hold information about signals of environmental forcings, such as changes in accommodation-creation rate or sediment supply. At present, though, we simply do not know how to use many of these properties to quantitatively reconstruct mesotimescale forcings or even whether these properties can be used for accurate signal extraction.

Existing theory provides a basis for exploring and interpreting grain size trends in SRSs. Several recent modeling studies have used a similarity solution for fluvial sediment fining by selective deposition (Fedele & Paola, 2007) in mass-conserving models to predict how deposit texture should vary in response to changing environmental conditions. Field deployment of such models remains hampered by outstanding questions about how to spatially average grain size observations to account for local variability imparted by autogenic processes (D'Arcy et al., 2017) and the fact that existing models only effectively describe bedload transport and are therefore generally suitable only for use in gravel and sand-dominated deposits. Overcoming these challenges would open up new depositional environments for quantitative environmental reconstructions for mesotimescale forcings from deposit texture.

Improvements in methods for dating detrital zircons have led to a dramatic increase in the number of sediment provenance studies in recent years (Blum et al., 2018; Cawood et al., 2012; Fedo et al., 2003). These methods complement traditional geochemical tracers (Weltje & von Eynatten, 2004), enhancing available methods for reconstructing sediment source areas and providing new constraints on paleosediment flux estimates. Detrital zircon studies provide an independent line of evidence for sediment flux estimates that compliment reconstructions achieved from deposit texture (Mahon & McElroy, 2018) or isopach reconstruction (Allen et al., 2013). Combining sediment provenance with other attributes measured from stratigraphic sections will help reduce uncertainty associated with nonunique solutions. As a demonstration, Sharman et al. (2019) used a landscape evolution model divided into two sediment provenance regions to track both the flux of material exiting catchments and the fraction of the material exiting each provenance for landscapes subject to either climatic or tectonic perturbations. The simultaneous tracking of sediment provenances and sediment fluxes enables the detection and differentiation of erosion rates that spatially vary due to either knickpoint propagation or slope adjustment triggered by precipitation changes (Paola & Swenson, 1998). More exploration of propagation and mixing of sediment along routing systems will be necessary, though, before they can be deployed for quantitative environmental signal reconstructions. Specifically, we need better theory to describe time and space scales of mixing relative to differences in provenance signatures of sediments.

Facies architecture and stacking patterns are long-standing evidence used to interpret changes in environmental conditions. Facies models of depositional environments, defined by foundational field observation like the composition, texture, form, and fossil content of sediment beds, provide a major opportunity to incorporate updated process-based understanding into stratigraphic interpretations. We hark back to a suggestion made by Paola (2016), who challenged the field data collection community, to track signals of known onset and duration from a signal source to sediment sink. This would allow the facies, facies association, and architectural element signature of cleanly preserved signals to be compared directly to the signature of purely autogenic stratigraphy resulting from depocenters that are efficient at signal shredding. However, we should remember that facies are the result of short-timescale processes and so we need to identify the spatial scales in strata that link to mesotimescales if we are to invert for mesotimescale forcings from stratigraphy.

6.4. Training Stratigraphers in the 21st Century

The first generation of computational stratigraphic models was developed in the late 1970s (Allen, 1978; Bridge & Leeder, 1979; Leeder, 1978). Since that time, the field of stratigraphy has undergone a quantitative revolution. This revolution was facilitated by parallel advances in our quantitative description of Earth surface dynamics (Paola, 2000; Slingerland & Smith, 1998; Whipple, 2001). We now have numerical models that use our detailed physical descriptions of fluid and sediment transport that can simulate the evolution of basin strata over geological timescales and their response to changing environmental conditions. We have also developed physical experimental techniques that allow us to explore how myriad environmental conditions influence sediment transport and the construction of strata. This quantitative revolution is only going to accelerate as we approach the middle of the 21st century. Techniques to harness Big Data, including machine learning, are starting to be employed in the fields of Earth surface processes and sedimentary geology and will undoubtedly help solve today's outstanding scientific problems in these fields and likely open avenues of subject understanding that we cannot begin to predict.

To ensure that we as a community are positioned to harness such innovations, a realistic assessment of our state-of-readiness is required. As one community we have at our disposal a vast array of skills, some of which are quantitative and are aligned with the “Big Data revolution” and others that are qualitative and are aligned with a more traditional way of collecting information in the field. While this range of skills attests to the good health of our science, a fundamental pedagogic or philosophical hurdle appears to be preventing us from attaining a desired state of readiness. There is still a tendency to offer student training in either quantitative processes stratigraphy or qualitative observational stratigraphy. As a result, some quantitative stratigraphers can become enamored with theory built from high spatial and temporal resolution data produced by numerical and physical experiments, whereas some field stratigraphers focus their attention on the complexities of individual field locales. This divergence over time has led to a community that speaks different scientific dialects in the 21st century. Some terms that field stratigraphers view as jargon, quantitative stratigraphers use because they have precise definitions rooted in mathematics and physics. On the other side, debates about nomenclature that some quantitative stratigraphers view as pedantic are essential to field stratigraphers needing to describe specific observations with important interpretive implications. While reasons exist for the two dialects, they prevent us from singing from the same song sheet in order to construct and test robust hypotheses.

Overcoming these divides will require quantitative and field stratigraphers working together to train a new generation of quantitative field stratigraphers. Achieving this goal can be facilitated by real and visible attempts by the leaders of today to set an example for the leaders of tomorrow. This will help us reach our goal of isolating the signals of a changing environment in strata, which surpass null hypotheses. This next generation should be trained both in the skills necessary to employee quantitative theory and the techniques to gather meaningful data from field-scale stratigraphy (whether from cores, logs, outcrops, or seismic data). This could be achieved with more opportunities for students to be jointly advised during their training by representatives of both camps. It will also be critical, though, for new stratigraphers to be encouraged to take courses in advanced mathematics and computational data analysis and to provide them ample opportunity to collect and interpret field data sets. Students comfortable with the methods central to the quantitative sciences and the realities of data collection from strata will be positioned to pose and test questions of societal importance and acknowledge the uncertainties of our interpretations of Earth's past.

7. Summary

The stratigraphic record contains an immense amount of information which can be used to improve our understanding of the Earth system, which, if properly mined, could enhance our understanding of Earth history and aid our predictions of its future. Similar records preserved on other planetary bodies can also be mined to unravel the history of the solar system. The stratigraphic recording process, however, occurs in fits and starts, which discretizes a continuous time series into preserved strata. This stratigraphic filter occurs after transmission of environmental signals across landscapes, which also has the tendency to degrade signals. We propose that the magnitude of these impediments to signal storage and recovery is set by lengthscales and timescales that arise from self-organized processes in landscapes and environmental stochasticity. Techniques to predict and measure these scales exist but need refining. Specifically, we need better models and data to help characterize the scales of stochastic surface fluctuations in regions where stratigraphic records are produced. In addition, much of the theory, developed from observations of numerical and physical experiments, still needs to be tested against field-scale stratigraphy. Advances along these lines will help us define uncertainty in the storage of time in stratigraphy and in discriminating the products of stochastic processes from the environmental records that many seek to extract from stratigraphy.

Acknowledgments
We wish to thank our respective research groups, both past and present, for scientific conversations that led to the thoughts and ideas presented here. In particular, the exhaustive experimental campaigns of Qi Li and Stephan Toby and their creative data analysis were critical for our ability to produce this synthesis. Farther afield, numerous conversations with Chris Paola, Philip Allen, David Mohrig, Vamsi Ganti, Peter Burgess, Sheila Trampush, and Peter Sadler greatly influenced the ideas contained here. This study was supported in part by the National Science Foundation (EAR-1424312) to K. M. S. (OCE-141663 and EAR-1455240) to E. A. H. and the National Geographic Society Research and Exploration Grant (9867-16) to B. Z. F. We appreciate the invitation from the JGR Earth Surface editorial board, specifically Amy East, to write this paper. We also extend our thanks to Jon Pelletier, Mike Blum, Peter DeCelles, and two anonymous reviewers for comprehensive reviews that helped add clarity to the paper. New calculations reported on in this this manuscript used two data sets, which can be found online (http://doi.org/10.5967/M07D2S7Q and http://doi.org/10.5967/M0HX19TT).

References

Aalto, R., Maurice-Bourgoin, L., Dunne, T., Montgomery, D. R., Nittrouer, C. A., & Guyot, J.-L. (2003). Episodic sediment accumulation on Amazonian flood plains influenced by El Niño/Southern Oscillation. *Nature*, *425*(6957), 493–497. https://doi.org/10.1038/nature02002
Aalto, R., & Nittrouer, C. A. (2012). ^{210}Pb geochronology of flood events in large tropical river systems, *Philosophical Transactions of the Royal Society A: Mathematical. Physical and Engineering Sciences*, *370*(1966), 2040–2074.
Abels, H. A., Aziz, H. A., Krijgsman, W., Smeets, S. J. B., & Hilgen, F. J. (2010). Long-period eccentricity control on sedimentary sequences in the continental Madrid Basin (middle Miocene, Spain). *Earth and Planetary Science Letters*, *289*(1-2), 220–231. https://doi.org/10.1016/j.epsl.2009.11.011
Abels, H. A., Kraus, M. J., & Gingerich, P. D. (2013). Precession-scale cyclicity in the fluvial lower Eocene Willwood Formation of the Bighorn Basin, Wyoming (USA). *Sedimentology*, *60*(6), 1467–1483.
Adams, P. N., Slingerland, R. L., & Smith, N. D. (2004). Variations in natural levee morphology in anastomosed channel flood plain complexes. *Geomorphology*, *61*, 127–142.
Ager, D. V. (1973). *The Nature of the Stratigraphical Record* (114 pp.). Wiley.
Allen, J. (1974). Reaction, relaxation and lag in natural sedimentary systems: General principles, examples and lessons. *Earth-Science Reviews*, *10*(4), 263–342.
Allen, J. R. (1963). The classification of cross-stratified units. *With notes on their origin, Sedimentology*, *2*(2), 93–114.
Allen, J. R. L. (1978). Studies in fluviatile sedimentation: An exploratory quantitative model for the architecture of avulsion-controlled alluvial sites. *Sedimentary Geology*, *26*, 617–644.
Allen, P. A. (2008). Time scales of tectonic landscapes and their sediment routing systems. In K. Gallager, S. J. Jones, & J. Wainwright (Eds.), *Landscape evolution: Denudation, climate and tectonics over different time and space scales* (pp. 7–28). London.
Allen, P. A. (2017). *Sediment routing systems: The fate of sediment from source to sink*. Cambridge, UK: Cambridge University Press.
Allen, P. A., Armitage, J. J., Carter, A., Duller, R. A., Michael, N. A., Sinclair, H. D., et al. (2013). The Qs problem: Sediment volumetric balance of proximal foreland basin systems. *Sedimentology*, *60*(1), 102–130. https://doi.org/10.1111/Sed.12015
Anderson, S. W., & Konrad, C. P. (2019). Downstream-propagating channel responses to decadal-scale climate variability in a glaciated river basin. *Journal of Geophysical Research: Earth Surface*, *124*, 902–919.
Anderton, R. (1985). Clastic facies models and facies analysis, *Geological Society*, *London*. *Special Publications*, *18*(1), 31–47.

Armitage, J. J., Allen, P. A., Burgess, P. M., Hampson, G. J., Whittaker, A. C., Duller, R. A., & Michael, N. A. (2015). Sediment transport model for the eocene escanilla sediment-routing system: Implications for the uniqueness of sequence stratigraphic architectures. *Journal of Sedimentary Research*, *85*(12), 1510–1524.

Armitage, J. J., Burgess, P. M., Hampson, G. J., & Allen, P. A. (2018). Deciphering the origin of cyclical gravel front and shoreline progradation and retrogradation in the stratigraphic record. *Basin Research*, *30*, 15–35.

Armitage, J. J., Duller, R. A., Whittaker, A. C., & Allen, P. A. (2011). Transformation of tectonic and climatic signals from source to sedimentary archive. *Nature Geoscience*, *4*(4), 231–235. https://doi.org/10.1038/Ngeo1087

Armitage, J. J., Jones, T. D., Duller, R. A., Whittaker, A. C., & Allen, P. A. (2013). Temporal buffering of climate-driven sediment flux cycles by transient catchment response. *Earth and Planetary Science Letters*, *369*, 200–210.

Ashton, A., Murray, B., & Arnault, O. (2001). Formation of coastline features by large-scale instabilities induced by high-angle waves. *Nature*, *414*, 296–300.

Aswasereelert, W., Meyers, S., Carroll, A., Peters, S., Smith, M., & Feigl, K. (2013). Basin-scale cyclostratigraphy of the Green River Formation, Wyoming. *Bulletin*, *125*(1-2), 216–228.

Aziz, H. A., Di Stefano, A., Foresi, L. M., Hilgen, F. J., Laccarino, S. M., Kuiper, K. F., et al. (2008). Integrated stratigraphy and (40)Ar/(39) Ar chronology of early Middle Miocene sediments from DSDP Leg 42A, Site 372 (Western Mediterranean). *Palaeogeography Palaeoclimatology Palaeoecology*, *257*(1-2), 123–138. https://doi.org/10.1016/j.palaeo.2007.09.013

Aziz, H. A., Hilgen, F. J., van Luijk, G. M., Sluijs, A., Kraus, M. J., Pares, J. M., & Gingerich, P. D. (2008). Astronomical climate control on paleosol stacking patterns in the upper Paleocene-lower Eocene Willwood Formation, Bighorn Basin, Wyoming. *Geology*, *36*(7), 531–534. https://doi.org/10.1130/G24734a.1

Baczynski, A. A., McInerney, F. A., Wing, S. L., Kraus, M. J., Morse, P. E., Bloch, J. I., et al. (2016). Distortion of carbon isotope excursion in bulk soil organic matter during the Paleocene-Eocene thermal maximum. *Bulletin*, *128*(9-10), 1352–1366.

Baisden, T. R. (2014). Piceance (Molina V-road). http://www.gigapan.com/gigapans/162418, web reference.

Barrell, J. (1917). Rhythms and the measurements of geologic time. *Geological Society of America Bulletin*, *28*(1), 745–904.

Beerbower, J. R. (1961). Origin of cyclothems of the Dunkard group (upper Pennsylvanian-lower Permian) in Pennsylvania, West Virginia, and Ohio. *Geological Society of America Bulletin*, *72*(7), 1029–1050.

Begin, Z. E., Meyer, D. F., & Schumm, S. A. (1981). Development of longitudinal profiles of alluvial channels in response to base-level lowering. *Earth Surface Processes and Landforms*, *6*(1), 49–68.

Bentley, S., Blum, M., Maloney, J., Pond, L., & Paulsell, R. (2016). The Mississippi River source-to-sink system: Perspectives on tectonic, climatic, and anthropogenic influences. *Miocene to Anthropocene, Earth-Science Reviews*, *153*, 139–174.

Bhattacharya, J. P., Miall, A. D., Ferron, C., Gabriel, J., Randazzo, N., Kynaston, D., et al. (2019). Time-stratigraphy in point sourced river deltas: Application to sediment budgets, shelf construction, and paleo-storm records. *Earth-Science Reviews*, *199*, 102,985.

Blair, N. E., & Aller, R. C. (2012). The fate of terrestrial organic carbon in the marine environment. *Annual Review of Marine Science*, *4*, 401–423. https://doi.org/10.1146/annurev-marine-120709-142717

Blum, M., Martin, J., Milliken, K., & Garvin, M. (2013). Paleovalley systems: Insights from Quaternary analogs and experiments. *Earth-Science Reviews*, *116*, 128–169. https://doi.org/10.1016/j.earscirev.2012.09.003

Blum, M., Rogers, K., Gleason, J., Najman, Y., Cruz, J., & Fox, L. (2018). Allogenic and autogenic signals in the stratigraphic record of the deep-sea Bengal Fan. *Scientific Reports*, *8*(1), 7973. https://doi.org/10.1038/s41598-018-25819-5

Boulesteix, K., Poyatos-Moré, M., Flint, S. S., Taylor, K. G., Hodgson, D. M., & Hasiotis, S. T. (2019). Transport and deposition of mud in deep-water environments: Processes and stratigraphic implications. *Sedimentology*.

Bowen, G. J., Maibauer, B. J., Kraus, M. J., Röhl, U., Westerhold, T., Steimke, A., et al. (2015). Two massive, rapid releases of carbon during the onset of the Palaeocene–Eocene thermal maximum. *Nature Geoscience*, *8*(1), 44.

Boyd, R., Dalrymple, R., & Zaitlin, B. (1992). Classification of clastic coastal depositional environments. *Sedimentary Geology*, *80*(3-4), 139–150.

Bridge, J. S., & Leeder, M. R. (1979). A simulation model of alluvial stratigraphy. *Sedimentology*, *26*, 617–644.

Bridgland, D. R., & Westaway, R. (2008). Preservation patterns of Late Cenozoic fluvial deposits and their implications: Results from IGCP 449. *Quaternary International*, *189*(1), 5–38.

Brunsden, D., & Thornes, J. (1979). Landscape sensitivity and change. *Transactions of the Institute of British Geographers*, *4*(4), 463–484.

Bull, W. B. (1991). *Geomorphic responses to climatic change*. Oxford, UK: Oxford University Press.

Burgess, P. M., Masiero, I., Toby, S. C., & Duller, R. A. (2019). A big fan of signals? Exploring autogenic and allogenic process and product in a numerical stratigraphic forward model of submarine-fan development. *Journal of Sedimentary Research*, *89*(1), 1–12.

Caldwell, R. L., & Edmonds, D. A. (2014). The effects of sediment properties on deltaic processes and morphologies: A numerical modeling study. *Journal of Geophysical Research: Earth Surface*, *119*, 961–982.

Carlson, B., Piliouras, A., Muto, T., & Kim, W. (2018). Control of basin water depth on channel morphology and autogenic timescales in deltaic systems. *Journal of Sedimentary Research*, *88*(9), 1026–1039.

Carmichael, M. J., Inglis, G. N., Badger, M. P., Naafs, B. D. A., Behrooz, L., Remmelzwaal, S., et al. (2017). Hydrological and associated biogeochemical consequences of rapid global warming during the Paleocene-Eocene Thermal Maximum. *Global and Planetary Change*, *157*, 114–138.

Carson, R. (1951). *The sea around us*. New York: Oxford University Press.

Castelltort, S., & Van den Driessche, J. (2003). How plausible are high-frequency sediment supply-driven cycles in the stratigraphic record? *Sedimentary Geology*, *157*(1-2), 3–13. https://doi.org/10.1016/S0037-0738(03)00066-6

Catuneanu, O., Willis, A. J., & Miall, A. D. (1998). Temporal significance of sequence boundaries. *Sedimentary Geology*, *121*(3-4), 157–178.

Cawood, P. A., Hawkesworth, C., & Dhuime, B. (2012). Detrital zircon record and tectonic setting. *Geology*, *40*(10), 875–878.

Chatanantavet, P., Lamb, M. P., & Nittrouer, J. A. (2012). Backwater controls of avulsion location on deltas. *Geophysical Research Letters*, *39*, L01402. https://doi.org/10.1029/2011gl050197.

Chow, V. T. (1959). *Open-channel hydraulics* (680 pp.). New York: McGraw-Hill.

Ciarletta, D. J., Lorenzo-Trueba, J., & Ashton, A. (2019). Mechanism for retreating barriers to autogenically form periodic deposits on continental shelves. *Geology*, *47*(3), 239–242.

Clift, P. D., Hodges, K. V., Heslop, D., Hannigan, R., Van Long, H., & Calves, G. (2008). Correlation of Himalayan exhumation rates and Asian monsoon intensity. *Nature Geoscience*, *1*(12), 875.

Clyde, W. C., Hamzi, W., Finarelli, J. A., Wing, S. L., Schankler, D., & Chew, A. (2007). Basin-wide magnetostratigraphic framework for the Bighorn Basin, Wyoming. *Geological Society of America Bulletin*, *119*(7-8), 848–859.

Coffey, T. S., & Shaw, J. B. (2017). Congruent bifurcation angles in river delta and tributary channel networks. *Geophysical Research Letters*, *44*, 11,427–11,436. https://doi.org/10.1002/2017gl074873

Colombera, L., Arévalo, O. J., & Mountney, N. P. (2017). Fluvial-system response to climate change: The Paleocene-Eocene tremp group. *Pyrenees, Spain, Global and Planetary Change*, *157*, 1–17.

Coulthard, T., & Van de Wiel, M. (2013). Climate, tectonics or morphology: What signals can we see in drainage basin sediment yields? *Earth Surface Dynamics*, *1*(1), 13–27.

Covault, J. A., Craddock, W. H., Romans, B. W., Fildani, A., & Gosai, M. (2013). Spatial and temporal variations in landscape evolution: Historic and longer-term sediment flux through global catchments. *The Journal of Geology*, *121*(1).

Covault, J. A., Romans, B. W., Fildani, A., McGann, M., & Graham, S. A. (2010). Rapid climatic signal propagation from source to sink in a Southern California sediment-routing system. *Journal of Geology*, *118*(3), 247–259. https://doi.org/10.1086/651539

Crouch, E. M., Dickens, G. R., Brinkhuis, H., Aubry, M.-P., Hollis, C. J., Rogers, K. M., & Visscher, H. (2003). The Apectodinium acme and terrestrial discharge during the Paleocene–Eocene thermal maximum: New palynological, geochemical and calcareous nannoplankton observations at Tawanui, New Zealand. *Palaeogeography, Palaeoclimatology, Palaeoecology*, *194*(4), 387–403.

Currano, E. D., Wilf, P., Wing, S. L., Labandeira, C. C., Lovelock, E. C., & Royer, D. L. (2008). Sharply increased insect herbivory during the Paleocene–Eocene Thermal Maximum. *Proceedings of the National Academy of Sciences*, *105*(6), 1960–1964.

Dade, W. B., & Friend, P. F. (1998). Grain-size, sediment-transport regime, and channel slope in alluvial rivers. *The Journal of Geology*, *106*(6), 661–676.

D'Arcy, M., Whittaker, A. C., & Roda-Boluda, D. C. (2017). Measuring alluvial fan sensitivity to past climate changes using a self-similarity approach to grain-size fining, Death Valley, California. *Sedimentology*, *64*(2), 388–424.

Davies, N. S., & Gibling, M. R. (2010). Cambrian to Devonian evolution of alluvial systems: the sedimentological impact of the earliest land plants. *Earth-Science Reviews*, *98*(3), 171–200.

Davies, N. S., Shillito, A. P., & McMahon, W. J. (2019). Where does the time go? Assessing the chronostratigraphic fidelity of sedimentary geological outcrops in the Pliocene–Pleistocene Red Crag Formation, eastern England. *Journal of the Geological Society*, *176*(6), 1154–1168.

Densmore, A. L., Allen, P. A., & Simpson, G. (2007). Development and response of a coupled catchment fan system under changing tectonic and climatic forcing. *Journal of Geophysical Research*, *112*.

Dodds, P. S., & Rothman, D. H. (2000). Scaling, universality, and geomorphology. *Annual Reviews of Earth Sciences*, *571-610*.

Donnelly, J. P., & Woodruff, J. D. (2007). Intense hurricane activity over the past 5,000 years controlled by El Niño and the West African monsoon. *Nature*, *447*(7143), 465–468. https://doi.org/10.1038/nature05834

Duller, R. A., Armitage, J. J., Manners, H. R., Grimes, S., & Jones, T. D. (2019). Delayed sedimentary response to abrupt climate change at the Paleocene-Eocene boundary, northern Spain. *Geology*, *47*(2), 159–162.

Duller, R. A., Warner, N. H., McGonigle, C., De Angelis, S., Russell, A. J., & Mountney, N. P. (2014). Landscape reaction, response, and recovery following the catastrophic 1918 Katla jökulhlaup, southern Iceland. *Geophysical Research Letters*, *41*, 4214–4221.

Duller, R. A., Whittaker, A. C., Fedele, J. J., Whitchurch, A. L., Springett, J., Smithells, R., et al. (2010). From grain size to tectonics. *Journal of Geophysical Research*, *115*, F03022. https://doi.org/10.1029/2009jf001495

Duller, R. A., Whittaker, A. C., Swinehart, J. B., Armitage, J. J., Sinclair, H. D., Bair, A., & Allen, P. A. (2012). Abrupt landscape change post-6 Ma on the central Great Plains, USA. *Geology*, *40*(10), 871–874.

Durkin, P. R., Hubbard, S. M., Holbrook, J., & Boyd, R. (2018). Evolution of fluvial meander-belt deposits and implications for the completeness of the stratigraphic record. *GSA Bulletin*, *130*(5-6), 721–739.

Edmonds, D. A., & Slingerland, R. L. (2010). Significant effect of sediment cohesion on delta morphology. *Nature Geoscience*, *3*(2), 105–109. https://doi.org/10.1038/Ngeo730

Einstein, H. A. (1950). *The bed-load function for sediment transportation in open channel flows Rep*. U.S. Dept. of Agriculture Technical Bulletin No. 1026.

Esposito, C. R., Di Leonardo, D., Harlan, M., & Straub, K. M. (2018). Sediment storage partitioning in alluvial stratigraphy: The influence of discharge variability. *Journal of Sedimentary Research*, *88*(6), 717–726.

Fan, M., Ayyash, S. A., Tripati, A., Passey, B. H., & Griffith, E. M. (2018). Terrestrial cooling and changes in hydroclimate in the continental interior of the United States across the Eocene-Oligocene boundary. *GSA Bulletin*, *130*(7-8), 1073–1084.

Fedele, J. J., & Paola, C. (2007). Similarity solutions for fluvial sediment fining by selective deposition. *Journal of Geophysical Research*, *112*.

Fedo, C. M., Sircombe, K. N., & Rainbird, R. H. (2003). Detrital zircon analysis of the sedimentary record. *Reviews in Mineralogy and Geochemistry*, *53*(1), 277–303.

Fernandes, A. M., Törnqvist, T. E., Straub, K. M., & Mohrig, D. (2016). Connecting the backwater hydraulics of coastal rivers to fluvio-deltaic sedimentology and stratigraphy. *Geology*, *44*(12), 979–982.

Finnegan, N. J., & Dietrich, W. E. (2011). Episodic bedrock strath terrace formation due to meander migration and cutoff. *Geology*, *39*(2), 143–146.

Finnegan, N. J., Roe, G., Montgomery, D. R., & Hallet, B. (2005). Controls on the channel width of rivers: Implications for modeling fluvial incision of bedrock. *Geology*, *33*(3), 229–232.

Flemings, P. B., & Jordan, T. E. (1989). A synthetic stratigraphic model of foreland basin development. *Journal of Geophysical Research*, *94*(B4), 3851–3866.

Flower, B. P., Hastings, D. W., Hill, H. W., & Quinn, T. M. (2004). Phasing of deglacial warming and Laurentide Ice Sheet meltwater in the Gulf of Mexico. *Geology*, *32*(7), 597–600.

Foreman, B. (2014). Climate-driven generation of a fluvial sheet sand body at the Paleocene–Eocene boundary in northwest Wyoming (USA). *Basin Research*, *26*(2), 225–241.

Foreman, B. Z., Heller, P. L., & Clementz, M. T. (2012). Fluvial response to abrupt global warming at the Palaeocene/Eocene boundary. *Nature*, *491*(7422), 92–95. https://doi.org/10.1038/Nature11513

Foreman, B. Z., & Straub, K. M. (2017). Autogenic geomorphic processes determine the resolution and fidelity of terrestrial paleoclimate records. *Science Advances*, *3*(9), e1700683.

Forzoni, A., Storms, J. E., Whittaker, A. C., & de Jager, G. (2014). Delayed delivery from the sediment factory: Modeling the impact of catchment response time to tectonics on sediment flux and fluvio-deltaic stratigraphy. *Earth Surface Processes and Landforms*, *39*(5), 689–704.

Furbish, D. J., Haff, P. K., Roseberry, J. C., & Schmeeckle, M. W. (2012). A probabilistic description of the bed load sediment flux: 1. Theory. *Journal of Geophysical Research*, *117*.

Gani, M. R., & Gani, N. D. (2011). River-margin habitat of Ardipithecus ramidus at Aramis, Ethiopia 4.4 million years ago. *Nature Communications*, *2*, 602.

Ganti, V., Lamb, M. P., & Chadwick, A. J. (2019). Autogenic erosional surfaces in fluvio-deltaic stratigraphy from floods, avulsions, and backwater hydrodynamics. *Journal of Sedimentary Research*, *89*, 1–18.

Ganti, V., Paola, C., & Foufoula-Georgiou, E. (2013). Kinematic controls on the geometry of the preserved cross sets. *Journal of Geophysical Research: Earth Surface*, *118*, 1296–1307.

Ganti, V., Straub, K. M., Foufoula-Georgiou, E., & Paola, C. (2011). Space-time dynamics of depositional systems: Experimental evidence and theoretical modeling of heavy-tailed statistics. *Journal of Geophysical Research*, *116*, F02011. https://doi.org/10.1029/2010jf001893

Gasparini, N. M., Whipple, K. X., & Bras, R. L. (2007). Predictions of steady state and transient landscape morphology using sediment-flux-dependent river incision models. *Journal of Geophysical Research*, *112*.

Gilbert, G. K. (1890). Lake Bonneville. *U.S. Geologic Survey Monograph*, *1*.

Gingerich, P. D. (2006). Environment and evolution through the Paleocene–Eocene thermal maximum. *Trends in Ecology & Evolution*, *21*(5), 246–253.

Gingerich, P. D. (2019). Temporal scaling of carbon emission and accumulation rates: Modern anthropogenic emissions compared to estimates of PETM onset accumulation. *Paleoceanography and Paleoclimatology*, *34*(3), 329–335.

Goldfinger, C.Nelson, C. H., Morey, A. E., Johnson, J. E., Patton, J. R., Karabanov, E. B., et al. (2012). Turbidite event history—Methods and implications for Holocene paleoseismicity of the Cascadia subduction zone (No. 1661-F). US Geological Survey.

Goodbred, S. L. (2003). Response of the Ganges dispersal system to climate change: A source-to-sink view since the last interstade. *Sedimentary Geology*, *162*(1-2), 83–104.

Goudge, T. A., Mohrig, D., Cardenas, B. T., Hughes, C. M., & Fassett, C. I. (2018). Stratigraphy and paleohydrology of delta channel deposits. *Jezero crater, Mars, Icarus*, *301*, 58–75.

Greenberg, E. B. (2016). Bighorn (boundary SS Saddle Mtn). http://gigapan.com/gigapans/190416, web reference.

Grotzinger, J. P., Arvidson, R., Bell Iii, J., Calvin, W., Clark, B., Fike, D., et al. (2005). Stratigraphy and sedimentology of a dry to wet eolian depositional system, Burns formation, Meridiani Planum, Mars. *Earth and Planetary Science Letters*, *240*(1), 11–72.

Grotzinger, J. P., Sumner, D. Y., Kah, L., Stack, K., Gupta, S., Edgar, L., et al. (2014). A habitable fluvio-lacustrine environment at Yellowknife Bay, Gale Crater, Mars. *Science*, *343*(6169), 1242777. https://doi.org/10.1126/science.1242777

Hajek, E. A., Heller, P. L., & Sheets, B. A. (2010). Significance of channel-belt clustering in alluvial basins. *Geology*, *38*(6), 535–538.

Hajek, E. A., & Straub, K. M. (2017). Autogenic sedimentation in clastic stratigraphy. *Annual Review of Earth and Planetary Sciences*, *45*, 681–709.

Hajek, E. A., & Wolinsky, M. A. (2012). Simplified process modeling of river avulsion and alluvial architecture: Connecting models and field data. *Sedimentary Geology*, *257*, 1–30.

Hancock, G. S., & Anderson, R. S. (2002). Numerical modeling of fluvial strath-terrace formation in response to oscillating climate. *Geological Society of America Bulletin*, *114*(9), 1131–1142.

Hasbargen, L. E., & Paola, C. (2000). Landscape instability in an experimental drainage basin. *Geology*, *28*(12), 1067–1070.

Hovius, N. (1996). Regular spacing of drainage outlets from linear mountain belts. *Basin Research*, *8*(1), 29–44.

Howard, A. D., & Knutson, T. R. (1984). Sufficient conditions for river meandering: A simulation approach. *Water Resources Research*, *20*(11), 1659–1667.

Hoyal, D. C. J. D., & Sheets, B. A. (2009). Morphodynamic evolution of experimental cohesive deltas. *Journal of Geophysical Research*, *114*, F02009. https://doi.org/10.1029/2007JF000882

Hutton, J. (1788). X. Theory of the Earth; or an investigation of the laws observable in the composition, dissolution, and restoration of land upon the globe. *Earth and Environmental Science Transactions of the Royal Society of Edinburgh*, *1*(2), 209–304.

Jerolmack, D. J. (2009). Conceptual framework for assessing the response of delta channel networks to Holocene sea level rise. *Quaternary Science Reviews*, *28*(17-18), 1786–1800. https://doi.org/10.1016/j.quascirev.2009.02.015

Jerolmack, D. J., & Mohrig, D. (2005). A unified model for subaqueous bed form dynamics. *Water Resources Research*, *41*, W12421. https://doi.org/10.1029/2005wr004329

Jerolmack, D. J., & Paola, C. (2007). Complexity in a cellular model of river avulsion. *Geomorphology*, *91*, 259–270.

Jerolmack, D. J., & Paola, C. (2010). Shredding of environmental signals by sediment transport. *Geophysical Research Letters*, *37*, L19401. https://doi.org/10.1029/2010GL044638

Jerolmack, D. J., & Sadler, P. (2007). Transience and persistence in the depositional record of continental margins. *Journal of Geophysical Research*, *112*, F03S13. https://doi.org/10.1029/2006JF000555

Jerolmack, D. J., & Swenson, J. B. (2007). Scaling relationships and evolution of distributary networks on wave-influenced deltas. *Geophysical Research Letters*, *34*, L23402. https://doi.org/10.1029/2007GL031823

Johnson, S. Y. (1992). *Phanerozoic evolution of sedimentary basins in the Uinta-Piceance Basin region, northwestern Colorado and northeastern Utah*. US Geological Survey: US Department of the Interior.

Jonell, T. N., Owen, L. A., Carter, A., Schwenniger, J.-L., & Clift, P. D. (2018). Quantifying episodic erosion and transient storage on the western margin of the Tibetan Plateau, upper Indus River. *Quaternary Research*, *89*(1), 281–306.

Kemp, D. B., Fraser, W. T., & Izumi, K. (2018). Stratigraphic completeness and resolution in an ancient mudrock succession. *Sedimentology*, *65*(6), 1875–1890.

Kemp, D. B., & Sexton, P. F. (2014). Time-scale uncertainty of abrupt events in the geologic record arising from unsteady sedimentation. *Geology*, *42*(10), 891–894.

Kemp, D. B., & Van Manen, S. M. (2019). Metre-scale cycles in shallow water carbonate successions: Milankovitch and stochastic origins. *Sedimentology*.

Kennett, J. P., & Stott, L. (1991). Abrupt deep-sea warming, palaeoceanographic changes and benthic extinctions at the end of the Palaeocene. *Nature*, *353*(6341), 225.

Kim, W., Fouke, B. W., Petter, A. L., Quinn, T. M., Kerans, C., & Taylor, F. (2012). Sea-level rise, depth-dependent carbonate sedimentation and the paradox of drowned platforms. *Sedimentology*, *59*(6), 1677–1694.

Kim, W., & Paola, C. (2007). Long-period cyclic sedimentation with constant tectonic forcing in an experimental relay ramp. *Geology*, *35*(4), 331–334.

Kleinhans, M. G., van Dijk, W. M., van de Lageweg, W. I., Hoyal, D. C., Markies, H., van Maarseveen, M., et al. (2014). Quantifiable effectiveness of experimental scaling of river-and delta morphodynamics and stratigraphy. *Earth-Science Reviews*, *133*, 43–61.

Kleinhans, M. G., Van Scheltinga, R. T., Van Der Vegt, M., & Markies, H. (2015). Turning the tide: Growth and dynamics of a tidal basin and inlet in experiments. *Journal of Geophysical Research: Earth Surface*, *120*, 95–119.

Knight, J., & Harrison, S. (2014). Limitations of uniformitarianism in the Anthropocene. *Anthropocene*, *5*, 71–75.

Knox, J. C. (1996). Late Quaternary Upper Mississippi River alluvial episodes and their significance to the Lower Mississippi River system. *Engineering Geology*, *45*(1-4), 263–285.

Koch, P. L., Zachos, J. C., & Dettman, D. L. (1995). Stable isotope stratigraphy and paleoclimatology of the Paleogene Bighorn Basin (Wyoming, USA). *Palaeogeography, Palaeoclimatology, Palaeoecology*, *115*(1-4), 61–89.

Koch, P. L., Zachos, J. C., & Gingerich, P. D. (1992). Correlation between isotope records in marine and continental carbon reservoirs near the Palaeocene/Eocene boundary. *Nature*, *358*(6384), 319.

Kolmogorov, A. N. (1951). *Solution of a problem in probability theory connected with the problem of the mechanism of stratification*. New York: American Mathematical Society.

Kraus, M. J., Woody, D. T., Smith, J. J., & Dukic, V. (2015). Alluvial response to the Paleocene–Eocene Thermal Maximum climatic event, Polecat Bench, Wyoming (USA). *Palaeogeography, Palaeoclimatology, Palaeoecology*, *435*, 177–192.

Lamb, M. P., Nittrouer, J. A., Mohrig, D., & Shaw, J. (2012). Backwater and river plume controls on scour upstream of river mouths: Implications for fluvio-deltaic morphodynamics. *Journal of Geophysical Research*, *117*.

Lauzon, R., & Murray, A. B. (2018). Comparing the cohesive effects of mud and vegetation on delta evolution. *Geophysical Research Letters*, *45*, 10,437–410,445.

Leeder, M. R. (1978). A quantitative stratigraphic model for alluvium, with special reference to channel deposit density and interconnectedness. In A. D. Miall (Ed.), *Fluvial Sedimentology* (Vol. 5, 587–596). Can. Soc. Petrol. Geol. Mem..

Lentsch, N., Finotello, A., & Paola, C. (2018). Reduction of deltaic channel mobility by tidal action under rising relative sea level. *Geology*, *46*(7), 599–602.

Li, Q., Gasparini, N. M., & Straub, K. M. (2018). Some signals are not the same as they appear: How do erosional landscapes transform tectonic history into sediment flux records? *Geology*, *46*(5), 407–410.

Li, Q., Matthew Benson, W., Harlan, M., Robichaux, P., Sha, X., Xu, K., & Straub, K. M. (2017). Influence of sediment cohesion on deltaic morphodynamics and stratigraphy over basin-filling time scales. *Journal of Geophysical Research: Earth Surface*, *122*, 1808–1826.

Li, Q., & Straub, K. M. (2017). TDB_13_1, edited, SEAD. https://doi.org/10.5967/M07D2S7Q

Li, Q., Yu, L., & Straub, K. M. (2016). Storage thresholds for relative sea-level signals in the stratigraphic record. *Geology*, *44*(3), 179–182. https://doi.org/10.1130/G37484.1

Lyons, S. L., Baczynski, A. A., Babila, T. L., Bralower, T. J., Hajek, E. A., Kump, L. R., et al. (2019). Palaeocene–Eocene thermal maximum prolonged by fossil carbon oxidation. *Nature Geoscience*, *12*(1), 54.

Mahon, R. C., & McElroy, B. (2018). Indirect estimation of bedload flux from modern sand-bed rivers and ancient fluvial strata. *Geology*, *46*(7), 579–582.

Mahon, R. C., Shaw, J. B., Barnhart, K. R., Hobley, D. E., & McElroy, B. (2015). Quantifying the stratigraphic completeness of delta shoreline trajectories. *Journal of Geophysical Research: Earth Surface*, *120*, 799–817.

Malmon, D. V., Dunne, T., & Reneau, S. L. (2003). Stochastic theory of particle trajectories through alluvial valley floors. *The Journal of Geology*, *111*(5), 525–542.

Malverti, L., Lajeunesse, E., & Métivier, F. (2008). Small is beautifull: Upscaling microscale experimental results to the size of natural rivers. *Journal of Geophysical Research*, *113*, F04004.

Marr, J., Swenson, J., Paola, C., & Voller, V. (2000). A two-diffusion model of fluvial stratigraphy in closed depositional basins. *Basin Research*, *12*(3-4), 381–398.

Martin, J., Sheets, B., Paola, C., & Hoyal, D. (2009). Influence of steady base-level rise on channel mobility, shoreline migration, and scaling properties of a cohesive experimental delta. *Journal of Geophysical Research*, *114*, F03017. https://doi.org/10.1029/2008jf001142

McElroy, B., Willenbring, J., & Mohrig, D. (2018). Addressing time-scale–dependent erosion rates from measurement methods with censorship. *GSA Bulletin*, *130*(3-4), 381–395.

McInerney, F. A., & Wing, S. L. (2011). The Paleocene-Eocene thermal maximum: A perturbation of carbon cycle, climate, and biosphere with implications for the future. *Annual Review of Earth and Planetary Sciences*, *39*, 489–516.

Métivier, F. (1999). Diffusivelike buffering and saturation of large rivers. *Physical Review E*, *60*(5), 5827.

Métivier, F., & Gaudemer, Y. (1999). Stability of output fluxes of large rivers in South and East Asia during the last 2 million years: Implications on floodplain processes. *Basin Research*, *11*(4), 293–303.

Miall, A. D. (1977). A review of the braided-river depositional environment. *Earth-Science Reviews*, *13*(1), 1–62.

Miall, A. D. (2015). Updating uniformitarianism: Stratigraphy as just a set of 'frozen accidents'. *Geological Society, London, Special Publications*, *404*(1), 11–36.

Miller, K. L.Kim, W., & McElroy, B. 2019. Laboratory investigation on effects of flood intermittency on fan delta dynamics. *Journal of Geophysical Research: Earth Surface*, *124*, 383–399. https://doi.org/10.1029/2017JF004576

Miller, T. (1965). Time in stratigraphy. *Palaeontology*, *8*(1), 113–131.

Mohrig, D., Heller, P. L., Paola, C., & Lyons, W. J. (2000). Interpreting avulsion process from ancient alluvial sequences: Guadalope-Matarranya (northern Spain) and Wasatch Formation (western Colorado). *Geoogical Society of America Bulletin*, *112*, 1787–1803.

Muto, T., & Steel, R. J. (2004). Autogenic response of fluvial deltas to steady sea-level fall: implications from flume-tank experiments. *Geology*, *32*(5), 401–404.

Muto, T., Steel, R. J., & Swenson, J. B. (2007). Autostratigraphy: A framework norm for genetic stratigraphy. *Journal of Sedimentary Research*, *77*(1), 2–12.

Muto, T., & Swenson, J. B. (2005). Large-scale fluvial grade as a nonequilibrium state in linked depositional systems: Theory and experiment. *Journal of Geophysical Research*, *110*.

Muto, T., & Swenson, J. B. (2006). Autogenic attainment of large-scale alluvial grade with steady sea-level fall: An analog tank-flume experiment. *Geology*, *34*(3), 161–164.

Myrow, P. M., Jerolmack, D. J., & Perron, J. T. (2018). Bedform disequilibrium. *Journal of Sedimentary Research*, *88*(9), 1096–1113.

Nittrouer, J. A., Shaw, J., Lamb, M. P., & Mohrig, D. (2012). Spatial and temporal trends for water-flow velocity and bed-material sediment transport in the lower Mississippi River. *Geological Society of America Bulletin*, *124*(3-4), 400–414.

Paola, C. (2000). Quantitative models of sedimentary basin filling. *Sedimentology*, *47*, 121–178.

Paola, C. (2016). A mind of their own: Recent advances in autogenic dynamics in rivers and deltas. *SEPM Special Publication*, *106*, 5–17.
Paola, C., & Borgman, L. (1991). Reconstructing random topography from preserved stratification. *Sedimentology*, *38*, 553–565.
Paola, C., Foufoula-Georgiou, E., Dietrich, W. E., Hondzo, M., Mohrig, D., Parker, G., et al. (2006). Toward a unified science of the Earth's surface: Opportunities for synthesis among hydrology, geomorphology, geochemistry, and ecology. *Water Resources Research*, *42*, W03S10. https://doi.org/10.1029/2005wr004336
Paola, C., Ganti, V., Mohrig, D., Runkel, A. C., & Straub, K. M. (2018). Time not our time: Physical controls on the preservation and measurement of geologic time. *Annual Review of Earth and Planetary Sciences*, *46*, 409–438.
Paola, C., Heller, P. L., & Angevine, C. L. (1992). The large-scale dynamics of grain-size variation in alluvial basins. 1: Theory. *Basin Research*, *4*, 73–90.
Paola, C., & Mohrig, D. (1996). Paleohydraulics revisited: palaeoslope estimation in coarse-grained braided rivers. *Basin Research*, *8*, 243–254.
Paola, C., Parker, G., Mohrig, D. C., Whipple, K. X. (1999). The influence of transport fluctuations on spatially averaged topography on a sandy, braided fluvial fan.
Paola, C., Straub, K. M., Mohrig, D., & Reinhardt, L. (2009). The “unreasonable effectiveness” of stratigraphic and geomorphic experiments. *Earth-Science Reviews*, *97*, 1–43.
Paola, C., & Swenson, J. B. (1998). Geometric constraints on composition of sediment derived from erosional landscapes. *Basin Research*, *10*(1), 37–47.
Parker, G. (1978). Self-formed straight rivers with equilibrium banks and mobile bed. Part 1. *The sand-silt river, Journal of Fluid Mechanics*, *89*(1), 109–125.
Pelletier, J., & Turcotte, D. (1996). Scale-invariant topography and porosity variations in fluvial sedimentary basins. *Journal of Geophysical Research*, *101*(B12), 28,165–28,175.
Pelletier, J. D., & DeLong, S. (2004). Oscillations in arid alluvial-channel geometry. *Geology*, *32*(8), 713–716.
Pelletier, J. D., & Turcotte, D. L. (1997). Sythetic stratigraphy with a stochastic diffusion model of fluvial sedimentation. *Journal of Sedimentary Research*, *67*, 1060–1067.
Peters, S. E., & Loss, D. P. (2012). Storm and fair-weather wave base: A relevant distinction? *Geology*, *40*(6), 511–514.
Phillips, C. B., & Jerolmack, D. J. (2016). Self-organization of river channels as a critical filter on climate signals. *Science*, *352*(6286), 694–697. https://doi.org/10.1126/science.aad3348
Piliouras, A., Kim, W., & Carlson, B. (2017). Balancing aggradation and progradation on a vegetated delta: The importance of fluctuating discharge in depositional systems. *Journal of Geophysical Research: Earth Surface*, *122*, 1882–1900.
Pirmez, C., & Imran, J. (2003). Reconstruction of turbidity currents in Amazon Channel. *Marine and Petroleum Geology*, *20*, 823–849.
Pisel, J. R., Pyles, D. R., & Kirschbaum, M. A. (2018). The influence of lateral topographic confinement on fluvial channel-belt clustering, compensation and connectivity–lower Wasatch Formation and Dakota Sandstone, Utah, USA. *Sedimentology*, *65*(2), 597–619.
Pizzuto, J., Keeler, J., Skalak, K., & Karwan, D. (2017). Storage filters upland suspended sediment signals delivered from watersheds. *Geology*, *45*(2), 151–154.
Powell, E. J., Kim, W., & Muto, T. (2012). Varying discharge controls on timescales of autogenic storage and release processes in fluvio-deltaic environments: Tank experiments. *Journal of Geophysical Research*, *117*, F02011. https://doi.org/10.1029/2011jf002097
Purkis, S. J., van de Koppel, J., & Burgess, P. M. (2016). Spatial self-organization in carbonate depositional environments. In *SEPM Special Publication* (Vol. 106, pp. 53–66).
Ratliff, K. M., Hutton, E. H., & Murray, A. B. (2018). Exploring wave and sea-level rise effects on delta morphodynamics with a coupled river-ocean model. *Journal of Geophysical Research: Earth Surface*, *123*, 2887–2900.
Rodríguez-Iturbe, I., Rinaldo, A., Rigon, R., Bras, R. L., Marani, A., & Ijjász-Vásquez, E. (1992). Energy dissipation, runoff production, and the three-dimensional structure of river basins. *Water Resources Research*, *28*(4), 1095–1103.
Rodriguez-Iturbe, I., Rinaldo, A., Rigon, R., Bras, R. L., Marani, A., & Ijjasz-Vasquez, E. J. (1992). Energy dissipation, runoff production, and the three-dimentsional structure of river basins. *Water Resources Research*, *28*(4), 1095–1103.
Romans, B. W., Castelltort, S., Covault, J. A., Fildani, A., & Walsh, J. (2016). Environmental signal propagation in sedimentary systems across timescales. *Earth-Science Reviews*, *153*, 7–29.
Romans, B. W., Normark, W. R., McGann, M. M., Covault, J. A., & Graham, S. A. (2009). Coarse-grained sediment delivery and distribution in the Holocene Santa Monica Basin, California: Implications for evaluating source-to-sink flux at millennial time scales. *Geological Society of America Bulletin*, *121*(9-10), 1394–1408. https://doi.org/10.1130/B26393.1
Rubin, D. M., Hunter, R. E. (1983). Reconstructing bedform assemblages from compound crossbedding. In *Developments in Sedimentology* (Vol. 38, 407–427). Elsevier/.
Sadler, P. M. (1981). Sediment accumulation rates and the completeness of stratigraphic sections. *Journal of Geology*, *89*, 569–584.
Sadler, P. M., & Jerolmack, D. J. (2015). Scaling laws for aggradation, denudation and progradation rates: The case for time-scale invariance at sediment sources and sinks. *Geological Society, London, Special Publications*, *404*(1), 69–88.
Sadler, P. M., & Strauss, D. J. (1990). Estimation of completeness of stratigraphical sections using empirical data and theoretical-models. *Journal of the Geological Society of London*, *147*, 471–485.
Scheingross, J. S., Lamb, M. P., & Fuller, B. M. (2019). Self-formed bedrock waterfalls. *Nature*, *567*(7747), 229.
Schmitz, B., & Pujalte, V. (2007). Abrupt increase in seasonal extreme precipitation at the Paleocene-Eocene boundary. *Geology*, *35*(3), 215–218.
Schumer, R., & Jerolmack, D. J. (2009). Real and apparent changes in sediment deposition rates through time. *Journal of Geophysical Research*, *114*, F00a06. https://doi.org/10.1029/2009jf001266
Schumer, R., Jerolmack, D. J., & McElroy, B. (2011). The stratigraphic filter and bias in measurement of geologic rates. *Geophysical Research Letters*, *38*, L11405. https://doi.org/10.1029/2011gl047118
Schumm, S. (1973). Geomorphic thresholds and complex response of drainage systems. *Fluvial geomorphology*, *6*, 69–85.
Schumm, S. A. (1979). Geomorphic thresholds: The concept and its applications. *Transactions of the Institute of British Geographers*, *4*(4), 485–515.
Schumm, S. A., & Parker, R. S. (1973). Implications of complex response of drainage systems for Quaternary alluvial stratigraphy. *Nature Physical Science*, *243*(128), 99.
Sharman, G. R., Sylvester, Z., & Covault, J. A. (2019). Conversion of tectonic and climatic forcings into records of sediment supply and provenance. *Scientific Reports*, *9*(1), 4115. https://doi.org/10.1038/s41598-019-39754-6

Sheets, B. A., Hickson, T. A., & Paola, C. (2002). Assembling the stratigraphic record: Depositional patterns and time-scales in an experimental alluvial basin. *Basin Research*, *14*, 287–301.

Shen, Z., Tornqvist, T. E., Autin, W. J., Mateo, Z. R. P., Straub, K. M., & Mauz, B. (2012). Rapid and widespread response of the Lower Mississippi River to eustatic forcing during the last glacial-interglacial cycle. *Geological Society of America Bulletin*, *124*(5/6), 690–704.

Simpson, G., & Castelltort, S. (2012). Model shows that rivers transmit high-frequency climate cycles to the sedimentary record. *Geology*, *40*(12), 1131–1134. https://doi.org/10.1130/G33451.1

Slingerland, R., & Kump, L. (2011). *Mathematical modeling of Earth's dynamical systems: A primer*. Princeton University Press.

Slingerland, R., & Smith, N. D. (1998). Necessary conditions for a meandering-river avulsion. *Geology*, *26*(5), 435–438.

Sluijs, A., Brinkhuis, H., Schouten, S., Bohaty, S. M., John, C. M., Zachos, J. C., et al. (2007). Environmental precursors to rapid light carbon injection at the Palaeocene/Eocene boundary. *Nature*, *450*(7173), 1218–1221. https://doi.org/10.1038/nature06400

Snow, R. S., & Slingerland, R. L. (1990). Stream profile adjustment to crustal warping—Nonlinear results from a simple-model. *Journal of Geology*, *98*(5), 699–708.

Southard, J. B. (1991). Experimental determination of bed-form stability. *Annual Review of Earth and Planetary Sciences*, *19*(1), 423–455.

Straub, K. (2019). Morphodynamics and stratigraphic architecture of shelf-edge deltas subject to constant vs. dynamic environmental forcings: A laboratory study. *Frontiers in Earth Science*, *7*, 121.

Straub, K. M., & Esposito, C. R. (2013). Influence of water and sediment supply on the stratigraphic record of alluvial fans and deltas: Process controls on stratigraphic completeness. *Journal of Geophysical Research: Earth Surface*, *118*, 1–13. https://doi.org/10.1002/jgrf.20061

Straub, K. M., & Foreman, B. Z. (2018). Geomorphic stasis and spatiotemporal scales of stratigraphic completeness. *Geology*, *46*(4), 311–314.

Straub, K. M., Ganti, V., Paola, C., & Foufoula-Georgiou, E. (2012). Prevalence of exponential bed thickness distributions in the stratigraphic record: Experiments and theory. *Journal of Geophysical Research*, *117*, F02003. https://doi.org/10.1029/2011jf002034

Straub, K. M., Li, Q., & Benson, W. M. (2015). Influence of sediment cohesion on deltaic shoreline dynamics and bulk sediment retention: A laboratory study. *Geophysical Research Letters*, *42*, 9808–9815.

Straub, K. M., Paola, C., Mohrig, D., Wolinsky, M. A., & George, T. (2009). Compensational stacking of channelized sedimentary deposits. *Journal of Sedimentary Research*, *79*(9), 673–688.

Straub, K. M., & Pyles, D. R. (2012). Quantifying the hierarchical organization of compensation in submarine fans using surface statistics. *Journal of Sedimentary Research*, *82*(11-12), 889–898.

Straub, K. M., & Wang, Y. (2013). Influence of water and sediment supply on the long-term evolution of alluvial fans and deltas: Statistical characterization of basin-filling sedimentation patterns. *Journal of Geophysical Research: Earth Surface*, *118*, 1–15. https://doi.org/10.1002/jgrf.20095

Strauss, D., & Sadler, P. M. (1989). Stochastic-models for the completeness of stratigraphic sections. *Mathematical Geology*, *21*(1), 37–59.

Talling, P. J., Stewart, M. D., Stark, C. P., Gupta, S., & Vincent, S. J. (1997). Regular spacing of drainage outlets from linear fault blocks. *Basin Research*, *9*(4), 275–302.

Thomas, E. (2003). Extinction and food at the seafloor: A high-resolution benthic foraminiferal record across the Initial Eocene Thermal Maximum, Southern Ocean Site 690. *Special Papers-Geological Society of America*, *319-332*.

Thomas, E., & Monechi, S. (2007). Cenozoic mass extinctions in the deep sea: What perturbs the largest habitat on Earth? *SPECIAL PAPERS-GEOLOGICAL SOCIETY OF AMERICA*, *424*, 1.

Tipper, J. C. (1983). Rates of sedimentation, and stratigraphical completeness. *Nature*, *302*(5910), 696–698.

Tipper, J. C. (2015). The importance of doing nothing: Stasis in sedimentation systems and its stratigraphic effects. *Geological Society, London, Special Publications*, *404*(1), 105–122.

Toby, S. C., Duller, R. A., De Angelis, S., & Straub, K. M. (2019). *A stratigraphic framework for the preservation and shredding of environmental signals*. Geophysical Research: Letters.

Tofelde, S., Savi, S., Wickert, A. D., Bufe, A., & Schildgen, T. F. (2019). Alluvial channel response to environmental perturbations: Fill-terrace formation and sediment-signal disruption. *Earth Surface Dynamics*, *7*(2), 609–631.

Trabucho-Alexandre, J. (2015). More gaps than shale: Erosion of mud and its effect on preserved geochemical and palaeobiological signals. *Strata and time: Probing the gaps in our understanding: Geological society, London, special publications*, *404*, 251–270.

Trampush, S., Hajek, E., Straub, K., & Chamberlin, E. (2017). Identifying autogenic sedimentation in fluvial-deltaic stratigraphy: Evaluating the effect of outcrop-quality data on the compensation statistic. *Journal of Geophysical Research: Earth Surface*, *122*, 1–23.

Trampush, S. M., & Hajek, E. A. (2017). Preserving proxy records in dynamic landscapes: Modeling and examples from the Paleocene-Eocene Thermal Maximum. *Geology*, *45*(11), 967–970.

Trower, E. J., Ganti, V., Fischer, W. W., & Lamb, M. P. (2018). Erosional surfaces in the Upper Cretaceous Castlegate Sandstone (Utah, USA): Sequence boundaries or autogenic scour from backwater hydrodynamics? *Geology*, *46*(8), 707–710.

Tucker, G. E., & Bras, R. L. (1998). Hillslope processes, drainage density, and landscape morphology. *Water Resources Research*, *34*(10), 2751–2764.

Turner, S. K. (2018). Constraints on the onset duration of the Paleocene–Eocene Thermal Maximum, *Philosophical Transactions of the Royal Society A: Mathematical. Physical and Engineering Sciences*, *376*(2130), 20170082.

Vail, P. R., Mitchum, R. M. Jr., & Thompson, S. III (1977). Seismic stratigraphy and global changes of sea level, Part 3: Relative changes of sealevel from coastal onlap. In C. E. Payton (Ed.), *Seismic Stratigraphy—Applications to Hydrocarbon Exploration, Memoir* (Vol. 26, 63–81). Tulsa, OK: American Association of Petroleum Geologists.

Van De Wiel, M. J., & Coulthard, T. J. (2010). Self-organized criticality in river basins: Challenging sedimentary records of environmental change. *Geology*, *38*(1), 87–90.

Van Wagoner, J., Posamentier, H., Mitchum, R., Vail, P., Sarg, J., Loutit, T., & Hardenbol, J. (1988), An overview of the fundamentals of sequence stratigraphy and key definitions.

Van Wagoner, J. C. (1995). Sequence stratigraphy and marine to nonmarine facies architecture of foreland basin strata, Book Cliffs, Utah, U.S.A. In J. C. Van Wagoner & G. T. Bertram(Eds.), *Sequence Stratigraphy of Foreland Basin Deposits–Outcrop and Subsurface Examples from the Cretaceous of North America, American Association of Petroleum Geologists, Memoir* (Vol. 64, 137–223).

Van Wagoner, J. C., Mitchum, R. M, Campion, K. M., & Rahmanian, V. D. (1990). Siliciclastic sequence stratigraphy in well logs, cores, and outcrops*: Concepts for high-resolution correlation of time and facies: American association of petroleum geologists*, *methods in exploration 7* (55 pp.).

Vázquez, G., Roy, P. D., Solis, B., Blanco, E., & Lozano-Santacruz, R. (2017). Holocene paleohydrology of the Etzatlán-Magdalena basin in western-central Mexico and evaluation of main atmospheric forcings. *Palaeogeography, Palaeoclimatology, Palaeoecology*, *487*, 149–157.

Vendettuoli, D., Clare, M., Clarke, J. H., Vellinga, A., Hizzet, J., Hage, S., et al. (2019). Daily bathymetric surveys document how stratigraphy is built and its extreme incompleteness in submarine channels. *Earth and Planetary Science Letters*, *515*, 231–247.
Ventra, D., & Nichols, G. J. (2014). Autogenic dynamics of alluvial fans in endorheic basins: outcrop examples and stratigraphic significance. *Sedimentology*, *61*(3), 767–791.
Villmoare, B., Kimbel, W. H., Seyoum, C., Campisano, C. J., DiMaggio, E. N., Rowan, J., et al. (2015). Early Homo at 2.8. *Ma from Ledi-Geraru, Afar, Ethiopia, Science*, *347*(6228), 1352–1355.
Walling, D. E. (1983). The sediment delivery problem. *Journal of Hydrology*, *65*(1-3), 209–237.
Wang, Y., & Straub, K. M. (2017). TDB_10_1, edited, SEAD. https://doi.org/10.5967/M0HX19TT
Wang, Y., Straub, K. M., & Hajek, E. A. (2011). Scale-dependent compensational stacking: An estimate of autogenic time scales in channelized sedimentary deposits. *Geology*, *39*(9), 811–814.
Watkins, S. E., Whittaker, A. C., Bell, R. E., McNeill, L. C., Gawthorpe, R. L., Brooke, S. A., & Nixon, C. W. (2018). Are landscapes buffered to high-frequency climate change? A comparison of sediment fluxes and depositional volumes in the Corinth Rift, central Greece, over the past 130 ky. *Bulletin*, *131*(3-4), 372–388.
Weltje, G. J., & von Eynatten, H. (2004). Quantitative provenance analysis of sediments: review and outlook. *Sedimentary Geology*, *171*(1-4), 1–11.
Whipple, K. X. (2001). Fluvial landscape response time: How plausible is steady-state denudation? *American Journal of Science*, *301*(4-5), 313–325.
Whipple, K. X., & Trayler, C. R. (1996). Tectonic control of fan size: the importance of spatially variable subsidence rates. *Basin Research*, *8*(3), 351–366.
Whittaker, A. C., & Boulton, S. J. (2012). Tectonic and climatic controls on knickpoint retreat rates and landscape response times. *Journal of Geophysical Research*, *117*, F02024. https://doi.org/10.1029/2011jf002157
Whittaker, A. C., Cowie, P. A., Attal, M., Tucker, G. E., & Roberts, G. P. (2007). Bedrock channel adjustment to tectonic forcing: Implications for predicting river incision rates. *Geology*, *35*(2), 103–106.
Wickert, A. D., Martin, J. M., Tal, M., Kim, W., Sheets, B., & Paola, C. (2013). River channel lateral mobility: Metrics, time scales, and controls. *Journal of Geophysical Research: Earth Surface*, *118*, 396–412. https://doi.org/10.1029/2012JF002386
Wickert, A. D., & Schildgen, T. F. (2019). Long-profile evolution of transport-limited gravel-bed rivers.
Wing, S. L., Harrington, G. J., Smith, F. A., Bloch, J. I., Boyer, D. M., & Freeman, K. H. (2005). Transient floral change and rapid global warming at the Paleocene-Eocene boundary. *Science*, *310*(5750), 993–996. https://doi.org/10.1126/science.1116913
Xie, X., & Heller, P. L. (2009). Plate tectonics and basin subsidence history. *Geological Society of America Bulletin*, *121*(1-2), 55–64.
Xu, K., Mickey, R. C., Chen, Q., Harris, C. K., Hetland, R. D., Hu, K., & Wang, J. (2016). Shelf sediment transport during hurricanes Katrina and Rita. *Computers & Geosciences*, *90*, 24–39.
Yu, L., Li, Q., & Straub, K. M. (2017). Scaling the response of deltas to relative-sea-level cycles by autogenic space and time scales: A laboratory study. *Journal of Sedimentary Research*, *87*(8), 817–837.
Zachos, J., Pagani, M., Sloan, L., Thomas, E., & Billups, K. (2001). Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science*, *292*(5517), 686–693. https://doi.org/10.1126/science.1059412
Zachos, J. C., Schouten, S., Bohaty, S., Quattlebaum, T., Sluijs, A., Brinkhuis, H., et al. (2006). Extreme warming of mid-latitude coastal ocean during the Paleocene-Eocene Thermal Maximum: Inferences from TEX86 and isotope data. *Geology*, *34*(9), 737–740.
Zachos, J. C., Wara, M. W., Bohaty, S., Delaney, M. L., Petrizzo, M. R., Brill, A., et al. (2003). A transient rise in tropical sea surface temperature during the Paleocene-Eocene thermal maximum. *Science*, *302*(5650), 1551–1554.
Zeebe, R. E., Ridgwell, A., & Zachos, J. C. (2016). Anthropogenic carbon release rate unprecedented during the past 66 million years. *Nature Geoscience*, *9*(4), 325.