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ENVIRONMENTAL SIGNAL PROPAGATION IN SEDIMENTARY SYSTEMS ACROSS TIMESCALES

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Abstract

Earth-surface processes operate across erosionally dominated landscapes and deliver sediment to depositional systems that can be preserved over a range of timescales. The geomorphic and stratigraphic products of this source-to-sink sediment transfer record signals of external environmental forcings, as well as internal, or autogenic, dynamics of the sedimentary system. Here, we evaluate environmental signal propagation across sediment-routing systems with emphasis on sediment supply, Qs, as the carrier of up-system forcings. We review experimental, numerical, and natural examples of source-to-sink sediment routing and signal propagation during three timescales: (1) Historic, which includes measurement and monitoring of events and processes of landscape change and deposition during decades to centuries; (2) Centuries to several millions of years, referred to as intermediate timescale; and (3) Deep time. We discuss issues related to autogenic

dynamics of sediment transport, transient storage, and release that can introduce noise, lags, and/or completely mask signals of external environmental forcings. We provide a set of conceptual and practical tools for evaluating sediment supply within a source-to-sink context, which can inform interpretations of signals from the sedimentary record. These tools include stratigraphic and sediment-routing system characterization, sediment budget determination, geochronology, detrital mineral analysis (e.g., thermochronology), comparative analog approaches, and modeling techniques to measure, calculate, or estimate the magnitude and frequency of external forcings compared to the characteristic response time of the sediment-routing systems.

1 Introduction

1.1 What is an 'Environmental Signal'?

From the perspective of sedimentary system analysis, signals are changes in sediment production, transport, or deposition, that originate from perturbations of environmental variables such as precipitation, sea level, rock uplift, subsidence, and human modifications. The origin of the perturbations can be 'natural' when they relate to tectonic and climatic processes that have happened over the course of Earth's history, or 'anthropogenic' if they are linked with human actions. Environmental signals occur over many temporal scales, ranging from several hours to millions of years in response to tectonic and climate changes. Signals involve a large range of spatial scales such as localized precipitation affecting small catchments to eustatic sea-level change that affects the globe.

An environmental signal can trigger a response of the Earth's surface in the form of erosion, sediment transport, and deposition, and the surface response may be local initially and further afield eventually as it propagates away. A sea-level fall, for example, can create local incision and shoreline regression, but also up-system knickpoint migration and down-system deposition in the deep sea. Similarly, an increase in precipitation can create a wave of incision, alluvial aggradation, and eventually a pulse of sediment discharge to the ocean. The overarching challenge of geomorphology and stratigraphy is to invert the history of environmental signals from landscape and rock records.

The transfer, or propagation, of signals is generally examined in the down-system direction, as this is the dominant direction of mass transfer (e.g., Castelltort and Van Den Driessche, 2003; Allen, 2008a, Jerolmack and Paola, 2010). However, up-system signal propagation driven by base level change has long been considered in the interpretation of the sedimentary record (e.g., Fisk, 1944), is important for distributive systems (e.g., backwater effect in deltas, Lamb et al., 2012), and is the subject of theoretical work (Voller et al., 2012).

Environmental signals are potentially preserved in the geomorphic expression of landscapes around us, as well as in the stratigraphic record of depositional basins. This review examines how signals propagate within the context of sediment routing systems with emphasis on the nature of sediment supply, or Qs, as the indicator of up-system forcings (Fig. 1A) (Allen et al., 2013). We think that reconstructing the rates and magnitudes of signal-generating processes from stratigraphy requires consideration of the nature of system response, and the potential modification of the original signal. It is also

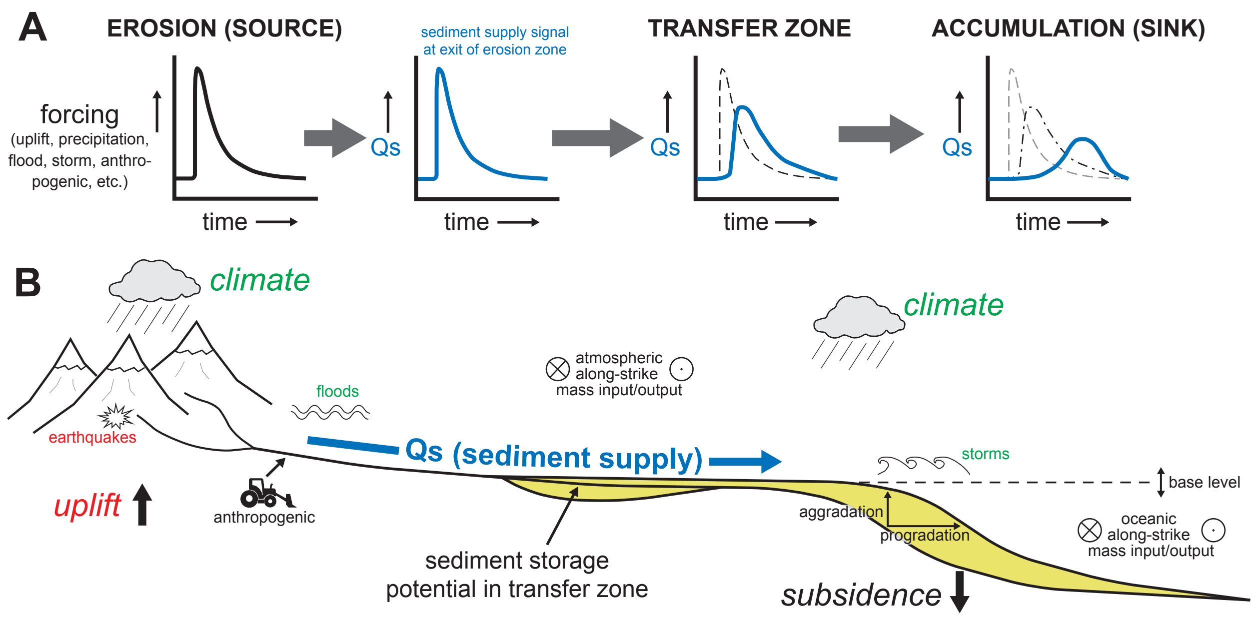

Figure 1: (A) Schematic portrayal of a sediment supply (Qs) signal from the erosion zone and how that signal propagates through the system. The leftmost Qs signal represents as measured at the exit of the erosion zone and for simplicity is the same as the original forcing of interest. The transfer zone Qs signal is measured within the transfer zone at some distance from exit of erosion zone and the rightmost signal represents that which reaches the accumulation zone and is an input for the stratigraphic record. Dashed lines refer to Qs signal in up-system segment(s) to illustrate that a signal can be modified during propagation. (B) 2-D profile of a generic sediment-routing system emphasizing erosion, transfer, and accumulation zones (potential for intermediate to deep time stratigraphic preservation in yellow) and important controls of tectonics (including earthquakes), climate (including storms), base level, and anthropogenic factors. Part B modified from Castelltort and Van Den Driessche (2003).

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important to recognize that signals can be masked or significantly altered by what can be referred to as 'noise.' In the present context, 'noise' has the broad meaning of any perturbation of the primary signal of interest, irrespective of its origin, frequency, or magnitude. It is one fundamental goal of stratigraphy to disentangle signal from noise, but what can be considered noise at one timescale may represent a signal at another. One notable type of noise is the result of internal, self-organized, dynamics of a sediment routing system (e.g., Jerolmack and Paola, 2010), that can potentially 'shred' environmental signals as a result of their large magnitude and period relative to the primary signal of interest (e.g., Jerolmack and Paola, 2010; Wang et al., 2011).

Deciphering signals has obvious implications for the meaning of the sedimentary record of Earth history: what do sediments and rocks tell us about the past? However, understanding the signal-to-noise character of the sedimentary record is also relevant to the prediction of land-to-sea export and burial of terrestrial organic carbon (e.g., Kao et al., 2014; Leithold et al., this volume), landscape resiliency and hazard management (e.g., Anthony and Julian, 1999), prediction of depositional systems for natural resource exploration and production (Bhattacharya et al., this volume), and response of hydrological systems to global climate change (e.g., Syvitski, 2003). We do not attempt to solve all the outstanding issues related to signal propagation and preservation in this contribution. Our goal is to provide the general Earth scientist a thorough review of the interesting and enigmatic questions and to promote a broader understanding that might attract other researchers to this multidisciplinary field of study.

1.2 Importance of Timescale of Investigation

We emphasize the importance of timescale in this review because of its association with the processes of signal generation, propagation, preservation, and analysis. The evaluation of signals requires consideration of the timescale(s) particularly in the context of internally generated 'noise.' Also, some signals occur over long durations (e.g., uplift and exhumation of a mountain belt) and, therefore, require a correspondingly long record from which to deduce the signal.

How do we put historical (past few centuries) measurements and observations within the framework of landscapes and stratigraphy constructed over timescales $\geq 10^3$ yr? Put another way, how do we accurately estimate short-term rates from geologic archives that have longer-term temporal resolution? For example, the < centennial stratigraphic record contains information about short-lived events, such as hurricane deposits, which can be reliably dated. The challenge is to extract meaningful insight from such records in the deeper past.

We organize this review of signal propagation and preservation within the context of three important timescales that span a minimum of seven orders of temporal magnitude: (1) Historic, which includes measurement and monitoring of events and processes of landscape change ($<10^2$ yr); (2) Centennial to several million years, herein referred to as the intermediate timescale (10^2-10^6 yr); and (3) Deep time ($\geq 10^7$ yr) (Fig. 2). These timescales are discussed in terms of age of the system as well as duration or period of forcing. The timescale of investigation also influences the application of concepts of steady state, response time, and other system dynamics indicators, which will be discussed in detail in the intermediate timescale Section 3.

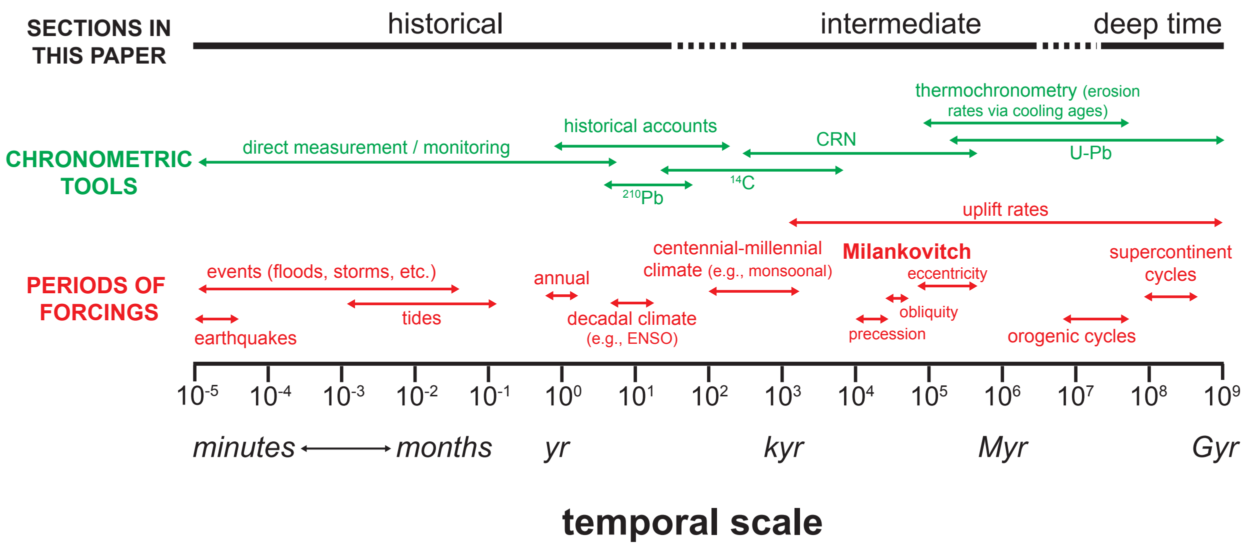

Figure 2: Overview of three timescales of investigation, some of the chronometric tools with which to constrain process rates, and periods of some of the forcings discussed in this review. Dashed lines at the top emphasize the continuum among the timescales. Temporal range of 'orogenic cycles' from DeCelles et al. (2009). Effective dating range of chronometric tools from Walker (2005).

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1.3 Sediment Routing Systems

Earth-surface processes operate within erosionally dominated landscapes coupled with depositional systems that can be preserved over a range of timescales. A simple and elegant way to consider an integrated sedimentary system was presented by Schumm (1977) wherein he subdivided a system into three spatial zones of dominant mass-flux behavior: denudation/erosion, transfer, and accumulation/deposition. Similar to Castelltort and Van Den Driessche (2003) and Sadler and Jerolmack (2014), we depict a generic sediment routing system in cross section and denote the prominent environmental forcings of interest in this review (Fig. 1B). The 'transfer zone' is assumed to be the segment of the sedimentary system that is neither net-denudational nor net-accumulative; rather, it is characterized by the balance between sediment removal/remobilization and sediment storage that feeds or starves down-system accumulation zones. Thus, this zone typically does not produce much sediment via bedrock erosion and, over sufficiently long timescales, it will transfer more mass than it produces or accumulates. We consider the morphology and process history of the transfer zone as an indicator of system response to perturbations, which is important for reconstructing paleo-sediment routing systems. A spatial scale is not shown on Figure 1B because the lengths of these zones vary significantly from system to system (e.g., Somme et al., 2009). For example, small/high-relief sediment routing systems (10-50 km long) typically have very short transfer zones, which results in negligible transient sediment storage, whereas large, continental-scale systems (100-1000 km long) commonly have long transfer zones containing sediment sinks that can store sediment temporarily or permanently given favorable subsidence conditions. The magnitudes and timescales of such mass transfer-and-storage behavior, which can be addressed through the estimation of sediment budgets, are fundamental to the propagation of signals. We focus on the down-system mass transfer of inorganic, dominantly siliciclastic, particulates through water-sediment flows and refer the reader to Leithold et al. (this volume) for a review of organic-carbon dynamics of source-to-sink systems. Additionally, we acknowledge the important and unique sediment-supply characteristics of glaciated systems but do not distinguish them here and refer the reader to Jaeger and Koppes (this volume).

2 Sedimentary Process-Response Over Historical (<10^2 yr) Timescales

Signals at the historical timescale are the result of individual events that last hours to days (e.g., floods, storms, and earthquakes) to longer-lived changes that occur over decades (e.g., watershed deforestation and other land-use alterations) (Fig. 1). The mechanisms involved in the formation and/or propagation of such signals from source to sink include a range of hillslope (e.g., sheetwash, landsliding), glacial, fluvial, volcanic, oceanic (e.g., tides and storm wave) processes and subaqueous mass movements and flows (e.g., turbidity currents). Data from instruments have provided opportunities to measure and quantify sedimentary dynamics, and the stratigraphic record is also examined to link process to product over longer time. The timescale of this section covers what some consider to be the period of significant anthropogenic influence on Earth surface systems (onset of the Industrial Revolution, or ~250 yr before present; Crutzen and Stoermer, 2000; Zalasiewicz et al., 2000).

We identify four potential challenges to leveraging historic records to understand millennial-scale and deeper-time geology. First, ancient events might have been non-actualistic; i.e., there is no adequate modern analog regarding process (Myrow and Southard, 1996). For example, globally distributed strata that were produced by the Cretaceous bolide impact (e.g., Bralower et al., 1998). Second, although a recent event may have had profound impact on society (e.g., 2005 Hurricane Katrina, 2011 Mississippi River flood), the geological record produced might be negligible or non-existent, depending on many factors including spatial variation in supply and erosion (Turner et al., 2006; Walsh et al., 2006; Goni et al., 2007; McKee and Cherry, 2009; Reed et al., 2009; Allison et al., 2010; Falcini et al., 2012; Kolker et al, 2014; Xu et al., 2014b). Third, the observation of modern sedimentary processes shows that strata are often destroyed within years after deposition as a result of physical and/or biogenic reworking (Wheatcroft et al., 2007 and references therein). Finally, there is the problem of discontinuous sedimentation and the likelihood of larger gaps in the record (i.e., time recorded as hiatus) as the time interval of sampling increases (Sadler, 1981; Sadler and Jerolmack, 2014), which will be discussed further in Section 4.

Erosional landforms provide a rich record of signals in the annual-to-centennial temporal range (e.g., Viles and Goudie, 2003), but the primary goal of our discussion is to understand signal propagation into the sedimentary record. Many studies at historical timescales are focused on specific processes and segments (e.g., hillslope erosion, shelf sedimentation) and do not strive to directly link source and sink through contemporaneous research. Moreover, simple relationships between event size (e.g., flood magnitude) and strata thickness may be the exception rather than the norm (Corbett et al., 2014 and references therein). As a result, the source-to-sink stratigraphic connection remains a challenge in many studies despite a wealth of data.

At the shortest end of the signal transfer spectrum (<1 yr), the potential for direct communication of a sediment supply signal to a sink is greatly limited. To produce a measurable signal in the stratigraphic record of the sink, events that drive sediment redistribution must move a relatively large volume of material over a short time. To understand modern system behavior, we recommend consideration of the source signal relative to the sink size; e.g., volume of event-scale Qs versus volume capacity of a sink. Furthermore, system size can impact the timescale of the signal. For example, a flood or earthquake-driven landslide into a confined mountain lake can be captured quickly (hours to days) and potentially with little post-depositional physical and/or biogenic modification (Schilleref et al., 2014) compared to a flood of the vast Mississippi River catchment into the Gulf of Mexico, the effects of which can persist for months (Allison et al.; 2000; Kolker et al., 2014; Xu et al., 2014b). Resolving events occurring in close succession is challenging because the signals might be truncated, overprinted, or commingled (e.g., hurricanes Katrina and Rita; Goni et al., 2007 or the Morokot earthquake and ensuing flood; Carter et al., 2012).

We first discuss key processes and rates of sediment production and transfer over human timescales. We then address the storage in sedimentary sinks and high-resolution dating typical of historical timescales. Finally, we examine two well-studied modern source-to-sink systems and the specificities of stratal preservation and sediment budgets over centennial timescales.

2.1 Sediment Production and Transfer Over Historical Timescales

Sediment production and movement in catchments and river systems is often described in a time-averaged perspective with the timescale of focus related to the measurement tool employed. Annual hillslope erosion rates (in mm/yr or t/ha/yr), sediment loads (t/yr) and yields (t/km^2/yr) may be used to compare and contrast systems and help evaluate their overall functioning (Milliman and Syvitski, 1992; Walling and Webb, 1996; Walling, 1999; Syvitski and Saito, 2007; Syvitski and Milliman, 2007; Milliman and Farnsworth; 2011; Covault et al., 2013). Loads and yields are commonly measured with stream gauges (e.g., Milliman and Farnsworth, 2011), which can be used to evaluate catchment erosion rates and/or alluvial storage (e.g., Meade et al., 1990; Walling and Collins, 2008). There are significant challenges to quantifying sediment transfer to the sea by rivers, particularly of large systems because of tidal influence on transport calculations and sediment storage in the lower river (e.g., Milliman et al., 1984; Allison et al., 2012). Historical measurements can be biased as a result of their limited duration or influences of anthropogenic catchment modification, including construction of dams and other land-use activities associated with agriculture, construction, and mining (Wilkinson and McElroy 2007; Milliman and Farnsworth, 2011). Erosion rates also can be measured with cosmogenically derived tracers (e.g., ^{10}Be; discussed further in Section 3) and radiochemically dated deposits (e.g., ^{14}C, ^{137}Cs or ^{210}Pb) from well-defined source areas (e.g., Walling and Collins, 2008). Technological advancements, specifically Light-Detection and Ranging and terrestrial laser scanners, have improved our ability to quantify morphological changes on land. Denudation rates from LiDAR, discharge measurements and ^{10}Be indicate variability depending on slope and other factors (commonly <0.5 mm/yr, but locally >3 mm/yr) (e.g., Hovius et al., 1997; Aalto et al., 2003; Roering et al., 2007; Korup et al., 2014). The contextual and temporal knowledge of precipitation and catchment characteristics usually exceeds what can be measured or inferred in ancient systems, as will be discussed in subsequent sections.

Water-driven transport, especially during intense floods, can generate recognizable sedimentary signals in sink areas. Intense rainfall and associated floods can rapidly (hours to days) move large volumes (>5 Mt [million metric tons]) of sediment through small (<5,000 km^2) catchments to offshore depositional areas (e.g., Sommerfield et al., 1999; Hale et al., 2014; Kniskern et al., 2014), and larger catchments (>50,000 km^2) can generate appreciable sediment supply (>10s Mt) signals to the sea over the course of days to weeks (e.g., Palinkas et al., 2005; Kolker et al., 2014). Subaerial and submarine landsliding and other mass movements are related to pre-conditioning factors, such as hillslope soil or rock strength, geomorphology, and short-term conditions (e.g., earthquake and hydrology) (Dietrich et al., 1995; Roering et al., 2007; Strasser et al., 2006; Goldfinger et al., 2012 and references therein).

An earthquake can disturb a catchment by increasing pore pressures and liquefying substrate, among other processes of manipulating gravitational loads on slopes, which can lead to abrupt increases in sediment loads (e.g., Dadson et al., 2004). Also, earthquake-triggered mass wasting can create conspicuous stratigraphic records in lakes and the deep sea (e.g., Heezen and Ewing, 1952; Piper and Aksu, 1997; Moernaut et al., 2007). Much research has explored coastal and marine sedimentary records to evaluate the recurrence interval for earthquakes and associated tsunamis (e.g., Atwater and Hemphill-Haley, 1997; Goldfinger et al., 2003; Strasser et al. 2006; Mournaut et al., 2007; Goldfinger et al., 2012; Barnes et al., 2013), including some recent detailed research focused on the Sumatra and Tomoko events (e.g., Szczucinski et al., 2012; Patton et al., 2013). There is still vigorous debate regarding deep-sea turbidite deposits as a reliable paleo-seismometer (e.g., Sumner et al., 2013; Atwater et al., 2014).

Over annual to centennial timescales, anthropogenic activities, such as deforestation and pollution, can create signals that become stored in sedimentary sinks (e.g., Paull et al., 2002; Cundy et al., 2003). Many natural and human factors (e.g., land use, dams) have significant influence on sediment yields and loads (Meade et al., 1990; Syvitski et al., 2005; Milliman and Farnsworth, 2011). As a consequence of the potential influence of human activities, Syvitski and Milliman (2007) included an anthropogenic factor in their BQART model that predicts global sediment flux to the oceans. Although intra-system storage might buffer some signals (i.e., low sediment delivery ratios; Phillips, 1991; Walling and Collins; 2008), catchment changes can notably increase Qs. Damming and leveeing can significantly diminish sediment supply into sink areas (Syvitski et al., 2005; Milliman and Farnsworth, 2011 and references therein), not only precluding new strata development but also yielding land loss in some areas (e.g., Day et al., 2007; Smith and Abdel-Kader, 1988).

2.2 Storage in Sedimentary Sinks Over Historical Timescales

To evaluate the presence of signals, including events, in stratigraphic records over historic timescales, ^{210}Pb and ^{137}Cs are commonly used to date deposits or determine sediment accumulation rates (Fig. 2) (e.g., Sommerfield and Nittrouer, 1999). Bathymetric and sub-bottom observations (i.e., seismic reflection) have revealed the geomorphic and stratigraphic complexity of subaqueous environments and such data are helpful to strategically position coring sites to obtain desired records (e.g., Goldfinger et al., 2012) or to inform spatial variability for determining sediment budgets (e.g., Miller and Kuehl, 2010; Gerber et al., 2010). Recent studies have shown how time-series bathymetric analysis with multibeam may yield new insight into the intermittent nature of fluvial deposition (Nittrouer et al., 2008) and subaqueous sediment density flows (e.g., Smith et al., 2005; Walsh et al., 2006; Paull et al., 2006; Xu et al., 2008; Hughes Clarke et al., 2012). Additionally, researchers are using innovative methods to track sediment transport and deposition, such as short-lived radiochemical tracers (i.e., ^{7}Be) for catchment and seaward sediment dispersal (e.g., Sommerfield et al., 1999; Dail et al., 2007; Walling, 2013), and mounted acoustic- and light-based sensors for measuring water and sediment movement (e.g., Xu et al., 2004; Dinehart and Burau, 2005; Cacchione et al., 2006). Deployed systems have provided flow measurements (i.e., velocity and sediment concentrations), which are essential to modeling sediment transport (e.g., Traykovski et al., 2007; Moriarty et al., 2014). However, field measurements remain limited especially during extreme and/or rare events when most sediment is moved (e.g., Ogston et al., 2000; Talling et al., 2013; Hale et al., 2014; Xu et al., 2014a; Stevens et al., 2014).

Sedimentary filling of hollows, ponds, lakes, floodplains, estuaries, and even sinkholes can provide information about individual events or decadal-to-centennial changes in the environment. Evidence for upstream changes includes increased sedimentation rates elevated trace metals, variations in pollen, microfossil organisms and/or assemblages, trace metals, and organic compounds (e.g., estuaries: Brush et al., 2001; Cooper et al., 2004; lakes: Noren et al., 2002, Girardclos et al., 2007; floodplains: Aalto et al., 2003; coastal deposits: Sorrel et al., 2012; Lane et al., 2011; shelves: Allison et al., 2012; deep sea: Soutar and Crill, 1977). Gilli et al. (2013) and Schillereff et al. (2014) provide reviews of flooding and climate changes from lake records.

Continental shelves, slopes, and deeper ocean segments are typically viewed as the ultimate depositional sinks, but their records are variably preserved as a result of post-depositional reworking and can be challenging to unravel (Nittrouer et al., 2007). Theory and modeling emphasize that event-layer preservation is a function of the rate of bioturbation, mixing depth, and layer thickness (Wheatcroft et al., 2007 and references therein). However, time-series coring studies of flood-related deposition on continental shelves offshore the Eel, Po, and Waipaoa sediment-routing systems have shown deep (>5 cm) biological reworking over the span of a few years (Wheatcroft et al., 2007; Tesi et al., 2012; Walsh et al., 2014). Areas of rapid sedimentation and physically reworked areas such as topset and foreset regions of clinoforms might have physical stratification preserved at depth, e.g., Amazon delta front (Kuehl et al., 1996; Sommerfield et al., 1999; Walsh et al., 2004; Rose and Kuehl, 2010). However, the presence of discontinuous, heterolithic bedforms can preclude recognition of event-specific beds (Goff et al., 2002; Walsh et al., 2014). Ocean areas with low or no dissolved oxygen inhospitable to benthic organisms (e.g., Soutar and Crill, 1977) are favorable for signal preservation (Allison et al., 2012). Continental margins and basin-margin deep-sea fans capture event records beyond historical timescales. However, during the sea-level highstand of the past several thousand years, off-shelf sediment transport is reduced in some settings (Posamentier and Vail, 1988; Covault and Graham, 2010), with shelf width serving as an important control

(Posamentier et al., 1991; Walsh and Nittrouer, 2003). As a result, limited sediment supply to some deep-sea fans has resulted in condensed sections recording few if any events at historical timescales.

2.3 Modern Sediment Routing System Examples

To further discuss historical ($<10^2$ yr) signal propagation, two differently sized sediment routing systems will be briefly discussed: the Eel River and the Ganges-Brahmaputra-Bengal system. The Eel is a small mountainous river system ($<10^3$ km^2) draining northern California, USA that has received intense scrutiny during and since the Office of Naval Research STRATA FORmation on Margins program (STRATAFORM; 1995-2004; Nittrouer et al., 2007). Small mountainous rivers are important for understanding sediment flux to the sea because of the minimal onshore sediment storage (Milliman and Syvitski, 1995; Kuehl et al., 2003; Covault et al., 2011). We contrast this work with the much larger Ganges-Brahmaputra-Bengal sediment-routing system, where abundant sediment is stored onshore, on the shelf, and in the canyon today (Kuehl et al., 2005; Walsh et al., 2013). Collectively, large systems provided potentially a third to a half of the sediment to the sea prior to human alterations (Milliman and Farnsworth, 2011; Walsh et al., 2013).

The Eel River is one of the most comprehensively studied modern sediment routing system over the historical timescale (<500 yr). Its 9,400 km^2 catchment in a tectonically active setting of outcropping sedimentary rocks is estimated to discharge ~12-16 Mt of sediment to the sea annually (Sommerfield and Nittrouer, 1999; Warrick, 2014; Sommerfield and Nittrouer, 2014) (Fig. 3). Landslides are common in steep portions of the catchment (de la Fuente et al., 2006), but almost 70% of the load comes from the

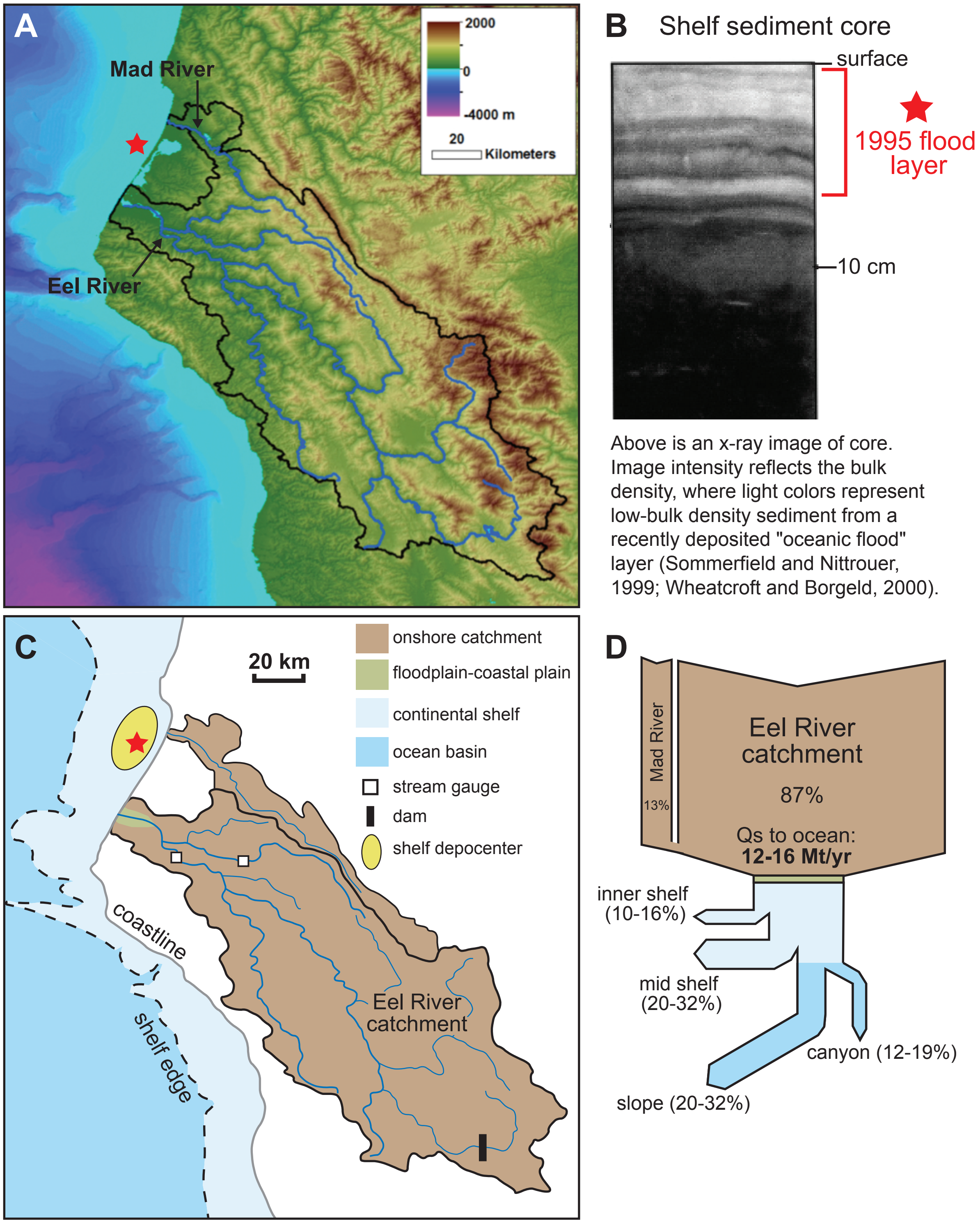

Figure 3: (A) Topography and drainage network of Eel and Mad river catchments, northern California, and bathymetry of the continental margin. Red star denotes location of shelf core x-radiograph shown in (B). (B) X-ray image of shelf reflects bulk density. Light colors (lower bulk density) interpreted as 1995 flood deposit (Sommerfield and Nittrouer, 1999; Wheatcroft and Borgeld, 2000). (C) Map of Eel-Mad sediment-routing system showing catchment area, areal extent of coastal floodplain, and shelf depocenter (yellow). Red star denotes location of shelf core image shown in (B). (D) Historical timescale sediment budget of the Eel-Mad sediment-routing system showing: 1) there is negligible onshore storage, 2) the shelf stores ~30-50% of the budget, and 3) the remainder moves to the canyon and continental slope. Budget estimations from Sommerfield and Nittrouer (1999) and Warrick (2014).

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central portion of the catchment where mélange outcrops are more erodible (Brown and Ritter, 1971). The largest recorded flood event occurred in December 1964, a year during which the Eel River is estimated to have discharged more than 160 Mt of sediment (Warrick, 2014). This is >13 times the annual average, with most discharge occurring over a few days. In 1995 (January and March) and 1997 (January), three floods occurred, and STRATAFORM scientists documented the deposition of a widespread layer on the shelf (Fig. 3) (Wheatcroft et al., 1997; Sommerfield and Nittrouer, 1999; Wheatcroft and Borgeld, 2000). The remarkable similarity between the flood deposits and decadal shelf sedimentation patterns demonstrate how important these events are to shelf construction. However, event and decadal sediment budgets indicate most (>50%) of the sediment is exported beyond the shelf (Fig. 3) giving testimony to the effective transport conditions associated with coherent discharge and energetic ocean conditions (Wheatcroft and Borgeld, 2000).

Instrument observations made in winter 1996-1997 revealed that a wave-enhanced sediment gravity flow associated with the floods transported an appreciable amount of sediment to the mid-shelf, exceeding other measured events by two orders of magnitude (Ogston et al., 2000; Traykovski et al., 2000). The widespread and distinctive shelf flood deposit is attributed to this mechanism; however, subsequent examination of the same deposit two years later indicated extensive reworking by physical and biological processes (Wheatcroft et al., 2007 and references therein). Although some shelf core records have stratigraphic and organic carbon evidence suggestive of older events (e.g., 1964 flood) (Sommerfield et al., 1999; Leithold et al., 2005), the documentation of post-event reworking indicates that the Eel shelf does not contain a laterally extensive or highfidelity record of flood signals (Goff et al., 2002; Wheatcroft et al., 2007 and references therein).

Subsequent coring and tripod research in the Eel Canyon documented the possibility of more direct sediment gravity flow to deeper water. Resuspension and transport of sediment via waves also were found to have an important control on this off-shelf export (Puig et al., 2003). Cores from the canyon indicate sedimentation is spatially and temporal complex, although export to deeper water is apparent (Mullenbach et al., 2004; Mullenbach and Nittrouer, 2006; Drexler et al., 2006). Nepheloid layers also transport fluvial sediment seaward of the Eel River mouth, allowing hemipelagic sedimentation to accumulate on the slope, but this modest input is easily reworked by the active benthic community precluding event layer formation (Alexander et al., 1999; Walsh and Nittrouer, 1999). These studies demonstrate that unravelling signals from Eel margin stratigraphic records is not straightforward, which a similar story for the Waipaoa Rivers of New Zealand (Kuehl et al., this volume). The apportionment of terrigenous sediment among shelf, slope, and deep-sea segments (Fig. 3D) suggests that shelf records might contain signals of sediment-production events that originated in the catchment, but post-depositional homogenization hampers event-scale determination.

The Ganges-Brahmaputra-Bengal is a large (1,656,000 km^2 catchment) sediment-routing system fed by tectonically active mountains. Sedimentation on the Bengal Fan (>2,000,000 km^2 depositional area), the ultimate sink for the system, has varied significantly since the Mesozoic because of plate tectonics (i.e., rifting and then collision in the Eocene) and associated sediment production (Curray, 2014 and references therein). Despite onshore foreland-basin accommodation created by ongoing collisional tectonics, sediments are moving through most of the system over historical timescales, from the Himalayas (>5000 m elevation) to the Bengal Fan (>4000 m water depth) (Fig. 4; Kuehl et al., 2005). Sediment production in the Ganges-Brahmaputra catchment (including the Meghna River) corresponds to an average catchment denudation rate of 365 mm/kyr, which is over an order of magnitude larger than the global average of 30 mm/kyr (Islam et al., 1999). The sediment load for the integrated catchment is ~1,000 Mt/yr, which equates to a system sediment yield of 556 t/km^2/yr. However, sediment yield varies significantly spatially across the catchment. For example, the Brahmaputra River yield is >140% that of the Ganges (Summerfield and Hulton, 1994; Islam et al., 1999), and most of the Brahmaputra sediment is sourced from a smaller portion of the catchment, the High Himalayas (Wesson, 2003).

Gauging stations for rivers are located about 300 km from the coast, and studies indicate ~30% of the sediment is stored landward of the coastline (Fig. 4) (Goodbred and Kuehl, 1999). Sediment sinks include levee, floodplain, and river-bed aggradation, and alluvial accumulation in tectonically subsiding areas (Allison, 1998; Goodbred and Kuehl, 1998). Longer timescale records show that since the middle Holocene (~7 ka) slowdown in sea-level rise, some locations have accumulated >20 m of sediment (Goodbred and Kuehl, 1998), and rates of filling since ~12 ka suggest significant climate forcing on sediment supply (Goodbred and Kuehl, 2000a). Over historical timescales, floodplain areas of the upper delta plain have linear sediment accumulation rates that generally decrease with distance from the river channel (e.g., from 4 cm/yr to <1 cm/yr, Allison, 1998). Sediment dynamics in the lower delta plain are influenced by processes that originate in the marine realm such as sea-level rise, waves, tides, and cyclones

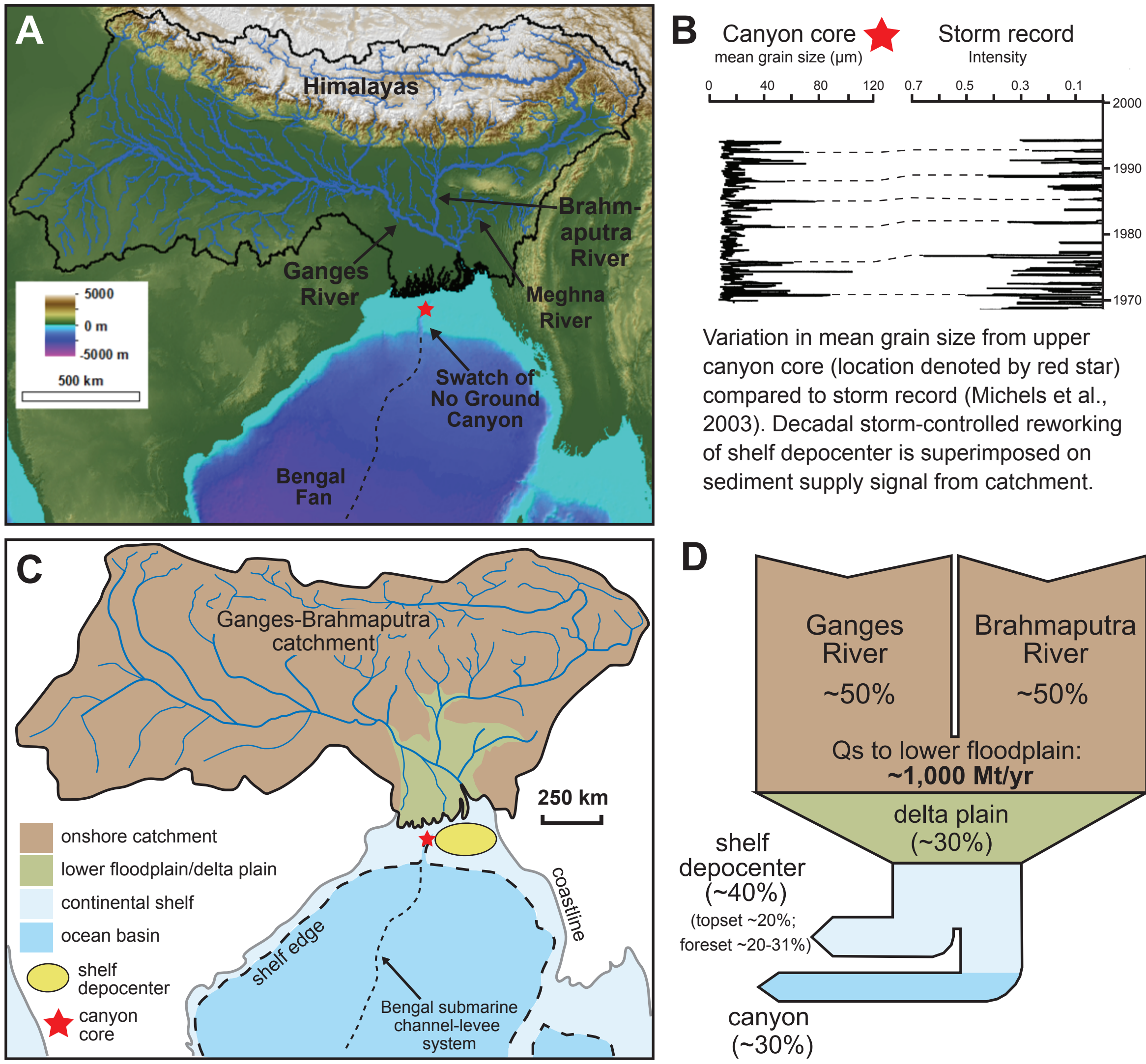

Figure 4: (A) Topography and drainage network of Ganges, Brahmaputra, and Meghna rivers and bathymetry of shelf, Swatch of No Ground submarine canyon, and part of the Bengal submarine fan system. Red star denotes location of core record shown in (B). (B) Core from upper canyon showing variation in mean grain size with time compared to storm record from eastern Bengal shelf. Data is from core 96 KL as reported in Michels et al. (2003). (C) Map of Ganges-Brahmaputra-Bengal sediment-routing system showing catchment area, the large delta plain area, shelf depocenter (yellow) and the Bengal submarine channel-levee system. Red star denotes location of core record shown in (B). (D) Historical timescale sediment budget of Ganges-Brahmaputra-Bengal sediment-routing system showing that almost one-third of the budget is stored on the delta plain, ~40% accumulates in the shelf depocenter, split between the topset and foreset regions, and the remaining ~30% is delivered to the canyon and Bengal submarine fan. Budget estimations from Kuehl et al. (2005) and references therein.

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(Allison and Kepple, 2001; Hanebuth et al., 2013). Shoreline areas show a complex pattern of erosion and accretion (Allison, 1998; Shearman et al., 2013), but radiochemical analyses indicate sediment accumulation generally decreases with distance from the coast, reflecting import of fluvial sediment (Allison and Kepple, 2001).

Seismic-reflection profiling has established the presence of a sizable subaqueous delta clinoform on the shelf (Kuehl el., 1997; Michels et al. 1998). Bathymetric and shoreline changes indicate that ~20% of the fluvial load is building the topset of the clinoform (Fig. 4) (Allison, 1998). Based on core and seismic-reflection data, the foreset region of the clinoform sequesters another 20-31% over historical timescales (Fig. 4) (Michels et al., 1998; Suckow et al., 2001). Transparent layers visible in seismic reflection profiles of the clinoform have been suggested to represent mass flows triggered by earthquakes (Michels et al., 1998). As a result of westward along-shelf currents reworking the delta front, sediment is at present being advected into the head of the Swatch of No Ground submarine canyon and episodically to the Bengal submarine fan (Kuehl et al., 1997; Kuehl et al., 2005). Weber et al. (1997) showed that late Holocene sedimentation occurred on the channel-levee complex (on the middle fan, ~500 km seaward of the shelf), but at a reduced rate compared to latest Pleistocene to early Holocene. Cyclones are hypothesized to be responsible for stratigraphic layering visible on the shelf and in the upper canyon (Fig. 4) (Kudrass et al., 1998; Suckow et al., 2001; Michels et al., 2003). Cyclones also serve as a possible trigger mechanism for episodic mass wasting events (Rogers and Goodbred, 2010). Canyon sedimentation and down-canyon transport, including evidence for turbidite deposition on the Bengal Fan, are hypothesized to account for ~30% of the fluvial load over historical timescales (Fig. 4)

(Goodbred and Kuehl, 1999; Kuehl et al., 2005). A terrestrial erosion-zone signal is being driven down this system, but it has been and continues to be significantly modulated by other processes (e.g., cyclones) along the way. As a result, alluvial storage areas might be the best sites for extracting source forcings over the historical timescale.

The Ganges-Brahmaputra-Bengal and Eel systems research highlight how historic stratigraphic records, accumulation rates, and sediment budgets can inform system functioning and source-to-sink transfer. This work also demonstrates that, although historical timescale records may be data rich and highly temporally resolved relative to intermediate and deep-time records, evaluation of sediment supply signals generated in upland catchments can be difficult. A more detailed and quantitative documentation of processes, rates, and spatial distribution of sedimentation does not necessarily equate to a better understanding of linkages between system segments. Better preserved and potentially more complete records in proximal storage areas, such as lakes, might allow more detailed records to be captured up system, but the localized nature might not reflect broader system functioning (e.g., Orpin et al., 2010). Combining observations from multiple localities will be essential to defining robust regional or broader, global signals (e.g., Noren et al., 2002). Other insights about catchment sediment production can be provided from the geomorphic record of erosional landforms. The sedimentary signature of events, such as floods, earthquakes, and storms, is likely more easily relatable to its forcing if process and response occur within the same or immediately adjacent segment(s) of the sediment-routing system (e.g., coastal overwash fan deposits from landfalling hurricanes; Boldt et al., 2010). The variability in sediment transport and associated deposits generated at $<10^2$ yr timescales is commonly considered noise over longer timescales as a consequence of combining event-scale and 'background' sedimentation into a time-averaged rate. However, the findings from historical timescale studies show that there are signals embedded within the noise.

3 Sediment Routing at Intermediate (10^2-10^6 yr) Timescales

The timescale from just beyond historical (several centuries before present; discussed above) to several millions of years is a critical temporal range in Earth surface dynamics because fundamental climate forcings (i.e., Milankovitch cycles) that control the global climate are prominent over this timescale (Hays et al., 1976). Sustained rates of rock uplift and deformation in tectonically active areas lead to exhumation, sediment production, and morphological change at $\geq 10^5$ yr timescales (Burbank and Anderson, 2011). Moreover, it is in this temporal range during which sedimentary deposits can be sufficiently buried to become rock and preserved into the stratigraphic record – durations often referred to as 'geological timescales' (e.g., Allen et al., 2013).

We first discuss sedimentary system dynamics and associated signal implications based on numerical and physical models. Unlike the short-term timescales during which an integration of direct observation, monitoring, and modeling informs our understanding of source-to-sink signal propagation, modeling and theory become even more critical for intermediate (10^2-10^6 yr) timescales. Examples of recent work on paleo-sediment budgets for sediment-routing systems are also discussed.

3.1 Model Predictions of Intermediate Timescale Signal Propagation

We review how tectonic or climatic signals with periods of 10^2–10^6 yr are propagated through different portions of the sediment routing system (Fig. 2). We emphasize sediment supply (Qs) as the principal vector for environmental signal propagation and aim to provide a review on the current state of knowledge with respect to the following important questions: (1) Does the erosion zone produce sediment supply signals in response to climate and tectonic perturbations with periods able to generate stratigraphic patterns? (2) Does the transfer zone faithfully transmit signals to the sedimentary basin, or does it modify signals coming from the erosion zone?

Signal transfer through a system depends on whether its period is smaller or larger than the response time of the system (Paola, 1992; see also Allen, 1974). Moreover, the action of internal, or autogenic, dynamics in any or all of the mass-flux zones can influence Qs behavior, which affects signal propagation. We first define and review knowledge of response times for the erosion zone of hillslopes and bedrock channels, then focus on the transfer zone of mixed alluvial and bedrock channels to alluvial channels with floodplains, and its linkage to the accumulation zone in sedimentary basins (Fig. 1).

3.1.1 Qs Signal Generation and Propagation in the Erosion Zone

It is beyond our scope to present a comprehensive review of investigations into how climate and/or tectonic forcings are recorded in net-erosional landscapes (e.g., Burbank and Anderson, 2011; and references therein). Rather, we emphasize the propagation of those perturbations out of the erosion zone in the form of sediment supply.

The concept of steady state as applied to landscape evolution (e.g., Willett and Brandon, 2002) refers to a state in which Earth's surface elevation relative to a datum is broadly constant as a result of a balance between rock uplift and erosion. Thus, the rate of sediment supply out of an area in steady state is, in the simplest case, also constant when averaged beyond timescales of individual events. A perturbation in the form of varying rates of tectonic movement or precipitation induces a response of the landscape system in the form of varying rates of sediment production. A characteristic equilibrium, or response, time for the system is the time that it takes for this transient landscape to respond to this perturbation and then return to a steady state (Beaumont et al., 2000) (Fig. 5). Allen (2008b) termed landscapes that have a response time shorter than the repeat time of the perturbation as 'reactive' and those with response times longer than perturbation repeat time as 'buffered' landscapes. This equilibrium time is critical to the discussion of signal propagation because it is the *variability* of Qs out of the erosion zone that can result in recognizable variations in deposit character down system. Here, we focus on the relevance of steady state in terms of denudation because of its close association with sediment production.

The physical laboratory experiments of Bonnet and Crave (2003) highlighted the important observation that climate signals, because they can affect the totality of an area at once, can trigger an immediate response of the landscape. In their case, steady state is characterized by a constant mean elevation (Montgomery, 2001; Willet and Brandon, 2002) and, thus, a response is a change in mean elevation. This contrasts to rock uplift signals, expressed in the form of baselevel changes (see Schumm, 1993), which propagate as waves of headward incision and diachronously affect the landscape (Bonnet

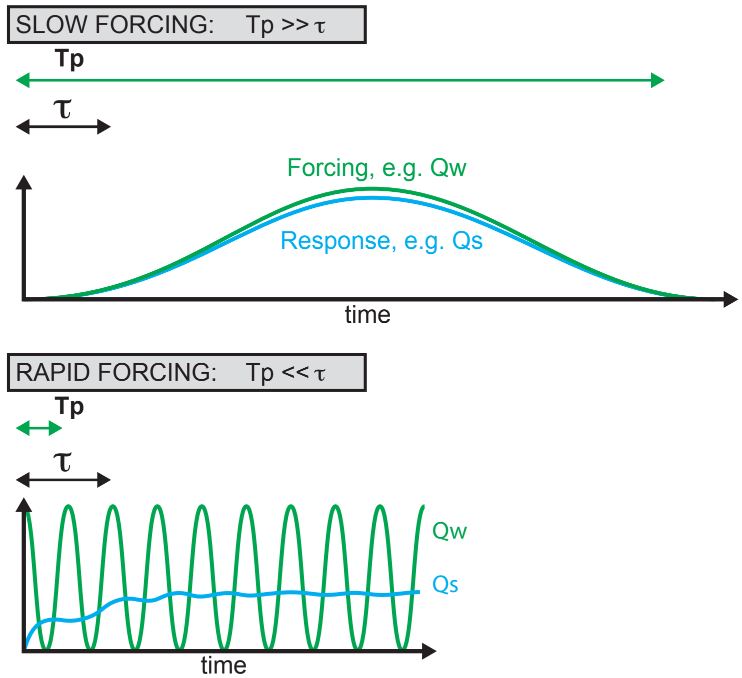

Figure 5: The ratio between the timescale of a perturbation (Tp) and the characteristic equilibrium timescale (τ) of the considered system describes the system response to forcing (after Beaumont et al., 2000; see also Allen, 2008b). A forcing of water discharge (Qw) and a response of sediment supply (Qs) are shown for (A) a reactive response when response time is much shorter than timescale of forcing and (B) a buffered response when response time is longer than timescale of perturbation.

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and Crave, 2003). In a study of the response of bedrock channels to tectonic and climate signals using generic stream-power fluvial incision, Whipple (2001) showed response times ranging from 250 kyr to 2.5 Myr to both tectonics and climate. In this study, the response time is the time required for the landscape to return to a steady state defined as a statistically invariant topography (i.e., constant mean elevation; Montgomery, 2001; Willet and Brandon, 2002) and constant denudation rate. When climate and tectonics act jointly, the response of a stream-power fluvial landscape may essentially be immediate (i.e., response time tends to zero; Whipple, 2001). Improvements of the stream-power erosion law produce divergent results as to landscape reactivity. Among these, the consideration of dynamic adjustment of channel width during perturbations induces faster reaction of fluvial landscapes than if channel width is not considered (Attal et al., 2008; Whittaker et al., 2007). Conversely, a series of stream-power inspired models (e.g., Gasparini et al., 2007) including a degree of dependency to saltating bedload tends to suggest longer response times than those predicted by detachment-limited stream power (such as those of Whipple, 2001; see above).

Using a nonlinear 1D diffusive model of catchment erosion, Armitage et al. (2013) showed that small (10-20 km long) catchments, such as those draining normal-fault bounded footwalls, are reactive to single-step, sustained changes of precipitation but tend to temporally buffer cyclic precipitation variations with Milankovitch periodicities (i.e., 100 kyr, 400 kyr, 1.2 Myr). Using a 2D model of landscape evolution including diffusive hillslopes and detachment-limited stream-power-governed bedrock incision, Godard et al. (2013) found that a given landscape possesses a characteristic resonance periodicity for which landscape response to corresponding climatic oscillation is maximized in terms of sediment supply. For more easily erodible lithologies, landscape response to orbitally controlled climate signals could be a significant increase or decrease of the amplitude of sediment supply variations. Thus, some landscapes respond to, might even amplify, climate and tectonic signals. In landscapes dominated by diffusive hillslopes, however, such as in soil-mantled, low-relief settings, diffusion itself might be very efficient at filtering climatic or tectonic oscillations because of slow signal propagation (e.g., Furbish and Fagherazzi, 2001).

3.1.2 Qs Signal Generation and Propagation in the Transfer Zone and Preservation in the Accumulation Zone

In many instances, the terminal depositional sink is not immediately adjacent to the source area but linked to it by a fluvial system. In such cases, it was recognized that the fundamental problem becomes whether climate and tectonic sediment supply signals that originate in the erosion zone are propagated by the transfer system to the sedimentary basin (Castelltort and Van Den Driessche, 2003). Paola et al. (1992) developed the idea that to understand stratigraphic response to external factors it was fundamental to consider the periodicity of cyclic signals with respect to the characteristic equilibrium, or response, time (T_{eq}) of a sediment-routing system (Fig. 5). They expressed T_{eq} (time unit) for a 1D fluvial profile as a function of characteristic system length (L) and diffusivity (K):

$$T_{eq} \sim L^2/K$$

Thus, the larger the system (i.e., the longer the transfer zone), the longer its response time is whereas the more diffusive the transfer zone, the shorter its response time. A prediction of this model is that cyclic perturbations with periods less than T_{eq} are buffered by the system's response time. In contrast, variations of boundary conditions with periodicities greater than T_{eq} produce stratigraphic patterns in the sedimentary basin, but these patterns might be similar for subsidence and sediment supply variation (Paola et al., 1992; see also Marr et al., 2000 and Allen, 2008b).

On the basis of a comparison between the modern river sediment discharge of some large Asian rivers and the average sediment discharge deduced from sedimentary basins over the last 2 million years, Métivier and Gaudemer (1999) suggested that large alluvial systems of Asia behave as diffusive entities buffering the high-frequency climate change known for the late Cenozoic (see also Schaller et al., 2001; and Wittmann et al., 2011). Métivier and Gaudemer (1999) computed equilibrium times of >1 Myr for such rivers using an expression they proposed for the diffusivity (K) of large rivers as a function of sediment discharge (Qs), river channel or channel-and-floodplain width (W), and slope (S):

$$K=Qs/(W*S)$$

Following Métivier and Gaudemer (1999) results, Castelltort and Van Den Driessche (2003) calculated the diffusive response time of 93 of the largest modern rivers to investigate the down-system stratigraphic response to high-frequency (10^4 yr) cycles of sediment supply. Castelltort and Van Den Driessche (2003) find that the characteristic response times of transfer zones comprising large rivers, which typically include extensive floodplains, are 10^5-10^6 yr, exceeding the 10^4 yr climate oscillations. When channel width rather than alluvial valley width is used in this relationship the resulting response times are minimum response times.

These diffusion-based investigations suggest that temporary, and in some cases permanent, storage of sediment in catchment and/or transfer-zone sinks (see also Allen, 2008a; Wittmann et al., 2011; Covault et al., 2013; and references therein) can mask the down-system stratigraphic record of external perturbations to the sediment-routing system. In the case of a large, hinterland-river-continental margin sediment-routing system, this transient storage of sediment can result from deposition in floodplains (Allen, 2008b). Larger catchments can retain sediment for longer periods as a result of more space available for sediment storage and consequent resistance to complete hinterland-to-continental margin sediment transfer in response to short-term, small-magnitude external perturbations, such as local storms and earthquakes (Allen, 2008a). Métivier and Gaudemer (1999) suggested that rivers and floodplains proportionally adjust to climate changes and upstream denudation in buffered catchments in which sediment loads are approximately balanced over different timescales. That is, if upstream denudation is reduced, the river will incise its floodplain to keep the sediment load at the outlet constant. Conversely, if climate changes force greater upstream denudation, the river is likely to use that increased sediment load to recharge its previously excavated floodplain. In this way, the steady transfer of reworked floodplain sediment to an outlet can be maintained over a range of timescales (Métivier and Gaudemer, 1999; Phillips, 2003; Phillips and Slattery, 2006; Covault et al., 2013; among many others). The ubiquitous alluvial terrace fills that ornate many river systems worldwide are witnesses of the residence time of sediments in the transfer zone.

These theoretical results contrast with the sensitivity to Late Quaternary climate change apparently displayed by some large fluvial systems such as the Ganges-Brahmaputra (Goodbred and Kuehl, 1999; 2000a, 2000b; Goodbred, 2003) and suggest that, although alluvial systems may behave diffusively in response to sediment supply variations, they may be sensitive to perturbations of water discharge, which can increase or decrease diffusivity (Simpson and Castelltort, 2012). Using physical laboratory models of river response to water discharge and sediment supply change, Van Den Berg Van Saparoea and Postma (2008) show that experimental rivers respond faster to changes of discharge than to perturbations of up-system sediment supply. Van Den Berg Van Saparoea and Postma (2008) concluded that high-frequency cyclic patterns in marine delta-shelf successions were most likely controlled by high-frequency changes in discharge driven by climate, whereas the low-frequency sequences were likely a result of low-frequency changes in sediment supply driven by tectonic deformation.

We recognize that the results of diffusion-based approaches may be dependent on our current ability to estimate parameters of the diffusion laws. Simpson and Castelltort (2012) explored the response of a 1D alluvial river bed to sediment concentration and water discharge pulses using a physically based numerical model of interacting water flow and sediment transport without *a priori* assumption of diffusive behavior. Consistent with the experiments of Van Den Berg Van Saparoea and Postma (2008), in this model the strong coupling between water discharge and river gradient induces amplified sediment supply variations in response to oscillations of water discharge, whereas sediment supply oscillations are dampened because of the negative feedback between sediment concentration and channel gradient. In the future, additional constraints on the behavior of sediment transfer will result from other approaches such as computational fluid dynamics (e.g., Edmonds and Slingerland, 2007) or cellular automata (e.g., Murray and Paola, 1997).

3.1.3 Potential Influence of Internal Dynamics on Qs Signal Recognition

In addition to the buffering of signals linked with the processes reviewed in the previous sub-section, perturbations to sedimentary signal propagation arise from sedimentary processes occurring within the river-floodplain and/or river-coastal plain segments that need not be driven by up-system forcings. Such self-organizing processes, referred to as autogenic dynamics (Paola et al., 2009), and first emphasized by Beerbower (1964), can create organized depositional architecture (e.g., Hoyal and Sheets, 2009). Critical to this discussion is the potential for climate or tectonic signals that originated in the catchment to be significantly masked, modified, or 'shredded' by such autogenic dynamics (Jerolmack and Paola, 2010). Variability in sediment transport is a result of the following general autogenic cycle: transient storage of sediment, exceedance of some critical threshold, and release of sediment during relaxation following failure. Jerolmack and Paola (2010) likened the threshold behavior of sediment storage and release to morphodynamic turbulence, analogous to turbulence in fluid flows.

Recently, Ganti et al. (2014) developed a quantitative framework to isolate autogenic, morphodynamic processes from external, environmental forcings in the stratigraphic record. They showed that the calculated advection length (l_a) for settling sediment sets bounds on the scale over which autogenic processes operate:

$$l_a = u\, h_s/w_s$$

where u is the flow velocity, h_s is the average sediment settling height, and w_s is the settling velocity. The advection length scale is the horizontal length over which an average particle is transported in the flow before falling to the bed. Ganti et al. (2014) argued that morphodynamic feedbacks, or autogenic 'shredding,' can only occur if the length scale of interest, e.g., the system size, is larger than l_a.

Wang et al. (2011) recognized the aforementioned work on damping or 'shredding' of upstream, external signals by autogenic sediment transport processes, and used numerical and physical experiments, as well as some field data, to gain insight into the timescale of compensational stacking of deposits within a basin. This compensation timescale (T_c) is defined as:

$$T_c = l/r$$

where l is a roughness length scale, equal to the amount of topographic 'mounding' due to local channel deposition produced between each avulsion, and r is the basin-wide, long-term sediment accumulation rate. This equation suggests that the geometry of deposits carries the signature of stochastic autogenic dynamics during the time necessary to fill a basin to a depth equal to the amount of surface roughness in a sediment-routing system. T_c provides an estimate of temporal scales below which stratigraphers should be cautious about interpreting signals. As a case in point, Wang et al. (2011) calculated T_c for the Lower Mississippi Delta in which they consider that the roughness length scale, l, was represented by the mean channel depth for the Lower Mississippi River of 30 m and a sediment accumulation rate of 0.26 m/kyr, estimated for the past 8 Myr (Straub et al., 2009). T_c is 115 kyr, which is ~100 times larger than the ~1300 yr recurrence of large avulsions of the Lower Mississippi River (Aslan et al., 2005). However, subsequent field data from the Lower Mississippi River indicate a rapid response to glacio-eustatic variation since Oxygen Isotope Stage 7 (~200 ka) (Shen et al., 2012). Large amplitude sea-level rise and fall prompted rapid and widespread fluvial aggradation and incision, respectively, the effects of which extended >600 km upstream from the present shoreline (Shen et al., 2012).

The models and experiments discussed above highlight that signal buffering as a result of sediment storage in up-system segments as well as depositional dynamics in the sink can mask the stratigraphic record of external perturbations to the sediment-routing system, although the quantitative expression of this are still being resolved. In summary, signals of a forcing can be passed to a basin and preserved in the stratigraphic record when their period exceeds the characteristic equilibrium time of the sediment-routing system, but this is valid only if their period is also larger than the characteristic timescale of autogenic sediment transport fluctuation and/or when the magnitude of the forcing is larger than the magnitude of internal oscillations (e.g., on the order of the size of catchment and alluvial accommodation) (Jerolmack and Paola, 2010). In the next section we review sediment budgets of natural sedimentary systems, which allow for accounting of our principal vector of relevance, Qs.

3.2 Paleo-Sediment Budgets of Natural Systems and Implications for Signal Propagation

In this section we will review work on sediment budgets at 10^2-10^5 yr timescales and implications for signal propagation via three sediment-routing systems: (1) tectonically active, small systems of southern California; (2) tectonically quiescent, larger systems of the northwestern Gulf of Mexico; and (3) tectonically active, larger systems of southern Asia. By focusing on sediment delivery from onshore catchments to the deep sea, which is the ultimate sink for coarse-grained terrigenous material, we highlight the role of the shelf as a Qs gateway and filter.

3.2.1 Methods for Paleo-Sediment Budget Reconstruction at Intermediate Timescales

Just as microfossils are the carriers of isotopes used to reconstruct geochemical signals, sediment supply is here considered the carrier of climate and tectonic signals. Thus, determining a paleo-sediment budget, the spatial and temporal partitioning of mass removed, transferred, and deposited within a routing system, is valuable for the interpretation of signal propagation and preservation. For the sake of brevity, we do not present a comprehensive review of the application of sediment budget concepts to timescales beyond direct measurement and instead refer the reader to a recent review by Hinderer (2012). Determining accumulated mass from stratigraphic volumes is straightforward in concept, but can be challenging in practice as a result of lack of appropriate data (e.g., seismic-reflection with proper chronologic control) and/or uncertainties in post-depositional stratal preservation (Sadler and Jerolmack, 2014). The geochemistry and mineralogy of sediment is often used to determine routing pathways as well as the relative contributions and potential residence times of terrigenous versus marine-derived material.

Two of the three systems reviewed below combine cosmogenic radionuclide (CRN) analysis for catchment-integrated denudation and radiocarbon dating for continental-margin deposition to reconstruct sediment budgets. Advances in CRN analysis provide catchment-integrated denudation rates and sediment loads at 10^2–10^5 yr timescales (von Blanckenburg, 2005), which are comparably similar to the timescales of deposition measured in offshore basins with radiocarbon ages (generally <50 ka; Reimer, 2012). CRNs are produced *in situ* as secondary cosmic rays interact with rocks within meters of Earth's surface; longer exposure to secondary cosmic rays as a result of slower denudation produces more nuclides. Sediment can be liberated from these rocks, mixed in the catchment through hillslope and fluvial transport processes, and ultimately deposited near the catchment outlet. Accordingly, the CRN abundance measured in sediment deposited near the catchment outlet can be used to divulge the catchment-wide denudation rate, which is inversely proportional to nuclide abundance (Brown et al. 1995; Bierman and Steig 1996; Granger et al. 1996).

Regardless of the specific tools used, it is of critical importance that all mass inputs and outputs to the system are considered and accounted for. Attempting to close a sediment budget at timescales beyond direct measurement provides an opportunity to evaluate other inputs and outputs that might not be evident with a qualitative interpretation.

3.2.2 Small and Tectonically Active Systems of Southern California

Tectonically active southern California is an ideal setting in which to investigate millennial-scale mass balance as a result of close proximity of sediment-routing components: onshore erosion zones are located adjacent to short alluvial-coastal plain depositional environments and offshore, confined sedimentary basins of the California Continental Borderland (Fig. 6A). The confinement of the offshore basins facilitates complete accounting for detrital mass relative to open-ocean basins, such as the Arabian Gulf and Bay of Bengal (Weber et al., 1997; Curray et al., 2002). Furthermore, many of the submarine canyon and fan systems of the California Continental Borderland are consistently linked to the shoreline and maintain connectivity even during Holocene highstand (Normark et al., 2009).

Covault et al. (2011) used CRNs from the Peninsular Ranges of southern California to calculate catchment-integrated denudation rates, which varied from 0.07 to 0.24 mm/yr since 10 ka. These denudation rates were calculated to be 1.9-2.4 Mt/yr and integrated across the total area of drainage basins (>6 x 10^3 km^2) delivering sediment to the offshore Oceanside littoral cell and the La Jolla submarine canyon and fan system. Based on radiocarbon-constrained seismic-reflection mapping (Covault et al., 2007) the mass accumulation rate of the La Jolla submarine fan was calculated to be 2.6-3.5 Mt/yr since the Last Glacial Maximum. Although the mass of material denuded from Peninsular Ranges catchments is in close agreement, of the same order of magnitude, as the mass of material deposited in the La Jolla submarine fan, deep-sea deposition exceeds terrestrial denudation by 11%-89%. This additional supply of sediment could be owed to enhanced dispersal of sediment across the shelf caused by sea cliff erosion during postglacial shoreline transgression and initiation of submarine mass wasting.

The terrestrial source to deep-sea sink mass balance does not show orders of magnitude inequalities that might be expected in the wake of major sea-level changes since the Last Glacial Maximum. Thus, sediment-routing processes in a globally significant class of small, tectonically active systems might be fundamentally different from those of larger systems that drain entire orogens, in which sediment storage in coastal plains and wide continental shelves can exceed millions of years (Milliman and Syvitski, 1992). Furthermore, in such small systems, depositional changes in the deep offshore can reflect onshore changes when viewed over timescales of several thousands of years to more than 10 kyr. For example, Romans et al. (2009) and Covault et al. (2010) examined Holocene deposition of the Hueneme and Newport deep-sea depositional systems offshore of southern California. Integrated datasets of radiocarbon ages from sediment cores and seismic-reflection profiles demonstrated that variability in rates of Holocene deep-sea turbidite deposition is related to complex ocean-atmosphere interactions, including enhanced magnitude and frequency of El Niño-Southern Oscillation (ENSO) cycles, which increased precipitation and fluvial water and sediment discharge in southern California (Fig. 7). Thus, millennial-scale climate forcings are represented as a measureable signal in the stratigraphic record of the deep-sea segment.

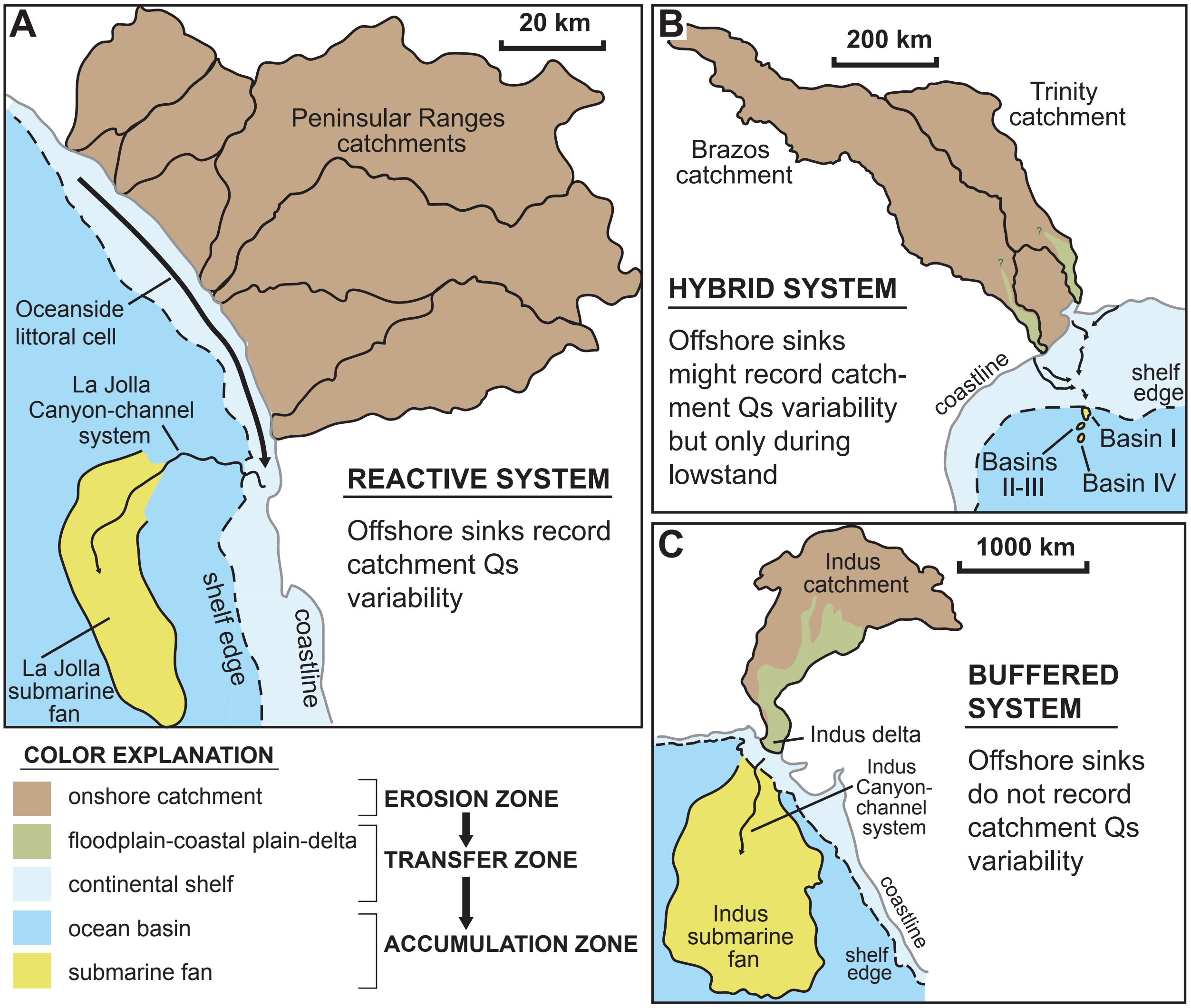


Figure 6: Examples of natural sediment routing systems examined at intermediate timescales. (A) Small and tectonically active system, Peninsular Ranges and Continental Borderland of southern California (Covault et al., 2011); (B) Large and tectonically quiescent system, Texas coastal plain and western Gulf of Mexico (Hidy et al., 2014; Pirmez et al., 2012); (C) Large and tectonically active system, Indus River and Indus submarine fan (Clift et al., 2014). Indus River floodplain extent from Milliman et al. (1984).

Romans et al. -- Figure 6

3.2.3 Large and Tectonically Quiescent Systems of the Western Gulf of Mexico

A sediment budget for the Brazos and Trinity rivers linked to offshore depositional systems in shallow-marine and deep-water environments of the northwestern Gulf of Mexico can be balanced by integrating recent work of Hidy et al. (2014) and Pirmez et al. (2012). In contrast to the small and tectonically active southern California catchments, the Brazos and Trinity rivers drain a large (~2 x 10^5 km^2) tectonically quiescent, non-glaciated, low-relief landscape (Fig. 6B). Hidy et al. (2014) evaluated how denudation rates from CRNs responded to climate change during the last glacial cycle (~15-45 ka): Brazos River CRNs yielded a mass load of 5.3 Mt/yr since 35 ka; and Trinity River CRNs yielded a mass load of 2-4 Mt/yr. Furthermore, Hidy et al. (2014) analyzed the CRN ratio of $^{26}Al/^{10}Be$ in river sediment to evaluate its transient storage in the catchment in route to its final depositional site (see Wittmann and von Blanckenburg, 2009; Wittmann et al., 2011). Mass storage on the coastal plain was interpreted to have been greater during glacial periods with lower sea level. Denudation rates and mass loads were calculated to be larger during interglacial periods, which suggest that increased weathering rates associated with warmer climates accelerated landscape erosion. Furthermore, increased mass load measured during warm interglacial periods is interpreted to reflect stronger reworking and delivery of sediment to the river mouth than during cooler glacial periods. An implication of this relationship between temperature and mass load is that global sediment and potentially dissolved load delivery to the ocean from analogous, tectonically quiescent, non-glaciated, low-relief landscapes might have been larger during the warm Pliocene than the cooler Quaternary (Hidy et al., 2014). However, any transient storage of sediment prior to preservation in terrace deposits would complicate the interpretation of the CRN data as representative of catchment denudation.

Pirmez et al. (2012) developed a robust chronostratigraphic framework from Oxygen Isotope Stage 6, ~120 ka, through the Last Glacial Maximum for the sediment deposited in four deep-water, salt-withdrawal slope basins of the northwestern Gulf of Mexico (see also Prather et al., 2012) (Fig. 6B). The deep-water depositional systems were linked to the Brazos and Trinity river-deltas only during lowstands of sea level, when the shoreline had regressed >100 km from the present-day beach to the shelf edge (Mallarino et al., 2006; Anderson et al., this volume). Chronostratigraphy was interpreted from an integrated database of 3D seismic-reflection data, age control from analysis of cores from Integrated Ocean Drilling Program Expedition 308, analysis of proprietary cores from Shell Oil Company, and published literature (Pirmez et al., 2012). The majority of sediment, ~50 km^3, was calculated to have been deposited during a period of relatively low sea level, between ~15-24 ka, yielding a mass accumulation rate during this period of 5.5 Mt/yr, which is within the same order of magnitude of the CRN-derived mass load of the Brazos and Trinity rivers of 7.3-9.3 Mt/yr (Hidy et al., 2014) during a similar period, generally <35 ka. The imbalance in rates, with diminished deep-water slope basin mass accumulation, is likely a result of Brazos-Trinity river-delta deposition on the subaerially exposed shelf as the shoreline had regressed to the shelf edge between ~15-24 ka. Indeed, Pirmez et al. (2012) estimate a maximum volume of ~25 km^3 of deltaic sediment was deposited between ~15-24 ka, which yields a mass accumulation rate of 2.8 Mt/yr. This mass added to the deep-water basin fill yields a total mass accumulation rate of 8.3 Mt/yr, which is within the range of CRN-derived mass load (Hidy et al., 2014) of the Brazos and Trinity rivers.

In this system, sea level is interpreted to control the delivery of terrigenous sediment to the deep-water slope basins over 10^5 yr timescales: during periods of relatively high sea level, when the shoreline had transgressed, the slope basins were interpreted to receive only hemipelagic, fine-grained sediment. During periods of relatively low sea level and a subaerially exposed shelf, relatively coarse-grained terrigenous sediment was deposited in the slope basins. However, mass accumulation during periods of relatively low sea level was interpreted to have varied between the four slope basins as a result of a complex interaction between river-delta sediment routing and dynamic, salt-withdrawal slope-basin evolution. The complex history of sediment deposition, storage, and remobilization in the zone between the modern coastline and the shelf edge over the past ~125 kyr (Anderson et al., this volume) highlights the role of the shelf as an additional filter of signals generated in the catchment. Therefore, in contrast to the southern California examples, offshore depositional records are hypothesized to primarily reflect sea-level-driven accommodation changes as opposed to Qs variability from the onshore catchments. However, during glacial periods of terrigenous sediment delivery to deep water, depositional records might reflect Qs variability from rivers (Fig. 6B). This might be common to other sediment-routing systems, where deep-water canyon heads are stranded at the edges of drowned continental shelves during interglacial periods (Blum and Hattier-Womack, 2009; Covault and Graham, 2010).

3.2.4 Large and Tectonically Active Systems of Southern Asia

Since the Last Glacial Maximum (LGM), the Indus sediment-routing system comprised a steep (total relief of nearly 8 km), tectonically active hinterland and large (~10^6 km^2) catchment (Milliman and Farnsworth, 2011), a delta located on a wide (~120 km) shelf, and a submarine canyon that fed the second largest accumulation of terrigenous sediment in the world, the Indus submarine fan (Fig. 6C). Clift et al. (2014) used seismic-reflection data, radiocarbon ages, and analyzed the geochemistry (a suite of major and trace elements, including Zr/Rb, K/Rb, and Nd; Limmer et al., 2012) and mineralogy of sediment of the Pakistani continental margin to investigate sediment routing from the Indus river-delta to the upper submarine canyon (<1300 m below present sea level) since the LGM. Seismic stratigraphic interpretations of deltaic clinoforms and radiocarbon ages indicate that the majority of Holocene Indus river-delta sediment is stored on the shelf (Giosan et al., 2006; Clift et al., 2014). Clift et al. (2014) interpreted a variety of deltaic clinoform seismic reflections and concluded that sediment used to construct the shelf-edge delta deposits was reworked and dispersed from mid-shelf locations basinward. Neodymium isotopes presented by Limmer et al. (2012) suggests transient storage on the shelf was significant prior to delivery to the canyon and fan system. Neodymium isotope ratios indicate different values compared to those expected from a fluvial source, which points to reworking of marine sediment deposited during the LGM (Clift et al., 2014). Deposition at the head of the Indus Canyon was measured to be rapid during the Holocene, with evidence for annual delivery of Indus river-delta sediment. However, downstream, ~1300 m below present sea level, the youngest deposits are greater than approximately 7 ka, and no terrigenous sand has reached the upper submarine fan during the Holocene (Clift et al., 2014).

The Indus sedimentary record at the Pakistani continental margin since the LGM indicates reworking and transient storage of sediment on the shelf and within the submarine canyon en route to the deep Arabian Sea. Thus, deep-water deposits of the Indus fan likely do not faithfully record Qs variability related to climatic or tectonic events onshore over timescales of 10^3-10^4 years (Clift et al., 2014). This conclusion highlights the role of the shelf segment as a critical process boundary as well as a Qs gateway between land and sea. This is similar to the Brazos-Trinity sediment-routing system (Fig. 6C), where the deep-water depositional record primarily reflects sea-level changes as opposed to Qs variability from the onshore catchments. Moreover, sediment storage in the vast Indus floodplain (Fig. 6C) likely buffers climatic or tectonic signals generated in the up-system headwaters of the Himalayas. Thus, in these large sediment-routing systems (see also the Amazon system, e.g., Wittman et al., 2011), millennial to million-year Qs signals that originated in the erosion zone are potentially recorded in alluvial and floodplain deposits.

3.3 Synthesis of Intermediate Timescale Signal Propagation

The theory, models, and natural systems reviewed above inform us about the plausibility of different forcings and how they might generate Qs signals that are then transmitted by the transfer zone to the ultimate depositional sink. These concepts are synthesized in Figure 8, which contains schematic representations of many of the scenarios reviewed in preceding sections, which can also be viewed as hypotheses about sediment-routing system dynamics worthy of further investigation.

Tectonic signals (e.g., uplift rates) with periods of >50 kyr are likely to produce Qs signals at the outlet of the erosion zone (Fig. 8A). Such signals may be transmitted to the accumulation zone if the transfer zone is short (<300 km) but will be buffered if the transfer zone is long (>300 km), unless their period exceeds 100 kyr. Tectonic signals with periods of <50 kyr are likely to be already buffered by the dynamics of the erosion zone itself; i.e., before they reach the outlet of the erosion zone (Fig. 8B).

Several antithetic results exist with respect to climate signals (e.g., precipitation changes) in the erosion zone. Different models propose that climate signals might be buffered, faithfully transmitted, or even amplified by the erosion zone (Fig. 8C-E). In some natural examples, where sediment budgets have determined the magnitude of sediment supply exiting the erosion zone, there is a relationship between climate (e.g., variation in precipitation) and Qs signals as recorded in the sink (Covault et al., 2009; Romans et al., 2009; Fig. 7). However, it is challenging to test whether climate signals are amplified since no predictive understanding exists between a given climate change and the corresponding amplitude of catchment response in terms of sediment supply. However, as discussed above, numerical models of sediment transport and deposition offer the opportunity to elucidate the question of climate signal amplification (e.g., Armitage et al., 2011). Nonetheless, intermediate climate signals in the form of water discharge (Qw) seem to trigger a strong response of the transfer zone in terms of sediment supply and may thus be transmitted faithfully or even amplified to sedimentary basins (Fig. 8C-E) (Godard et al., 2013; Simpson and Castelltort, 2012).

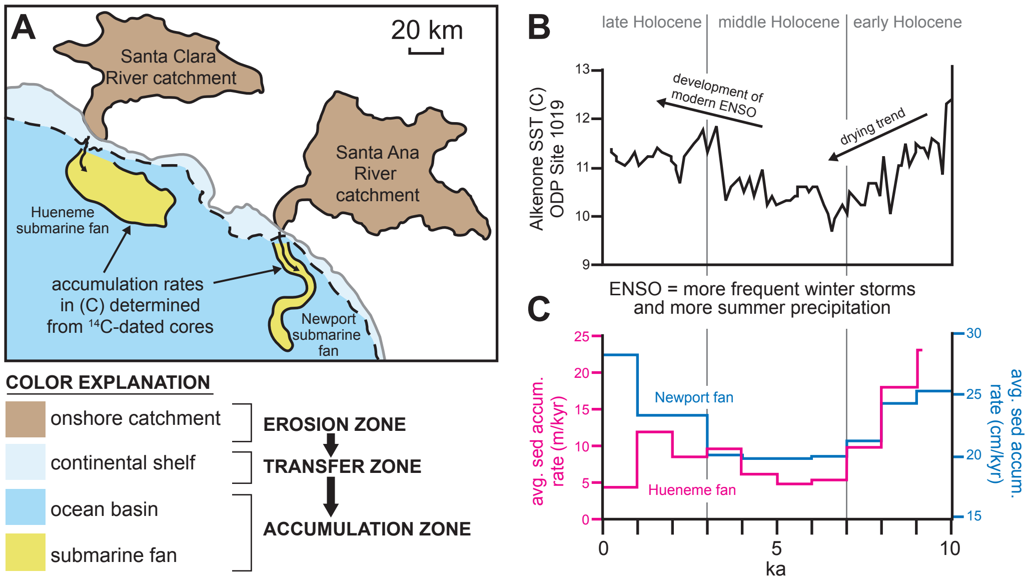
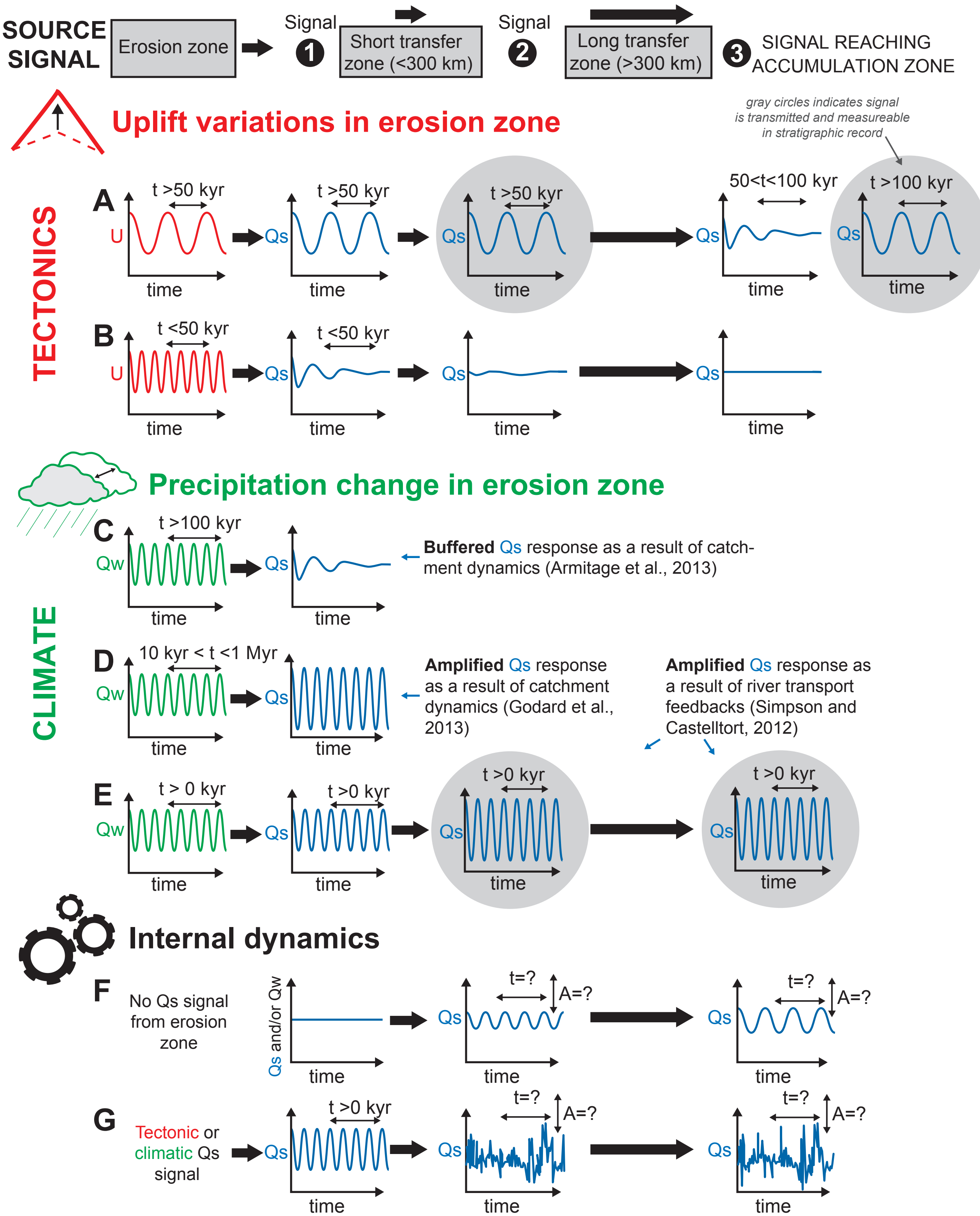


Figure 7: (A) Map showing two southern California sediment-routing systems, each with negligible onshore sediment storage at millennial timescales and correspondingly rapid transfer to offshore submarine fan systems. (B) Alkenone sea surface temperature (SST) proxy for the California coastal region showing a drying trend in the early Holocene followed by the development of the modern El Niño-Southern Oscillation (ENSO), which is known to be sensitive to increased SSTs (Barron et al., 2003). (C) Radiocarbon-constrained weighted-average sediment accumulation rates from Hueneme and Newport submarine fans (Romans et al., 2009; Covault et al., 2011) showing a general correlation of sediment supply to the basin to precipitation regime and, thus, propagation of climatic signal to the stratigraphic record

Romans et al. -- Figure 7



Romans et al. -- Figure 8 (see caption at end of manuscript text)

At the moment, few constraints exist on the characteristic saturation timescales and amplitudes of internally generated Qs fluctuations in natural systems. Current estimates of the typical timescale for channel stacking in large river systems (e.g., Hajek et al., 2010) suggest that autogenic dynamics may be able to completely mask or even destroy sedimentary signals with periods of less than 100 kyr.

4 Deep-Time (≥10^7 yr) Sediment Routing

4.1 Challenges and Uncertainties in Deep-Time Signal Propagation Analysis

As sedimentary systems age to tens of millions of years and older, the ability to explicitly measure or calculate source-to-sink sediment supply becomes increasingly challenging because: (1) sediment production areas are poorly preserved or not preserved at all, (2) there is increased uncertainty regarding boundary conditions such as tectonic setting and climate regime, (3) of the diminishing resolution of existing chronological tools, and (4) of the completeness of the stratigraphic record (Romans and Graham, 2013).

Reading the sedimentary record in deep time also requires understanding forcings with long periods. The maximum response times resolvable for erosional and depositional processes in most sedimentary systems is commonly ~10^6 yr (e.g., Paola et al., 1992; Whipple, 2001; Castelltort and Van Den Driessche, 2003; Allen, 2008b). Thus, tectonic and climatic signals with periods of at least several millions of years could induce a measureable equilibrium response of the Earth's surface that is potentially recorded in stratigraphic successions. In other words, long-period stratigraphic archives have more immunity to the internally generated dynamics that plague intermediate timescales. Examples of long-period forcings include the development of orogens and their coupled sedimentary basins (Dickinson, 1974; DeCelles et al., 2009), Phanerozoic changes of Earth's sea level (e.g., Miller et al., 2005), and significant shifts in global climate such as the transition from Cretaceous-Eocene greenhouse to Oligocene-present icehouse conditions (e.g., Zachos et al., 2008).

When peering back even further in time ($\geq 10^8$ yr) we lose details about the fundamental boundary conditions of tectonic setting and environmental conditions that are explicitly known for the modern or reliably reconstructed for historical to intermediate timescales. Thus, in many cases, reconstructing those boundary conditions is the primary goal of sedimentary analysis. Linking strata of such old ages to sediment source areas is challenging as a result of major tectonic regime changes (e.g., closing and opening of ocean basins; Wilson, 1966), poorly understood oceanic and atmospheric conditions, and non-actualistic Earth processes.

Our ability to reconstruct deep-time Earth surface conditions is based on rock availability: preserved as intact depositional architecture and/or detrital material representative of long-gone source areas. The following sections briefly review methodologies for characterizing the unpreserved sediment-production zones of ancient systems and potential value for interpreting Qs signal propagation.

4.2 *Inferring Catchment Characteristics and Sediment Supply From Stratigraphic Architecture*

Erosion and transfer zones are inherently not preserved in deep time and thus we must rely on preserved stratigraphy to reconstruct their characteristics. In this section we briefly review methods for estimating catchment area from stratigraphic architecture.

The dimensions of modern river channels scale to flood, or bankfull, discharge (Bridge and Demicco, 2008; and references therein), which affords estimation of water discharge from preserved fluvial channel stratigraphic architecture (e.g., Bridge and Tye, 2000; Bhattacharya and Tye, 2004; Adams and Bhattacharya, 2005; Blum et al., 2013; Bhattacharya et al., this volume). Analyses of modern river systems demonstrate a relationship between water discharge and catchment area (e.g., Hack, 1957; Rodier and Roche, 1984; Matthai, 1990) and a global empirical model shows that catchment area and relief are first-order controls on sediment supply (BQART model of Syvitski and Milliman, 2007). These relationships suggest that an estimate of paleo-catchment area can be determined from stratigraphy with the proviso that the channelized strata measured are truly representative of the alluvial system (see Blum et al., 2013 and Bhattacharya et al., this volume for discussion of river channel and alluvial valley scaling with respect to sediment delivery dynamics). However, discharge-to-area relationships as well as sediment load-to-area relationships from modern rivers are shown to vary as a function of precipitation and runoff patterns, vegetation, soil type, and geology (e.g., Milliman and Farnsworth, 2011; Covault et al., 2013 and references therein). Regional hydraulic curves, which capture such characteristics from modern systems (e.g., Leopold and Maddock, 1953), can be used to further constrain the estimate of catchment area if some aspects of the paleoclimate can be determined. Davidson and North (2009) provide a comprehensive discussion of the values and limitations of the regional hydraulic curve approach, including example applications from the deep-time rock record.

Sediment yield can be approximated with the regional hydraulic curve approach, providing insight about the paleo-sedimentary system. In practice, however, the calculation of any mass supply is only as accurate as the chronologic control available; mass supply averaged over >10^6 yr will obviously not capture shorter-period fluctuations. Furthermore, such paleo-hydrologic methods are burdened with uncertainties that are challenging to quantify, which limits accuracy to an order of magnitude (Holbrook and Wanas, 2014). These methods are also susceptible to aliasing the record of Qs. For example, a single, static paleogeographic reconstruction might be used to inform paleo-hydrology over a large duration of geologic time, which provides average Qs during that time. However, geologic evolution, and especially Qs, is dynamic and influenced by the extreme events. Some applications can be satisfactorily addressed with order-of-magnitude estimates, such as the selection of modern analogs for ancient systems (Bhattacharya and Tye, 2004; Bhattacharya et al., this volume). How to better link paleo-catchment reconstructions with interpretations of signal propagation remains a challenge. Ultimately, because these methods incorporate information from modern systems, the reliance on an actualistic approach should be acknowledged.

4.3 Source Area Signals From Detrital Material Analysis

Another approach to reconstruct aspects of the erosion zone of deep-time sediment-routing systems is to characterize the detrital material that is preserved in sedimentary rocks. Provenance analyses focused on detrital products released from the erosion zone has long been used to reconstruct and interpret deep-time paleo-drainage systems and their relationship to tectonic forcings (Dickinson, 1974; Graham et al., 1986; McLennan et al., 1983; among many others). More recently, radioisotope provenance studies have been employed to detect sediment supply signals in deep time by identifying source terranes (Fig. 9) and tracking their evolution through a basin fill (e.g., Dickinson and Gehrels, 2003; Weislogel et al., 2006; Romans et al., 2010; Carrapa, 2010; Blum and Pecha, 2014 and references therein). The common pre-conditions for application of such methods are related to specific characteristic of the source areas, which should be composed of rocks with different tectonic histories, distinctive crystallization and cooling ages, and the presence of the unique mineral(s) (Lawton, 2014).

Combining different thermochronometers on single detrital mineral grains such as zircon, monazite, white mica, and apatite can be used to determine cooling ages associated with different closure temperatures (depths) with which to reconstruct tectono-thermal events (Rahl et al., 2003; Carrapa et al., 2003; Carrapa, 2010; Lawton, 2014). This has potential application for the interpretation of the propagation of a tectonic signal across a paleo-landscape by constraining separate events (e.g., crystallization and cooling) of a single grain such that durations and rates can be determined. These methods provide rates for mountain belt emplacement and exhumation, helping to refine timing and magnitude of tectonic processes.

For example, when zircon U-Pb crystallization ages are coupled with zircon (U-Th)/He exhumation ages, apatite fission-track (AFT), and/or apatite (U-Th)/He methods, we gain critical insight into the timing of rock cooling, inferred exhumation, lag times

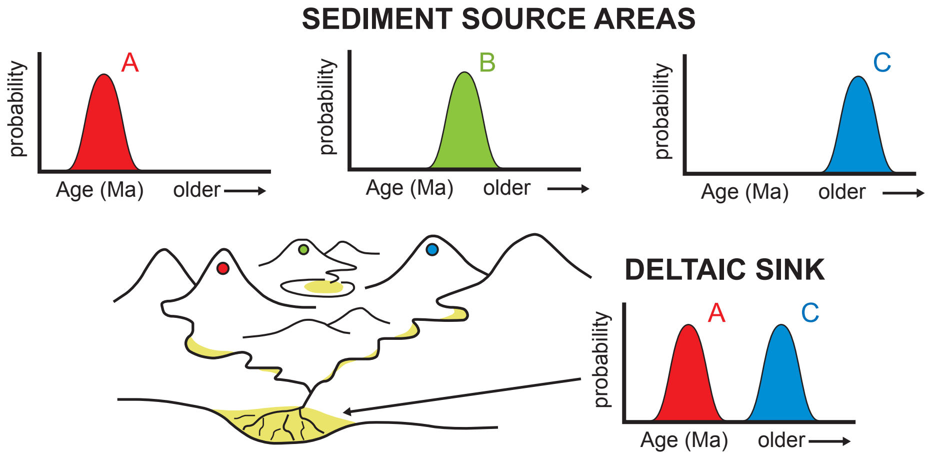


Figure 9: Conceptual diagram of crystallization age of detrital minerals (e.g., zircon U-Pb geochronology) as indicator of sediment-routing system connectivity, which can be used to aid reconstruction of sediment supply history from deep-time stratigraphic archives.

Romans et al. -- Figure 9

between different closure temperature depths, as well as lag times between cooling and deposition (Fig. 10) (e.g., Rahl et al., 2003; Reiners and Brandon, 2006; Painter et al., 2014). Lag time was originally defined as the difference between the cooling age of a detrital mineral and the depositional age of its host strata (Brandon and Vance, 1992; Garver et al., 1999). According to theoretical modeling, the variability in cooling age-depositional age lag times through a stratigraphic succession could be used to infer changing erosion rates in an orogenic belt (Rahl et al., 2007), which is a potentially powerful tool to estimate Qs in deep time (Fig. 10). Most of these studies focus on synorogenic basins with inferred short transfer zones and correspondingly short duration of transient storage in order to interpret orogenic signals. Minerals yielded by distinct source regions could display overlapping crystallization ages related to different volcanic events but virtually undistinguishable if using a U-Pb dating technique (e.g., Saylor et al., 2012). Coupling these ages with (U-Th)/He ages can help distinguish older exhumed regions from younger, especially within the context of the omnipresent Grenville U-Pb age populations of the Appalachian orogenic belt (Rahl et al., 2003).

Tectonic and/or sedimentary burial can reset thermochronometers, complicating a simple exhumation to erosion relationship. In some cases, however, orogenic recycling can be constrained by integrating double dating with time-temperature modeling of burial history (Fosdick et al., 2014). These techniques will continue to be used to reconstruct paleo-drainage system connectivity and evolution, which will help track changes in sediment supply over $\geq 10^7$ yr timescales (Carrapa et al., 2003).

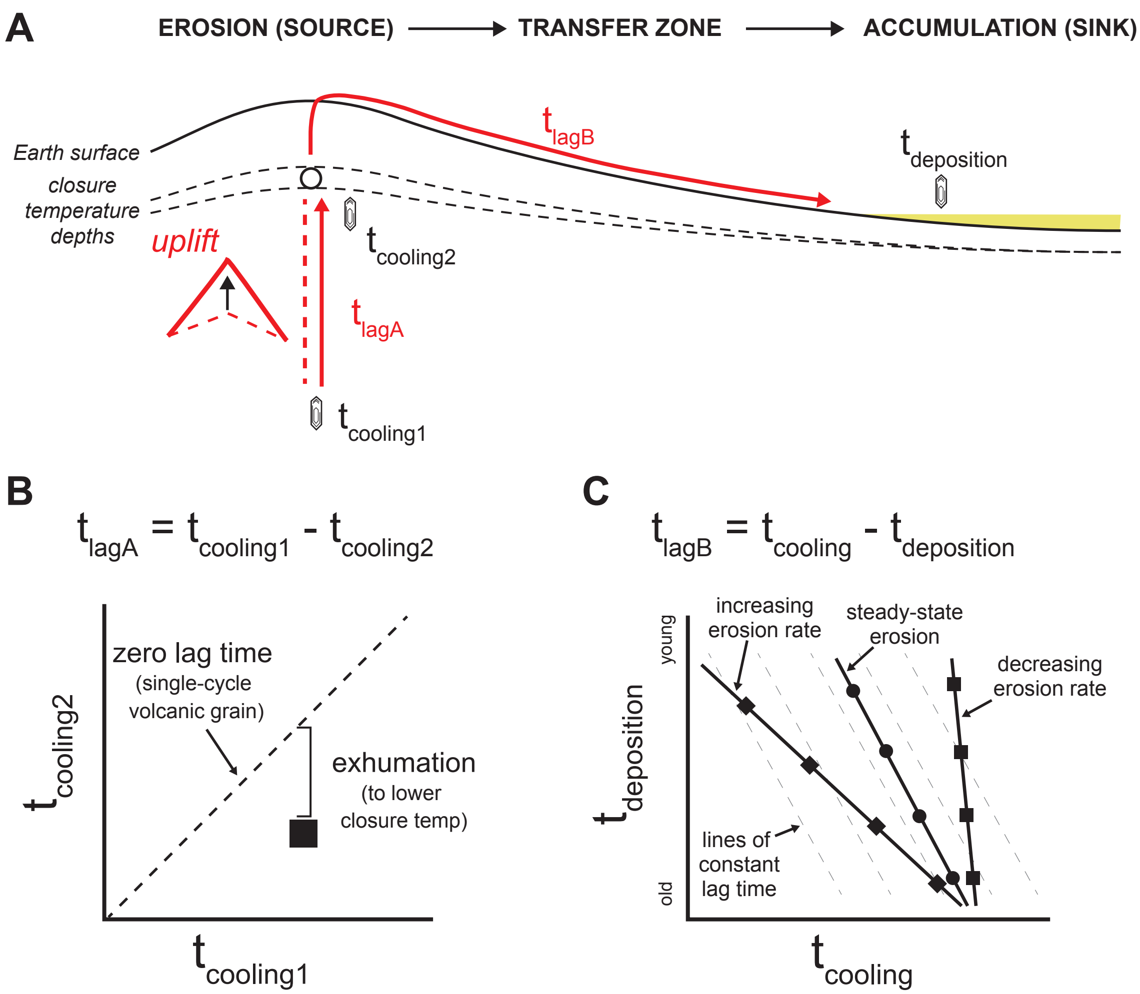


Figure 10: Conceptual diagrams of two different types of lag times to be calculated with combinations of cooling ages and depositional age. (A) Schematic cross section depicting trajectory of a particle through cooling via erosional exhumation followed by transport and deposition. (B) Lag time type A is the difference between higher-temperature cooling age (e.g., crystallization age) and lower-temperature cooling age of a single grain (e.g., zircon) and represents exhumation from depth. (C) Lag time type B is the difference between a cooling age and depositional age and represents the time from closure temperature depth to surface exhumation plus transport (and transient storage) in the sediment-routing system prior to deposition (adapted from Rahl et al., 2007). A consistent Lag time type B calculated through a stratigraphic succession indicates steady erosion whereas departures from that indicate changing erosion rates through time. Note that time within partial annealing/retention zones as well as sedimentary and/or tectonic burial can complicate simple lag time determinations.

Romans et al. -- Figure 10

4.4 Sedimentary System Mass Balance in Deep Time

As reviewed in preceding sections, a full accounting of mass supply (and storage) among erosion, transfer, and accumulation zones of a sediment-routing system can aid interpretation of signal propagation. For example, transient storage of sediment in floodplains and coastal plain segments at intermediate timescales can buffer the transmission of Milankovitch-controlled supply entering the sink (Fig. 6C). These same concepts can be applied to deep-time systems, but with the significant challenges that erosion/transfer zones are typically not morphologically preserved and chronology is more poorly resolved.

Despite these uncertainties, theory and methodologies have been and are being developed with the goal of characterizing mass-balance in ancient systems. Paola and Martin (2012) built on previous work of Paola and Voller (2005) and Strong et al. (2005) to apply mass-balance concepts to quantitative characterization of sedimentary basin fills. This and similar studies (e.g., Whittaker et al., 2011; Carvajal and Steel, 2012; Petter et al. 2013) aim to improve the estimation of Qs from time-averaged stratigraphy. Sadler and Jerolmack (2014) point out that well-documented 1D measurement-interval effects on rates of denudation (e.g., Gardner et al., 1987) and accumulation (e.g., Sadler, 1981) are eliminated with full spatial averaging because sediment generation and deposition must balance. Closing the sediment budget for an ancient system by accounting for all inputs and outputs is challenging (Hinderer, 2012; Allen et al., 2013). However, Sadler and Jerolmack (2014) make the case that avoiding linear rate measurements (i.e., maximizing volumetric analysis) and avoiding rates of any kind derived from the transfer zone can significantly minimize the measurement-interval bias.

In the same context as box models or other budget diagrams (e.g., Walling and Collins, 2008), Hay et al. (1989) developed a method for generating mass-balanced paleogeographic maps. These maps aim to depict the source-to-sink redistribution of mass through time by tracking paleotopographic evolution at $\geq 10^7$ yr timescales. Similarly, recent efforts by Meyers and Peters (2011) aim to reconstruct stratigraphic volume and mass distribution through time in relation to long-period ($\geq 10^7$ yr) tectonic and/or sea-level cycles. The utility of such methods to signal propagation has yet to be explicitly addressed. Following Allen et al. (2013), wherein the challenge of estimating sediment supply from strata is termed 'The Qs problem', Michael et al. (2013) determine a sediment budget for Late Eocene (~42-34 Ma) foreland basin strata and use resultant grain-size partitioning information to reconstruct tectonic subsidence.

The tectonically active region of southern Asia includes high rates of denudation and mass redistribution, making it an ideal locale to develop longer-term sediment budget concepts. Johnson (1994) accounted for the volume of sediment deposited in foreland basins, deltas, and the Indus and Bengal submarine fans forward of the Himalayas to explore Cenozoic sediment-routing evolution and its relationship to uplift and exhumation. Using information from literature, Johnson (1994) calculated that the Himalayan Main Central Thrust, which is interpreted to have been emplaced ~20 Ma, to be of insufficient volume to balance the depositional systems. Thus, Johnson (1994) posited that regions outside the Himalaya, including Tibet and the Karakoram, might have been significant sediment source areas during the Cenozoic, especially prior to the emplacement of the Himalayan Main Central Thrust.

Clift et al. (2001) used a seismic stratigraphic interpretation to account for the Cenozoic deposition of the Indus Fan. The erosional record on land was constrained from provenance analysis using Pb and Nd isotopic compositions and published thermochronology. Cenozoic land-to-deep sea Indus sediment routing shows a balance of erosion and deposition, with a greater volume of eroded rock during the Neogene, and corresponding greater deposition on the Indus submarine fan during the Neogene. However, the Paleogene was characterized by rapid erosion and coupled accumulation, with deposition focused in the regions of the Katawaz Basin, the Makran accretionary wedge, and the Indus foreland. More recently, Clift et al. (2008) compared published thermochronology from the Himalaya to weathering and climate proxies recorded in Neogene deposits from the South China Sea, Bay of Bengal, and Arabian Sea. Erosion of the Himalaya was interpreted to have intensified ~23-10 Ma, and slowed to ~3.5 Ma, but then began to increase during the Late Pliocene and Pleistocene, which correlates with monsoon intensity interpreted from climate proxies (Clift et al., 2008).

In these studies of sediment budgets of Cenozoic sediment routing systems, the detrital record provides insights into sediment routing evolution, as an indicator of Qs variability, and insights into the tectonic and climatic controls on erosion and deposition. Similar to the use of modern systems to guide the interpretation of transport and depositional processes from preserved strata, we can use historical-timescale (Figs. 3 and 4) and intermediate-timescale (Fig. 6) sediment-routing systems as analogs of signal-propagation dynamics for deep-time systems. Studies of Cenozoic systems are of particular value to better understand deep-time signal propagation because they combine long-period forcings with relatively well-documented boundary conditions that pre-

Cretaceous systems lack as a result of tectonic reorganization (Romans and Graham, 2013).

5 Discussion and Research Directions

External forcings are initially transformed into an Earth-surface signal through the production of mobile mass that is then redistributed down system as detritus and solutes (Fig. 1) (Allen et al., 2013). The character of that signal and to what extent a signal is preserved in sedimentary records are dependent on the magnitude and frequency of the initial forcing (Fig. 8), on their initial recording or destruction (Wheatcroft et al., 2007), on the responses of the different segments of the sediment-routing system (Castelltort and Van Den Driessche, 2003), on their morphology (e.g., for instance promoting buffered versus reactive sediment and signal transfer) (Covault et al., 2011), and on the ratio of signal to noise, in particular with respect to autogenic dynamics (Jerolmack and Paola, 2010). In this review, we emphasized the importance of sediment supply (Qs) as the main carrier of signals that originate in erosion zones. Thus, approaches for determining Qs by direct observation and measurement (Figs. 3 and 4), calculation from measurement of related process (e.g., denudation via cosmogenic radionuclides; Fig. 6), or estimation from stratigraphy and/or detrital minerals (Figs. 9 and 10) that are reviewed in this paper are critical to understanding how those signals are transmitted through the system. Furthermore, the development of new computational techniques for numerical modeling of source-to-sink sediment transport and deposition promise a deeper understanding of sediment and signal propagation. The benefit of unravelling processes of sediment and signal propagation is an enhanced understanding of the coupling of Earth surface systems and improved capability to invert stratigraphic and geomorphic records that relate to broader Earth dynamics.

The investigation of signal propagation requires a systems approach, which is provided by the sediment-routing system, or source-to-sink, framework (Allen, 2008a) (Fig. 1). The concept of sediment mass balance is embedded within such a framework and variations in system morphology provide insight into signal propagation and preservation. We emphasized the importance of the transfer zone (Fig. 1) because of its potential role as a 'buffer' (e.g., along-system diffusion and temporary floodplain deposition) and, correspondingly, the effect on rates and magnitudes of Qs carried to down-system segments. Such transfer-zone buffering of up-system signals is highly relevant to decoding the meaning of coastal and marine stratigraphic archives (Figs. 4 and 6). The transfer-zone concept at historical timescales ($<10^2$ yr) is elusive because sediment storage can occur across the entire system, including in the erosion zone, leading to potential buffering over short distances. Reconstructing erosion and transfer zones in deep time ($\geq 10^7$ yr) is challenging as a result of incomplete or no preservation of these sediment-routing segments. Characteristics of deep-time sediment production and transfer areas can be interpreted by employing provenance tools, detrital mineral analysis, or application of empirical relationships based on modern systems, and tested with conceptual, analytical, and numerical models.

Implications of signal detection over noise are of paramount importance to interpreting the stratigraphic record of Qs variability. Internal, or autogenic, dynamics of sediment transport, transient storage, and release can introduce noise, lags, and/or completely mask the signal of interest. Experimental and theoretical work has shown that a Qs signal can be passed to a basin and preserved in the stratigraphic record when its period is similar to or exceeds the characteristic response time of the sediment-routing system, but this is valid only if their periods or magnitudes are also larger than the characteristic timescale of autogenic sediment transport fluctuations (Jerolmack and Paola, 2010). Moreover, autogenic 'shredding' is potentially more significant if the length scale of interest, e.g., the system size, is larger than the advection length scale of a particle of sediment (Ganti et al., 2014). Field studies focused on the coarse-grained sediment fraction in small, tectonically active sediment-routing systems of southern California (Fig. 6A) have shown that millennial-scale climate forcings are represented as a measureable signal in the stratigraphic record of the deep-sea segment (Fig. 7) (Romans et al., 2009; Covault et al., 2010). These systems are reactive (*sensu* Allen, 2008b) and comprise sediment-routing segments in close proximity: erosion zones are located adjacent to short transfer zones and offshore confined basins that make up the accumulation zone. Noise over historical timescales can be especially problematic as the observational window is small and the number of signals potentially large (e.g., Sommerfield and Wheatcroft, 2007).

Timescale of observation is fundamentally important for signal analysis in sedimentary systems. Temporal aspects discussed in this review include the duration and period of forcings, the resolution of chronologic tools with which to evaluate Qs (Fig. 2), and preservation into the record. At historical timescales the signal of interest is commonly an individual event, such as an earthquake, flood, or storm. At longer timescales, shifts in the rate and style of sedimentation in the cumulative record can be related to longer-period forcings. We devoted much of our treatment of signal propagation at the intermediate timescale (10^2-10^6 yr) because we consider this to be a critical temporal range in which to understand these dynamics at the scale of entire sediment-routing systems. The shorter-duration end of this range can be linked to direct observation and measurement and the longer, million-year end of this range can serve as a bridge to deep time. The intermediate timescale is the timescale of global climate cycles and the timescale at which climate and tectonic forcings overlap. Moreover, this is the timescale at which meso-scale stratigraphic architecture and high-frequency stratigraphic cycles are created. Our subdivision of timescales in this review is only to aid communication of dominant aspects within each timescale, but we emphasize that sedimentary system research strives to integrate across these timescales.

Our review emphasizes the role of sediment supply, yet we acknowledge the role of accommodation fluctuations in the accumulation zone as an important forcing of stratigraphic patterns (e.g., Anderson et al., this volume). Deciphering the relative contributions of sediment supply versus accommodation changes to the creation of stratigraphy has been discussed for at least a century (e.g., Grabau, 1913; Barrell, 1917) and examined via modeling studies for decades (e.g., Jervey, 1988; Allen and Densmore, 2000; Paola, 2000; Armitage et al., 2013). The nature of signal generation in the sink and potential up-system propagation effects (e.g., Voller et al., 2012) deserve further attention. Deconvolving the various external forcings from each other and from the products of internal dynamics encoded in the geologic record remains a prime challenge in Earth science (National Research Council, 2010).

This review provides a set of conceptual and practical tools for reaching informed interpretations of landscape dynamics from the stratigraphic record. These tools include stratigraphic and sediment-routing system characterization, sediment budget determination, geochronology, detrital mineral analysis (e.g., thermochronology), comparative analog approaches, and modeling techniques to measure, calculate, or estimate the magnitude and frequency of sediment supply signals compared to the characteristic response time of the sediment-routing systems. However, significant research challenges remain, which we distill into four research directions:

1. **Improved documentation of sediment production, transfer, and accumulation rates in natural systems.** The propagation of a signal through a system can be characterized as a phase velocity and, thus, knowledge of time is required. Research aimed at developing new chronometric techniques and studies applying existing techniques in novel ways should continue to be a focus in Earth surface dynamics research. Such work should include the study of and linkage between erosion, transfer, and accumulation zones across a spectrum of system sizes and morphologies as well as across a range of timescales. Within this context, the question of to what degree patterns of stratigraphy (i.e., patterns documented in the absence of absolute chronology) reflect process rates should be further explored.

2. **Grain-size partitioning and signal propagation.** What is the size distribution of sediments exiting the erosion zone as a function of forcings? Our discussion of how sediment supply signals propagate through a system and how they might be preserved in the stratigraphic record is simplistic in that only bulk mass balance is

considered. The variability in transit distances of different grain-size classes for a given forcing might result in the fractionation of catchment-generated signals with certain grain-sizes temporarily stored along the transfer zone. Attempts to capture downstream grain-size fining and to invert it to time-averaged grain-size trends in stratigraphy are promising but still in their embryonic stage (e.g., Whittaker et al., 2011; Michael et al., 2013). The link between forcings in the erosion zone and the probability density function of the produced grain-size distribution remains a major unknown and constitutes a fundamental input of Qs propagation models at all temporal and spatial scales.

3. **Integration of experimental and modeling approaches with natural systems.** Many of the modeling efforts reviewed here are based on diffusion assumptions and/or empirical relationships, neither of which truly model sediment transport. Several approaches are currently tackling sediment transfer more explicitly, including using increasingly complete physics of water flow and sediment transport (e.g., Delft3D, broadly labelled Computational Fluid Dynamics, e.g., Lesser et al., 2004); Reduced Complexity Models (such as cellular automata, e.g. Murray and Paola, 1997; Liang et al., 2014); and several other modeling approaches as part of CSDMS (Community Surface Dynamics Modeling System) (Syvitski, 2008). These efforts are complementary and promising, but it is important to maintain a link with natural systems in order to properly assess the appropriate degree of complexity for which model predictions can be compared to data from natural systems. Additionally, scaled-down physical experiments are

contributing valuable insights regarding the timescales of dynamics in depositional landscapes (e.g., autogenic channel avulsion frequency; Paola et al., 2009). However, how such timescales relate to natural-system timescales remains an open question. These experimental approaches must continue to strive to integrate with observation/measurement-based approaches and vice versa.

4. **Integration of particulate transfer dynamics with solute transfer and other geochemical signals.** The denudation of landscapes in erosion zones is the sum of physical and chemical products that are moved down system. Sedimentary archives containing chemical precipitates can be reliable recorders continental weathering as well as atmospheric and oceanic chemistry and have been used to detect climatic and/or oceanographic signals. However, studies that integrate geochemical signal analysis with the concepts and tools for signal propagation as a function of particulate transfer are rare. Additional work combining the particulate and (bio)geochemical perspectives to examine sedimentary system response to environmental change (e.g., Foreman et al., 2013) are necessary to develop a comprehensive understanding of the broader Earth surface system.

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This review of signal propagation concepts in sedimentary system analysis is purposefully broad in scope and, as a consequence, neither provides a comprehensive treatment of every relevant issue nor does it refer to all studies that have contributed. We hope that readers appreciate the challenge of preparing such a review. BWR, JAC, and

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Figure Captions

Figure 1: (A) Schematic portrayal of a sediment supply (Qs) signal from the erosion zone and how that signal propagates through the system. The leftmost Qs signal represents as measured at the exit of the erosion zone and for simplicity is the same as the original forcing of interest. The transfer zone Qs signal is measured within the transfer zone at some distance from exit of erosion zone and the rightmost signal represents that which reaches the accumulation zone and is an input for the stratigraphic record. Dashed lines refer to Qs signal in up-system segment(s) to illustrate that a signal can be modified during propagation. (B) 2-D profile of a generic sediment-routing system emphasizing erosion, transfer, and accumulation zones (potential for intermediate to deep time

stratigraphic preservation in yellow) and important controls of tectonics (including earthquakes), climate (including storms), base level, and anthropogenic factors. Part B modified from Castelltort and Van Den Driessche (2003).

Figure 2: Overview of three timescales of investigation, some of the chronometric tools with which to constrain process rates, and periods of some of the forcings discussed in this review. Dashed lines at the top emphasize the continuum among the timescales. Temporal range of 'orogenic cycles' from DeCelles et al. (2009). Effective dating range of chronometric tools from Walker (2005).

Figure 3: (A) Topography and drainage network of Eel and Mad river catchments, northern California, and bathymetry of the continental margin. Red star denotes location of shelf core x-radiograph shown in (B). (B) X-ray image of shelf reflects bulk density. Light colors (lower bulk density) interpreted as 1995 flood deposit (Sommerfield and Nittrouer, 1999; Wheatcroft and Borgeld, 2000). (C) Map of Eel-Mad sediment-routing system showing catchment area, areal extent of coastal floodplain, and shelf depocenter (yellow). Red star denotes location of shelf core image shown in (B). (D) Historical timescale sediment budget of the Eel-Mad sediment-routing system showing: 1) there is negligible onshore storage, 2) the shelf stores ~30-50% of the budget, and 3) the remainder moves to the canyon and continental slope. Budget estimations from Sommerfield and Nittrouer (1999) and Warrick (2014).

Figure 4: (A) Topography and drainage network of Ganges, Brahmaputra, and Meghna rivers and bathymetry of shelf, Swatch of No Ground submarine canyon, and part of the

Bengal submarine fan system. Red star denotes location of core record shown in (B). (B) Core from upper canyon showing variation in mean grain size with time compared to storm record from eastern Bengal shelf. Data is from core 96 KL as reported in Michels et al. (2003). (C) Map of Ganges-Brahmaputra-Bengal sediment-routing system showing catchment area, the large delta plain area, shelf depocenter (yellow) and the Bengal submarine channel-levee system. Red star denotes location of core record shown in (B). (D) Historical timescale sediment budget of Ganges-Brahmaputra-Bengal sediment-routing system showing that almost one-third of the budget is stored on the delta plain, ~40% accumulates in the shelf depocenter, split between the topset and foreset regions, and the remaining ~30% is delivered to the canyon and Bengal submarine fan. Budget estimations from Kuehl et al. (2005) and references therein.

Figure 5: The ratio between the timescale of a perturbation (Tp) and the characteristic equilibrium timescale (τ) of the considered system describes the system response to forcing (after Beaumont et al., 2000; see also Allen, 2008b). A forcing of water discharge (Qw) and a response of sediment supply (Qs) are shown for (A) a reactive response when response time is much shorter than timescale of forcing and (B) a buffered response when response time is longer than timescale of perturbation.

Figure 6: Examples of natural sediment routing systems examined at intermediate timescales. (A) Small and tectonically active system, Peninsular Ranges and Continental Borderland of southern California (Covault et al., 2011); (B) Large and tectonically quiescent system, Texas coastal plain and western Gulf of Mexico (Hidy et al., 2014;

Pirmez et al., 2012); (C) Large and tectonically active system, Indus River and Indus submarine fan (Clift et al., 2014). Indus River floodplain extent from Milliman et al. (1984).

Figure 7: (A) Map showing two southern California sediment-routing systems, each with negligible onshore sediment storage at millennial timescales and correspondingly rapid transfer to offshore submarine fan systems. (B) Alkenone sea surface temperature (SST) proxy for the California coastal region showing a drying trend in the early Holocene followed by the development of the modern El Niño-Southern Oscillation (ENSO), which is known to be sensitive to increased SSTs (Barron et al., 2003). (C) Radiocarbon-constrained weighted-average sediment accumulation rates from Hueneme and Newport submarine fans (Romans et al., 2009; Covault et al., 2011) showing a general correlation of sediment supply to the basin to precipitation regime and, thus, propagation of climatic signal to the stratigraphic record.

Figure 8. Summary of signal propagation from source to sink at intermediate timescales (10^2-10^6 yr). The figure considers uplift and climate signals in the erosion zone and their transformation and propagation into sediment supply (Qs) signals in both the transfer and accumulation zones. The gray shaded regions indicate the cases in which the original forcing has been faithfully transmitted to the accumulation zone. (A) An uplift signal with a period (t) > 50 kyr is transformed into a Qs signal with same amplitude and period by the erosion zone and transmitted by a short fluvial segment but buffered by a long transfer zone unless t > 100 kyr (e.g., Castelltort and Van Den Driessche, 2003). (B) An

uplift signal with t < 50 kyr is transformed into a buffered Qs signal by the diffusive catchment dynamics in the erosion zone (Allen and Densmore, 2000) and further buffered by diffusion in the transfer zone (e.g., Castelltort and Van Den Driessche, 2003). (C) A climate signal of water discharge (Qw) with t > 100 kyr in the erosion zone yields a buffered Qs response signal due to catchment dynamics (Armitage et al., 2013). (D) A climate signal of Qw with 10 kyr < t < 1 Myr in the erosion zone yields an amplified Qs response signal due to catchment dynamics (Godard et al., 2013). (E) Climate signal of Qw with t > 0 kyr in erosion zone is transferred to the fluvial system where it is transformed into an amplified Qs signal by river transport (Simpson and Castelltort, 2012). (F) In the absence of signals coming from the erosion zone, autogenic signals can emerge out of the transfer zone due to the internal dynamics of the fluvial system (e.g., compensational stacking in channelized systems, Hajek et al., 2010), with periodicities and amplitude poorly constrained. Wang et al. (2011) suggest autogenic periodicities of the order of 100 kyr for the Mississippi river delta. (G) Regardless of origin, Qs signal input to the transfer zone may be 'shredded' by the internal dynamics of sediment transport system when their period and amplitude fall within the range of 'morphodynamic turbulence' (Jerolmack and Paola, 2010).

Figure 9: Conceptual diagram of crystallization age of detrital minerals (e.g., zircon U-Pb geochronology) as indicator of sediment-routing system connectivity, which can be used to aid reconstruction of sediment supply history from deep-time stratigraphic archives.

Figure 10: Conceptual diagrams of two different types of lag times to be calculated with combinations of cooling ages and depositional age. (A) Schematic cross section depicting trajectory of a particle through cooling via erosional exhumation followed by transport and deposition. (B) Lag time type A is the difference between higher-temperature cooling age (e.g., crystallization age) and lower-temperature cooling age of a single grain (e.g., zircon) and represents exhumation from depth. (C) Lag time type B is the difference between a cooling age and depositional age and represents the time from closure temperature depth to surface exhumation plus transport (and transient storage) in the sediment-routing system prior to deposition (adapted from Rahl et al., 2007). A consistent Lag time type B calculated through a stratigraphic succession indicates steady erosion whereas departures from that indicate changing erosion rates through time. Note that time within partial annealing/retention zones as well as sedimentary and/or tectonic burial can complicate simple lag time determinations.

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