9.23 Fluvial Terraces

Frank J. Pazzaglia
Department of Earth and Environmental Sciences
Lehigh University
Bethlehem, PA 18015
610.758-3667; 3677 FAX
fjp3@lehigh.edu

Synopsis

Fluvial terraces are landforms and deposits that integrate tectonic, climatic, and geomorphic processes at the watershed scale. Terraces attest to a fundamental unsteadiness in the rate of vertical incision of a channel as it carves its valley. The most common sources of that unsteadiness are vegetative, geomorphic, and hydrologic responses to climate, which for the Quaternary, are dominated by 100-k.y. glacial-interglacial cycles; however, the precise response depends much on watershed substrate and the climatic, tectonic, and base level setting. When coupled with numeric ages, terraces can be used as a stratigraphic and geodetic marker to interpret correlation and process rates.

Keywords

Alluvium, base level, climate, fill, glacial-interglacial, grade, incision unsteadiness, longitudinal profile, river, strath, tectonics, terrace, tread.

Introduction

Fluvial terraces are common, globally-distributed landforms underlain by alluvial deposits that decorate the flanks of rivers valleys in a wide range of climatic and tectonic settings (**Fig. 1.**). Long utilized by humans for their strategic location along major rivers, flat surfaces conducive to agriculture and cities, and economic placer deposits, terraces constitute a core observable for geologists interested in fluvial processes, active tectonics, and Quaternary paleoclimatology. When coupled to Quaternary geochronology that provides numeric ages, terraces can be dated and used as a geodetic marker to infer tectonic and climatic process rates. This chapter presents a synthesis of important fluvial terrace studies and discusses models of their formation with the hope that terraces can be viewed as accessible to a wide range of geoscientists addressing diverse geologic problems.

The term “river terrace” and speculation on its origin first begins to appear in the modern physical geography literature in the latter part of the 19th century (Gilbert, 1877; Leslie, 1878; Dana, 1881; Nelson, 1893; **Fig. 2**). Before that, the engineering, mining, and natural science literature referred to flat landforms in river valleys in the general terms of topography useful for canals, irrigation, and agriculture (Hitchcock, 1824). The mineral value of terraces as a source of placer deposits (Schumm, 1977) and aggregate for construction and roads is a recent outgrowth of industrialized societies. Early scientific papers on terraces have provided us the legacy of terms we use to describe these landforms. For example, strath is a term used to describe the erosional base of a terrace (Bucher, 1932). Tread and berm (Campbell, 1929) are terms proposed to describe the terrace surface. River incision has long been recognized to be synonymous with terraces and an enduring, if not always correct legacy of early studies is that terraces are a landform caused by rock uplift. Similarly and from a glacial geology point of view, the early literature is rich in examples of river terraces as outwash at or near glacial margins (e.g. Carney, 1907; Penck and Bruckner, 1909).

This chapter is not intended to be a review and reconciliation of the vast literature on the subject of terraces. Rather, it is intended to provide, by means of example, a coherent synthesis and framework by which terrace genesis can be understood and terraces can be used to solve geologic problems. Virtually all of the ideas and concepts synthesized in this chapter are not the original work of the author, unless specifically referenced as such. Rather, the chapter leans heavily on the contributions of giants in the field, namely Bill Bull (Bull, 1991; 2007) and Stan Schumm (Schumm, 1977; Schumm et al., 1987) who have written outstanding textbooks on the subject.

The chapter will qualitatively describe terraces and the river valleys that contain them before offering some more quantitative approaches to modeling their genesis and age. Certain rivers and river valleys will figure prominently, not because they are the single best example of the points being made, but rather because they are good examples that have strongly shaped the experience of the author. These include the Susquehanna River of Pennsylvania, the Rio Grande and its major tributaries in New Mexico, the Clearwater River of Washington State, and the Reno River in the northern Apennines of Italy. The author is indebted to many students and professional colleagues who have engaged in observation and discussion of the ideas presented herein. The summary models presented are not agenda-driven, but rather represent a platform for continued research and scholarship on the topic.

Fluvial terrace definition and general description

In North America, the general definition and use of river terraces in the modern geologic literature is anchored in the foundational work of G.K. Gilbert (1877) and has been strongly influenced by the career contributions of Bill Bull and his students (Bull, 1991), Stan Schumm and his students (Schumm, 1977; Schumm et al., 1987), and Dale Ritter and his students (Ritter, 1967; Ritter et al., 2002). Similar, formative contributions to river terrace scholarship are found in Europe (Maddy, 1997; Maddy et al., 2001; Van den Berg, 1996 and subsequent studies; Bridgland, 2000), Asia (Ohmori, 1991; Sugai, 1993; Pan et al., 2003), Africa (Deacon and Lancaster, 1988; Partridge and Maud, 1987; de Wit et al., 2000), and in Australia (Bishop et al., 1985; Young and McDougall, 1993). The Fluvial Archives Group (FLAG) has been active over the past decade in documenting the global record and genesis of river terrace stratigraphy (Vandenberghe and Maddy, 2000). The terms and definitions adopted in this chapter can be sourced in these formative contributions.

River terraces are landforms underlain by alluvial deposits. They are distinguished from floodplains in that they are not inundated frequently, if ever, during floods. Floodplains may be terraces in the making, but need not necessary be preserved following river incision. When preserved, terraces are fixed as geomorphic markers in the river valley. They are expressed as elongated benches parallel to and above the river channel. Depending on the setting, terraces may adorn the flanks of a river valley far above the channel, extending all the way to and capping the drainage divide. More commonly, there is an elevation in the drainage basin below which terraces are preserved and above which they have been removed by erosion. Topographic data in the form of widely available topographic maps, air photos, remote-sensed satellite imagery, and increasingly available high resolution digital elevation models are used to map, describe, and correlate terraces.

Full characterization of the terrace landform and its underlying alluvial deposits are necessary in order to use them as geomorphic markers to address geologic and related problems (**Fig. 3**). Considering both, a terrace and terrace alluvium are best described as allostratigraphic units (Blum, 1993; Pazzaglia and Brandon, 2001) where the bounding unconformities are the flat terrace surface, called a tread, the erosional terrace base, called the strath, and the intervening slope connecting to a river floodplain, adjacent terrace or valley hillslope, called a riser. The bounding unconformities are all geomorphic markers that can be treated as paleogeodetic indicators of crustal deformation over geologic time scales (Bull, 2007). The enveloped alluvial deposits harbors paleohydrologic information regarding hillslope erosion, fluvial transport, and watershed-scale climatic processes (Bull, 1991; Blum and Valastro, 1994; Maddy et al., 2001). Paleohydrologic information is also contained the longitudinal gradients of the strath and the tread (Paola and Mohrig, 1996).

Terraces are commonly organized into two main categories based on their morphology and thickness of alluvial deposits (**Fig. 3**). A strath terrace is characterized by a distinct, sub-horizontal erosional base, commonly but not exclusively carved into bedrock, with a relatively thin alluvial cover. Bedrock is any substrate into which the river valley is being developed, including soft, poorly-consolidated sediments that can make the distinction between bedrock and river alluvium difficult to distinguish. The thickness of the alluvial cover is a long-discussed criteria in the definition of a strath terrace. The important points are that the alluvium is thin with respect to the size of the river, river valley, and watershed and the alluvium has played a role in the carving of the strath. This can be an alluvial cover, for example, of < 1 m for a small river with a drainage area < ~100 km^2, or 3-5 m for a large river that draining > ~10,000 km^2. The point is that for large rivers, frequent discharges may have a scour depth of 3-5 m and entrain an alluvial bed 3-5 m thick and if so, that is the alluvial cover the river can abandon upon the strath.

A strath terrace facies model shows the strath to be buried by medium- to thickly-bedded, imbricated, coarse gravel and sand commonly held to be the thalweg bedload (**Fig 4a**). This gravel fines up to thin- to medium-bedded, interbedded and commonly cross-stratified gravelly to pebbly sand. Strath terrace treads have 0.1 – 1 m of thin or massively-bedded sand and silt that conformably or unconformably overlies the sand and gravel. These fine-grained caps can be of overbank or aeolian (loess) origin.

The strath is a key, but still poorly understood result of river erosion into bedrock. Strath formation processes are more fully explored below, but much research remains to be completed on the topic. Observations in many rivers of different scale and tectonic and climatic setting indicate that bedrock channel erosion processes conspire to carve a sometimes erosionally-decorated, but generally sub-horizontal channel floor. The carving of wide straths, a largely lateral incision process that widens valley bottoms, and their subsequent abandonment by vertical incision that deepens the river valley are the ultimate causes for the formation of strath terraces.

A fill terrace is characterized by a thick alluvial deposit that (formerly) buried the river valley bottom (**Fig.3**). The base of a fill terrace can be sub-horizontal like a strath or it can also have buttress unconformities and other features common to buried topography with high local relief. A valley buried by a fill terrace commonly has an irregular basal contact in its center and more planar basal contacts with the valley walls on its flanks such that the cross-sectional shape looks like a valley with wings. Unlike a strath terrace, the thick alluvial deposits do not represent syn-bedrock erosion and bedload transport. Rather, the alluvium represents a period of valley-wide aggradation and the lifting of the river off of bedrock to become a purely alluvial channel. The valley fill is preserved as a terrace when the channel ceases aggrading and once again seeks out its former elevation, resulting in vertical incision of the alluvial deposits. These characteristics of the fill terrace have long been recognized to indicate rapid and profound environmental changes in a watershed, processes that will be more fully presented and discussed below.

A fill terrace facies model is characterized by common meandering and braided stream textures and bedforms including coarse thalweg gravel, laterally-accreted sand and gravel longitudinal, transverse, and point bars, and vertically accreted crevasse splay deposits, lacustrine silt and clay, and overbank mud (**Fig 4.**). The vertical stacking of these facies vary considerably with examples of coarsening-up, fining-up, and a coarse-fine-coarse stratigraphy known to prevail in different river valleys. Fill terrace alluvium varies in thickness and scales to the size of the watershed and river, but even for large watersheds of > ~10,000 km^2, mobile alluvial beds and channel scour depths are rarely in excess of 5 m so alluvial thickness greater than that are usually associated with fill terraces. Similar to strath terraces, fill terraces common are capped by fine grained overbank and aeolian deposits 0.1 – 1 m thick.

Features and processes of rivers and watersheds that contain terraces

An obvious but commonly overlooked observation is that not all river valleys contain terraces, so studies of terrace stratigraphy and genesis are necessarily biased towards that subset of fluvial and watershed-scale processes that conspire to make and preserve terraces. Similarly, the vast majority of terrace sequences are in the context of modern rivers and river valleys reflecting processes dominate during the Quaternary, or even more temporally restrictive, of the Holocene. The Quaternary, with its characteristic high-amplitude and high-frequency glacial-interglacial climate unsteadiness, may be diagnostic, but not representative of prevailing late Cenozoic environmental conditions. We may live in a world where river terraces are so common because they represent a landform reflecting the unique conditions of Quaternary environmental change (**Fig. 5**).

Terraces are an expression of contrasting erosional and depositional processes, played out at the watershed scale, and orchestrated by base level, hydrology, hillslope sediment flux, and sediment caliber. The most basic conclusion that terraces demand is that these conditions and processes are not steady or necessarily uniform. It is possible that autogenic processes within the confines of a watershed can drive the unsteadiness, but is more common that base level and climate, forces external to the watershed, dominate the unsteady erosion and depositional processes. Base level is a combination of eustasy and rock uplift with the former mostly lacking for more continental settings far from the coast. Climate refers not only to mean annual temperature and moisture, but also to precipitation intensity, seasonality, and inter-annual variability. For lofty mountainous settings and large watersheds climate is typically non-uniform, resulting in spatially variable terrace genesis and preservation.

Because of their close genetic link to base level and climate, terraces embody core, if not necessarily high-resolution paleohydrologic and paleogeodetic data, specific examples of which follow below. There are, however, common errors that studies make in trying to use terraces for these purposes. In terms of paleohydrology, it is erroneous to conclude that terrace genesis is uniformly in phase with a prevailing environmental setting. Rather, it is more common that terraces represent transient responses to an environmental change. Furthermore, the phase of that response can depend on aridity, substrate, and watershed relief.

Similarly, when used as a paleogeodetic marker, most studies focus on the terrace tread due to the relative ease in its identification in topographic and imagery data. Unfortunately, the tread and strath are not necessarily parallel surfaces, an observation that is particularly true for fill terraces (**Fig. 6**). For example, the along-valley gradient of the tread of a fill terrace is fixed by the hydrology of the alluvial channel that deposited it, rather than the modern channel for which it may have little in common. The elevation and gradient of a terrace tread contains components apportioned to the paleohydrologic conditions during deposition as well as post-depositional changes in base level and crustal deformation. There is no unique way to resolve this under-determined problem.

Perhaps even more erroneous is that many studies fail to recognize that hillslope and pedogenic processes begin to modify terrace treads as soon as they are formed (Ritter and Miles, 1973; Pelletier, 2008). Colluvial wedges shed from adjacent hillslopes and older terrace risers obscure the fluvial deposits introducing or erasing relief not related to the hydrologic conditions during tread deposition or post-depositional base level changes (**Fig. 7**). This problem is particularly acute for active tectonic studies that depend on subtle changes in terrace gradients to infer tectonically-drive changes in base level and river incision. For these reasons, the terrace strath emerges as the better geodetic marker with comparable paleohydrogic and base level conditions.

Terraces can be inset into the flanks of valleys as paired or unpaired landforms, terms that stem from their cross-valley discordant or concordant elevations respectively (**Fig. 3**). The preservation of flights of terraces, particularly when they are paired, seems to always imply progressive valley narrowing. When paired terraces are as wide, or wider than the modern valley bottom and situated directly across the valley from one another, valley narrowing is an inescapable conclusion. However, paired terraces are commonly staggered in their longitudinal position and need not indicate valley narrowing given the lateral mobility of the valley bottoms.

Terraces tend to be thickest, widest, and best preserved at tributary stream junctions. This observation suggests that the sediment delivered by the tributary actively both widens the main stem valley bottom and provides the alluvium to build a thick, preservable terrace deposit. Wide valleys obviously help in terrace preservation because the terraces themselves are wider, but it is the dominate hillslope process in the valleys that plays the ultimate role in determining where terraces can be preserved. Landslides, earthflows, and creep are hillslope processes that all contribute to removing terraces from valley walls. Terraces are similarly dissected by fluvial process, particularly where broad, impermeable treads impede infiltration. Pedogenesis, particularly where it is influenced by loess deposition, drives infiltration rates lower with the accumulation of translocated clays and/or soluble salts ($CaCO_3$) in soil B-horizons so as terraces age, they also tend to produce more runoff, enhancing the fluvial dissection process (Wells et al., 1987).

Graded and steady-state stream profiles and their relation to terraces

Equilibrium, or graded, river longitudinal profiles (Gilbert, 1877; Mackin, 1948; Leopold and Bull, 1979; Knox, 1975) are concave-up, a shape that stems from the downstream increase in discharge and decrease in bedload mean grain size (Sinha and Parker, 1996; **Fig. 8**). Graded streams delicately adjust reach-scale gradients so that the sediment being delivered from hillslopes is transported by the prevailing discharge (Mackin, 1948; Baker and Ritter, 1975; Bull, 1979). Graded profiles develop for bedrock and alluvial channels as it is the sediment being transported, rather than the substrate being incised that determines local channel gradient. A graded profile fixes its gradient to regional base level, the base level of erosion of Bull (1991) and it is implied, rather than specifically stated, that a steady base level is a necessary condition for a graded profile. There is little process unsteadiness in a perfectly graded profile so there are no occasions for the form of the profile or the river valley to change. Other than stochastic processes like channel meander migration, there are few opportunities for terraces to be created and preserved. However, it has long been noted that valley widening by wide strath carving occurs when the stream is in the graded condition for a long period of time (Gilbert, 1877; Mackin, 1937).

The graded profile needs to be distinguished from the steady state profile of which it is a subset. The steady state profile is fixed to the base level reference frame such that it neither rises nor falls in elevation with respect to a changing base level. For example, a river valley can be aggrading in response to subsidence (base level rise), a seemingly clear violation of the conditions for grade, but the elevation of the channel need not change as long as the rate of aggradation is balanced by the rate of subsidence. Similarly, a river valley can be actively deepening in response to rock uplift (base level fall), but as long as the incision is balanced by the rock uplift, the channel does not change its elevation and remains fixed to falling base level reference frame.

Like the graded profile, a perfect steady-state profile implies steady processes, resulting in uniform valley and channel forms, with little opportunities for terrace production. Environmental and base level changes impart transient processes to the steady-state profile, and a corresponding response in valley and channel form. It is this unsteadiness in process that presents many opportunities to create and preserve terraces.

The form of the steady-state profile can be readily quantified given the growing access to high-quality digital topography. Many recent studies have shown that when the upstream contributory area of a channel reach is plotted against channel slope of that reach a power law-relationship results where the downstream rate of gradient decrease, or concavity (θ), is the slope of the regression and the overall fall from the channel head to mouth, or steepness (k_s), is the regression's y-intercept (Hack, 1957; **Fig 8a**),

$$S = k_s A^{-\theta} \quad (1).$$

Rock uplift (Snyder et al., 2000; Kirby and Whipple, 2001), climate (Roe et al., 2002; Zaprowski et al., 2005), and bedrock (Duvall et al., 2004; Spagnolo and Pazzaglia, 2005) are known to impact profile concavity and steepness. These effects can be modeled when the bedrock channel erosional process is assumed to be detachment-limited and proportional to stream power,

$$E = KA^m S^n \quad (2),$$

where E is the rate of bedrock erosion, K is a proportionality constant that accounts for substrate resistance, A is basin area upstream of a given reach, S is channel reach slope, and m and n are scaling factors related to basin hydrology and base level of erosion respectively. Equation (2) stems from the fundamental assumption that bedrock channel erosion scales with shear stress on the bed, shear stress can be expressed as the reach-scale product of flow depth and channel gradient, that discharge grows linearly with basin area, and that channel width increases downstream as the square root of the discharge (e.g. Snyder et al, 2000).

The steady-state profile requires that vertical river incision (E) be perfectly balanced by rock uplift (base level fall, U) such that the elevation of the channel is steady (dz/dt =0),

$$\frac{dz}{dt} = 0 = U - E \quad (3).$$

Substituting equation (2) into equation 3 and solving for S yields

$$S = \left(\frac{U}{K}\right)^{\frac{1}{n}} A^{-m/n} \quad (4).$$

Equations (1) and (4) have the form of a straight line when plotted on logarithmic axes. Tectonics, expressed as the rate of rock uplift (U), is reflected in the profile steepness (k_s), and basin hydrology is primarily reflected as the profile concavity ($-m/n = \theta$).

Deviations from the idealized steady-state profile are generally taken to be transient adjustments to a tectonic or environmental change and are expressed in quantifiable changes in concavity and steepness. For example, steep channel reaches, or knickpoints (**Fig. 8b**), result from local offset across a fault or base level fall and have negative concavity (convexity). Channels in arid regions do not experience downstream growth in discharge and tend to have straighter, less concave profiles than those in wetter regions. It is the transient changes in these characteristics in the establishment of the steady-state profile that drives terrace genesis. Concavity changes in particular are accomplished by incision, a process that lowers the channel gradient at the reach scale, and aggradation, a process that steepens the channel gradient at the reach scale.

Terrace straths and treads represent paleo-steady-state profiles that need not be necessarily parallel. The base level and environmental conditions that prevailed during strath cutting are not necessarily the same when the river aggraded to the level of the tread, particularly in the case of a fill terrace. Parallelism, convergence, or divergence in strath, tread, and channel profiles results from the evolution in profile concavity or steepness as well as in any non-uniform tectonic process that accumulates in progressively deformed terrace geomorphic markers. Three general cases are possible considering the effects of unsteadiness in profile concavity or steepness in the comparison of river and terrace profiles (**Fig. 9**). The first case is a perfect steady-state profile that maintains a characteristic concavity and steepness with steady, uniform incision. Such a profile is formed and maintained by steady base level (or sea level) fall and uniform rock-uplift. Terrace straths formed under these conditions will have longitudinal profiles that are parallel to the active channel.

The second case is when the channel concavity is being increased or decreased because of non-uniform or unsteady rock uplift. Straths that converge or diverge with the channel are possible in this case, depending on their location with respect to the incision or aggradation accommodating the concavity change. A setting that illustrates this point involves a growing rate of uplift in the headwaters of a river valley situated on the fore-limb of a growing anticline and the subsequent tilting and fore-shortening of the entire valley towards the river mouth (Kirby and Whipple, 2001; Miller et al., 2007). In this case, the channel concavity and steepness increase as fold growth accelerates. Straths carved along a paleo-, less concave profile will be observed to diverge away from the channel where incision has increased concavity.

Similarly, it has been demonstrated through observation (Zaprowski et al, 2005; Dethier, 2001) and modeling (Roe et al., 2002; Wobus et al., in press) that channel concavity increases with increasing peak annual discharge and precipitation intensity (**Fig. 10)**. These results are particularly noteworthy given that climate change directly impacts discharge and precipitation characteristics. Global climate over the past ~ 25 m.y. is synonymous with a perceived increase in river erosion and delivery of sediment to the world's oceans (Zhang et al., 2001). It is possible, but remains to be unequivocally demonstrated, that valley incision, narrowing, and terrace formation is a global response to these recent environmental changes.

The third major case for profile concavity changes is attributed to glacio-eustasy, the headward migration of knickpoints, and related changes in hillslope sediment flux. Watershed responses to glacial-interglacial environmental changes are modulated by latitude, altitude, eustasy (Blum, 1993), and bedrock (Wegmann and Pazzaglia, 2009) which results in unsteady production of discharge and sediment (Schumm, 1969; Bull, 1991). These changes in the source of the upper parts of the profile are balanced by base level in the sink where the transported sediment is deposited. Valley aggradation steepens the local transport slope in response to greater quantity and caliber of sediment, reducing concavity in the process. In this case, a terrace strath will have a profile generally less steep than the terrace tread because the environmental conditions defining the steady-state profile at the time of strath cutting when the channel was in contact with bedrock are different than those that existed at the time of tread formation when the channel was purely alluvial.

Knickpoints dramatically, but locally steepen a long profile and commonly reduce concavity to negative values (convexity). Knickpoints are transients in a river profile that represent the step-wise transfer of base level lowering from the lower to upper part of a long profile (Whipple, 2001; Crosby et al., 2007; **Figs. 8 b**). The concavity of the steady-state lower and upper profiles astride the knickpoint is the same, but the elevation of the lower profile has fallen the base level reference frame. In this case, the paleo –steady state profile is represented by a terrace that is everywhere parallel to the modern profile downstream of the knickpoint.

Particularly in the lower part of a river profile that drains to the sea, base level change can have dramatic effects on profile concavity. The glacio-eustatic range has been +12 m to -140 m since the middle Pleistocene and the long-term average sea level has been -74 m for the entire Pleistocene (Pazzaglia and Brandon, 1996). Generally speaking base level fall has been accompanied by incision and the profile change described above for knickpoint migration has occurred. Conversely, base level rise generally results in aggradation; however, the precise response of the shoreline to accommodation space, sediment flux, and discharge vary geographically and in concert with the ratio of the gradient of the river and gradient of the shelf (Schumm, 1993; Woolfe et al., 1998; Pazzaglia and Brandon, 2001). For example, when the gradient of the shelf is the same or less than the gradient of the river, a eustatic fall results in a large horizontal translation of the coast, forcing the river channel to prograde out across the low-gradient shelf (**Fig. 11**). In order to maintain a transport gradient for its bedload, the channel will have to rise as it progrades, resulting in aggradation at the coast in response to a eustatic fall. This response is well documented for coastlines with high sediment fluxes (Bull and Knuepfer, 1987; Pazzaglia and Brandon, 2001; Nesci and Savelli, 2003) and carries with it an important cautionary note that the treads of such fill terraces are the result of bedload transport gradient steepening, not post-depositional tectonic deformation.

All of the scenarios described above describe time-depended transients that work to change the steady-state profile of a river. As a result, terraces, although physically contiguous and mappable as allostratigraphic units, are also diachronous. The degree of diachroneity in the carving of a strath or abandonment of a tread is best illustrated by a terrace created by the migrating knickpoint example (Zaprowski et al., 2001). In this case which is detailed below, both the terrace strath and tread are born at the knickpoint, and become progressively older downstream. In contrast, a concavity change driven by a climatically-induced increase in discharge would be more uniformly distributed along the entire profile resulting in a terrace tread with less diachroneity. In either case, it is generally assumed in studies that use terraces to infer river incision and aggradation processes that the amount of time represented in the creation of the terrace by some transient change in the steady-state profile is short compared to the amount of time that the profile remains in the steady state condition.

Strath Genesis

Straths are the geomorphic expression of valley bottoms carved by graded or steady state longitudinal profile (**Fig. 12**). Lateral incision processes that widen the valley bottom during graded or steady-state profile conditions carve wide straths that can be preserved following subsequent vertical incision (Mackin, 1937; Montgomery, 2004). It is important to point out that narrow, local straths, perhaps only the width of the channel, are always being carved by lateral incision at the reach-scale, but these straths are rarely mantled by alluvial deposits and preserved as a terrace. Rather, it is only when a channel attains the steady-state profile and can maximize valley widening that the lateral accommodation space for alluvial deposition and terraces is created. The width of the river valley bottom, like the concavity and steepness of the longitudinal profile, is fixed by mean discharge, resistance of rock substrate, and sediment flux (Pazzaglia et al., 1998). Like the rate of vertical incision, the rate of lateral incision fixing the valley bottom width is not steady, but varying in response to changing environmental conditions. Terraces are naturally associated with those environmental conditions when lateral incision processes that widen the valley bottom are maximized and sustained.

The carving of a river strath by lateral incision is an understudied process with respect to its importance in terrace genesis. What is known is that a river both lowers its bed and widens its valley, carving a strath, through several processes including abrasion, plucking, and cavitation (Whipple et al., 2000; Chatanantavet and Parker, 2009), all of which are influenced by the substrate being carved and wetting and drying cycles (Montgomery, 2004). Abrasion is accomplished by the alluvium in transport that acts as tools to wear on the bedrock substrate both vertically and laterally in the channel. Erosion due to abrasion (E_a) is typically taken to scale with shear stress on the bed (τ_b),

$$E_a = \tau_b^{5/2} \qquad (5).$$

Field and analogue experiments have demonstrated that there is an optimum amount of tools that will maximize the abrasion process (Sklar and Dietrich, 1998). When the channel is under capacity, there are not enough tools to accomplish abrasion and similarly, when the sediment flux to the channel is high, the channel is lifted off the bedrock, not allowing moving tools to contact the bed. A channel at capacity maximizes abrasion through a combination of abundant tools that are also in frequent contact with the bed. In fact, the common, thin alluvial cover of strath terraces is consistent with the idea that the alluvium represents a mobile bed whose thickness is the effective scour depth of the prevailing flows. Abrasion tends to smooth the bed, reducing local relief and generating a sub-horizontal strath.

Like abrasion, plucking and cavitation work to erode protrusions and exploit weaknesses in the channel bed that may be following fractures or other heterogeneities in the bedrock. Cavitation is a highly non-linear process dependent on velocity, flow depth, and stage of development of the cavitation vortex. It is typically expressed as proportional to a critical velocity (U_c) that is dependent upon flow depth, the Reynold's Number, fine sediment concentration, atmospheric pressure, and dissolved air content,

$$E_c = (U - U_c)^q \quad (6),$$

where q is a poorly constrained value less than or equal to 7.

Plucking in a purely detachment-limited channel is typically described as a loosening rate (L) of fracture-bounded bedrock fragments by a Meyer-Peter-Mueller bedload transport rate equation (q_s) of the form

$$L = q_s^p = (\tau_b - \tau_c)^{5p/2} \quad (7)$$

where the basal shear stress is defined by the depth-slope product, τ_c is the critical shear stress, typically defined by the Shield's criterion for gravel-sized clasts, and p is an unknown quantity thought to be close to unity. The plucking process is maximized where substrates are particularly sensitive to repeated wetting and drying cycles that open fractures (Montgomery, 2004) and effectively reduces the τ_c term in equation (6). Wetting and drying cycles occur seasonally and are characterized by a nearly perfectly-horizontal interface – the water surface – rising and falling through the channel bottom. Some rock types, most notably bedded sedimentary rocks, are particularly susceptible to weakening through wetting and drying cycles. Plucking controlled by a horizontal interface, along with abrasion, are the two most logical reasons why straths are so flat.

In the same way that vertical incision is modeled as scaling with stream power (ω), lateral incision (E_l) is thought to also depend on stream power (Hancock and Anderson, 2002).

One study suggests that lateral incision scales with the square root of stream power (Suzuki, 1982) and this makes sense considering that the width of a channel is one expression of lateral incision and channel width scales as the square root of discharge.

$$\omega = \rho g Q S \qquad (8),$$

$$E_l = \omega^{0.5} \qquad (9).$$

Field observations of straths and their alluvial cover reveal several clues that help indicate the prevailing environmental conditions during their genesis. First, the grain size of many strath terraces is coarser than the grain size of the corresponding active channel (Baker, 1974; Pierce and Scott, 1982). Second, the channel patterns inferred from terrace alluvial facies are different than the corresponding active channel (Baker, 1974; Ritter, 1967; Nesci and Sevelli, 1990; Coltorti et al., 1991; Vandenberghe, 2003). Lastly, paleo-basin-wide erosion rates indicated by terrace alluvium are several times greater than those inferred by the sediment flux of the corresponding channel (Schaller et al., 2002; 2004; Fuller et al., 2009), a point that is expanded upon below. Collectively, these observations suggest that the ratio of discharge (essentially stream power) to sediment flux (Schumm, 1969) modulate the strath carving process. When the lateral incision is maximized, wide straths are cut. Changes in the rate of vertical incision is not a necessary corollary to a change in the rate of lateral incision, but several field studies either measure or infer that vertical and lateral incision are inversely related (Bull, 1991).

Channel pattern changes, strath cutting, and subsequent terrace formation has been observed in flume studies where discharge and sediment flux has been intentionally or unintentionally varied (Meyer, 1986; Gardner, 1983; Wohl and Ikeda, 1997; Frankel et al., 2007; **Fig. 13**). Autogenic complex responses (Womack and Schumm, 1977) are highlighted in these studies and underscore the importance of non-externally forced unsteadiness in natural lateral and vertical river erosion. It has been argued that fill-cut terraces form primarily by this mechanism (Schumm et al., 1987).

For example, erosion of a simulated bedrock reach in an otherwise alluvial sand-bed flume has been observed by building a 30-cm wide rib of glacio-lacustrine clay within the sand (Frankel et al., 2007; **Fig. 13a**). Under constant base level and discharge conditions and no upstream introduction of sediment, an alluvial channel is allowed to develop and stabilize in the flume including the reach underlain by the simulated bedrock material. Erosion is introduced by instantaneous base level fall at the river mouth, creating a knickpoint that migrates rapidly head ward through the unconsolidated sand. The knickpoint is halted, temporarily at the bedrock reach as it steepens and begins to erode the clay, carving a narrow canyon. As soon as the knickpoint and downstream base level fall signal begins to propagate past the bedrock reach, the upstream alluvial reach responds instantaneously to the base level fall, delivering a pulse of sediment to the bedrock reach. This pulse of sediment quickly buries most of the canyon, steepens the transport gradient, and begins wearing away laterally at the unburied walls of the canyon, carving a strath. Eventually, base level stability is re-established at the bedrock reach and sediment delivery from the alluvial reach upstream wanes. Still under constant discharge, the ratio of discharge to sediment flux increases for the bedrock, resulting in a renewed phase of incision and canyon cutting. As that base level fall is propagated upstream to the alluvial reach, a second pulse of sediment is delivered to the canyon, partially burying it, and initiating the carving of a lower, inset strath.

Similar alternation in vertical and lateral erosion for a bedrock reach have been observed in the field for steep channels in the rapidly uplifting eastern Central Mountain Range of Taiwan (Hartshorn et al., 2002). Channel bedrock surveys of the gauged LiWu River with evenly-spaced recessed and permanent benchmarks record the channel widening and deepening over a two year period. The river has a very high sediment load of 10^7 metric tons annually and is affected by very high peak daily discharges approaching 10^3 m^3/s associated with wet-season typhoons, interspersed with more modest base flow peak annual discharges 2 orders of magnitude smaller. Measurements repeated before and after typhoons show that most of the vertical incision occurs during base flow conditions when the channel has a greater discharge to sediment ratio. Typhoons increase both discharge and sediment to the stream, but their ratio decreases and it is during these brief, but intense storms that the channel widens, presumably because it has access to sediment and high-energy abrasion, plucking, and cavitation that can be directed against the valley walls.

A drainage basin-scale natural experiment that combines features of the Frankel et al (2007) flume and Hartshorn et al. (2002) field studies has been documented for Pancho Rico Creek in the coast ranges of California where a recent stream capture has changed the discharge to sediment flux ratio (Garcia, 2006; **Fig. 14**). Pancho Rico Creek is underlain and formerly drained ~ 100 km^2 of friable, homogeneous marine sedimentary rocks of the Pancho Rico Formation. The valley of the trunk stream is flanked by two wide, paired strath terraces mantled with alluvium exclusively of Pancho Rico Formation provenance, the younger of which has a Late Glacial Maximum (LGM) OSL age of ~ 20 ka. These terraces are only preserved where the Pancho Rico valley has accumulated at least 60 km^2 of upstream catchment.

The Pancho Rico drainage instantaneously grew to ~150 km^2 following a capture event shortly after creation of the LGM strath terrace, triggering a cycle of incision and terrace formation in the former Pancho Rico trunk valley. Pancho Rico Creek has incised 60 m or more and carved four or more unpaired strath, fill, and fill-cut terraces inset into the LGM terrace level (**Fig. 14**). The capture not only provided Pancho Rico Creek with more discharge, but also with more sediment of a different caliber and hardness. As in the pre-capture Panco Rico valley, the post-capture valley carved terraces at the point where there is at least 60 km^2 of upstream catchment. Post-capture provenance of the terrace alluvium includes resistant Franciscan rock types that are much better abrasive tools than the friable Pancho Rico Formation sedimentary rocks and form much better alluvial mantles on straths, protecting them from erosion. The post-capture Pancho Rico valley has recorded terraces with a much higher fidelity than the pre-capture valley, presumably because of changes in drainage area and sediment flux. The Pancho Rico Creek study shows that there is a critical threshold in drainage area (discharge and stream power) where terraces can form in a valley and that the number, type, and preservation of those terraces are influenced by the provenance of the alluvium in transport and the discharge – sediment ratio.

A one-dimensional, physically-based numeric model provides one of the most comprehensive and insightful illustrations of the how the discharge – sediment ratio controls strath genesis (Hancock and Anderson, 2002). The study takes advantage of a rich set of observations of terraces in the basins of the Rocky Mountain foreland, many of which can be physically traced to heads of outwash, as the boundary conditions and input for the model. The basic assumptions and rules of the model are patterned after Tucker and Slingerland (1997) and differ from earlier numeric approaches (Boll et al., 1988; Veldcamp and Vermeulen, 1989; Veldcamp, 1992; Veldkamp and van Dijke, 2000; Bogaart and van Balen, 2001; Tebbens et al., 2001) in that the results stem from the interactions of a few, simple, physically-based erosion and sediment transport laws. The model accounts for discharge (conservation of mass), sediment transport and grain size that scales with shear stress, and vertical and lateral erosion of bedrock that scale with unit stream power. The lateral erosion law also takes into account the width of the valley wall as narrow valleys will have frequent contact with a laterally-eroding channel whereas already wide valleys with a wide floodplain will have less frequent contact with the channel.

The model is stable under boundary and input conditions representative of the Wind River of western Wyoming. With no change in discharge or sediment yield, the model predicts a gently concave-up longitudinal profile and a valley with no terraces after 800,000 model years. Most of the long profile and valley change by vertical and lateral incision occur early in the simulation and decrease monotonically. All other simulations where discharge, sediment yield, or both are forced periodically or variably produce unsteady vertical and lateral incision, straths, and strath terraces. A key finding in these simulations is that the ratio of vertical to lateral incision rate is time dependent and is determined almost exclusively by the rate of vertical incision. The model shows that both vertical and lateral incision rates scale similarly with changes in discharge such that there is little change in the vertical-lateral incision rate ratio (**Fig. 15a**). In contrast, lateral incision rates are unchanged while vertical incision rates fall to near zero as sediment supply is increased. This key finding indicates that the potential lateral incision rate of a stream is fixed by discharge, and modulated by rock-type resistance, specific erosion process, and width of the valley. Vertical incision rates are slowed presumably because high sediment yields lift the channel off of the bedrock and insulate it from the various erosion processes. In the meantime, the lateral incision rate continues unabated, carving a wide strath during the time of arrested vertical incision.

The model produces stair step terraces of uniform width and height when forced by period changes in discharge or sediment yield (**Fig. 15b**). A more realistic flight of terraces of variable width and height is produced when the discharge and sediment yield input is scaled to the well known oxygen isotopic curve of glacial-interglacial climate cycles (**Fig. 5**). In these simulations the channel spends 50-90% of the time laterally carving straths of variable width, and brief periods of time vertically incising transforming the strath into a strath terrace. Given the variable strength (amplitude) of glacial-interglacial transitions, not all climate cycles generate a terrace that is preserved in the valley. Only when a tilt of the valley orthogonal to the river axis is imposed is the accommodation space created to carve and preserve a terrace for each climate cycle. Such a tilt occurs in nature for many rivers, including those that drain the northern Apennine mountain front in northern Italy. Here, rivers like the Reno River at Bologna have a high fidelity record of unpaired terraces on their western valley walls that appear to record every major and minor climate cycle since the middle Pleistocene (Picotti and Pazzaglia, 2008; Wegmann and Pazzaglia, 2009), a feature not observed where valleys are not being tectonically tilted and terraces are paired.

Knowing the amount of time a river spends carving a strath by lateral incision is limited by geochronologic methods for dating terrace alluvium. Although there are several excellent studies that have dated and correlated long terrace records to glacial-interglacial climate cycles (van den Berg and van Hoof, 2001), the uncertainties and paucity of numeric ages makes it difficult to know precisely how much time the river spends carving the strath and how much time is spent vertically deepening the valley. Most studies conclude that the times of vertical incision must be short because long-term rates of incision that average over several glacial – interglacial cycles are almost always slower than short-term rates that account for only the incision since the LGM (Gardner et al., 1987; Mills, 2000).

A study of Holocene terraces with an unusually large number of radiocarbon numeric ages in the terrace alluvium (Wegmann and Pazzaglia, 2002; **Fig. 16**) sheds some light on the vertical – lateral incision rate ratio and the veracity of the Hancock and Anderson (2002) modeling results. These terraces are preserved in the Clearwater River drainage on the western flank of the Olympic Mountains of Washington State. The Holocene rate of incision for this river is 0.8 – 3 mm/yr, some three times more rapid than the long-term (Pleistocene) rate of vertical incision of 0.5-0.8 mm/yr. There are at least three major Holocene strath terraces that presumably represent times of little to no vertical incision, so the average Holocene rates must be eclipsed by even faster vertical incision rates when the channel is not laterally carving a strath. The strath ages are interpreted from 38 calibrated radiocarbon dates collected from their overlying alluvium. When plotted in a stacked probability diagram that accounts for calibrated age uncertainties, the age range for a given strath and its alluvium is revealed (**Fig. 16d**). Even when accounting for the presence of wood and charcoal in the basin with inherited ages of up to 1,000 years, the ages cluster around three strath forming intervals ~3000-4000 years in duration. For a given terrace, the ages are skewed, clustering at the minimum age with tails towards the older ages. These age distributions are interpreted as three pulses of sediment that were delivered to the channel, slowing the rate of vertical incision and allowing lateral incision processes to carve a wide strath. The Clearwater River needs 3000-4000 years to transport this sediment through its valley bottom, during which time the strath is being carved. When the sediment source is eliminated and/or cleared out of the valley, the channel again is in frequent contact with the bedrock and vertical incision proceeds. The data indicate that periods of vertical incision last ~1000 years, or only 25% of the time represented by the strath carving episodes.

Similarly, recent measurement of vertical incision rates for large rivers draining the Atlantic Slope across hard, metamorphic rocks of the Appalachian Piedmont suggest short pulses of rapid vertical incision (Reusser et al., 2004; 2006) balanced against much longer periods of lateral incision and the carving or wide straths (**Fig. 17**). The incision rates for these large rivers such as the Susquehanna and Potomac river, are very slow, on the order of 10 m/m.y. averaged since the middle Miocene (Pazzaglia and Gardner, 1993). The bedrock floors of these streams are straths nearly a kilometer wide in places, punctuated by falls and rapids of debated origin (Pazzaglia et al., 2006). Cosmogenic exposure ages of the straths in at Holtwood Gorge on the Susquehanna River and Great Falls on the Potomac River indicate vertical incision rates of 500 m/m.y., a result incompatible with the long-term measurements or even the amount of relief in the gorges that does not exceed 200 m. The more likely explanation is that the vertical incision rate of these streams is very slow or zero for long periods of time when the channels are carrying sediment, presumably during most of the Pleistocene when glacial and periglacial processes are responsible for sediment input to the fluvial system. The removal of this sediment source, a characteristic of the Holocene and presumably Holocene-like interglacial times, results in the short, but rapid pulses of vertical incision.

Both the Clearwater and Atlantic Piedmont river results are in good agreement with the Hancock and Anderson (2002) model predictions. A background lateral incision rate scaled to discharge is always present in a river valley and capable of carving a strath. Wide straths that are likely to be preserved as strath terraces are carved when vertical incision is slowed by an increase in sediment load. Terraces are the evidence that vertical incision is unsteady. These strath terrace genesis models and observations conclude that vertical incision may occur as little as only 10 – 25% of the time in life history of a given valley's development.

Terrace Genesis

Field studies, analogue models, and numeric models collectively have provided a set of observations that guides a general, synthetic model for terrace genesis. These studies can be cast into two broad categories: those that focus on process (e.g. Schumm, 1987; Bull, 1991) and those that focus on stratigraphy and correlation (e.g. Vandenbergh and Maddy, 2000). This is not to imply that these and related studies do not contain elements of process and stratigraphy because as it turns out they do and both are fairly critical in construction of the general model. The purpose of this section is to explore in detail a few, but outstanding studies that integrate process with stratigraphy to stand as the best examples of river valley development over geologic time scales.

Terraces and Tectonics

Crustal deformation and rock uplift have always occupied a central role in the geomorphic thinking of terrace genesis, a legacy of the strong connection between sedimentary rocks and the mountains that are eroded to produce them (Vandenbergh and Maddy, 2000). Active tectonics influence terrace genesis insofar that rates of rock uplift are unsteady. For example, there are many studies of coastal marine terraces with upstream fluvial equivalents that record impulsive base level fall (rock uplift) as a phase of fluvial incision inset into relict landscapes of low-relief (e.g. Gardner et al., 1992; Molin et al., 2004; Sak et al., 2004). Such impulsive uplift is easy to demonstrate in a coastal setting where there is a direct connection to sea level and there are independent estimates of sea level change. Although similar impulsive uplift has been argued for intracratonic regions, the distant connection to sea level leaves the door open for other, local, processes, namely hydro-climatologic (described below) or local stream capture to also be considered.

The Laramide Rocky Mountains, a series of basement cored ranges and intervening sedimentary basins embedded in a high-standing plateau, has a strongly-debated post-orogenic rock uplift history (McMillian et al., 2006; Riihimaki et al., 2007; Carroll et al., 2008) that helps illustrate the complicated drivers of intracratonic river incision. Terraces throughout this region record unsteady incision during the Pleistocene. The coincidence of a major terrace forming event during the Lava Creek B caldera eruption of Yellowstone ~640 ka imprinted a widespread datum against which incision can be measured (Dethier, 2001). Since the middle Pleistocene, rivers have been deepening their canyons at a rate of 50 to 300 m/m.y., which is significantly faster than the erosion rates measured for uplands between major drainages of 7.6±3.9 m/m.y. (Small et al., 1997). The long-term rate of vertical incision is generating relief in the Rocky Mountain landscape. The high rates are focused in regions of highest mean elevation and tend to decrease downstream along any given drainage. Active rock uplift, locally or epeirogenic at 50-300 m/m.y. cannot be ruled out as the cause of the incision, but if the Rocky Mountain foreland attained its high mean elevation in the geologic past it can also be argued that Pleistocene climates and attendant glacial-interglacial cycles are driving all or a good part of the incision. Recent modeling of long profile adjustments to these climate changes suggests that the greatest changes in profile concavity would be felt in the headwaters reaches of major rivers, a result consistent with observation (Wobus et al., in press).

The ambiguity in discerning real unsteadiness in rock uplift is rooted in the fact that tectonic processes rates typically do not accelerate or decelerate as rapidly as changes in climate. Active rock uplift provides the potential for incision and production of accommodation space, but there are few, if any studies that can clearly demonstrate tectonic unsteadiness using a data set independent of river incision. Geodetic measurements are not of long enough duration to project deformation unsteadiness over geologic time scales and thermochronologic data do not have the resolution to discern unsteady cooling (erosion). Yet, rivers that traverse the world's great orogens have terraces that demand incision unsteadiness. Numeric dating of straths (Burbank et al., 2006), strath terraces (Lave and Avuouc, 2000), and fill terraces (Pan et al., 2003) in the Himalaya, for example, repeatedly conclude that terrace ages coincide closely with known times of major climatic change.

Ironically, it is in slower-deforming regions of the world where unsteadiness in rock uplift may be primarily responsible for terrace genesis. Rivers draining passive margins have broad, upland terraces of great antiquity (late Cenozoic) that cannot be directly correlated to known climatic or eustatic events (Partridge and Maud, 1987; Gunnell et al., 2003; Howard et al, 1993; Pazzaglia and Gardner, 1994; Young and McDougall, 1993; Matmon et al., 2002). Given the slow rate of landscape development in these settings, changes in the rate of rock uplift driven by dynamic mantle flow or related processes (Moucha et al., 2008) can plausibly drive unsteady incision and terrace genesis. A strong case for this style of tectonically-controlled terrace genesis is argued for a pulse of uplift in the Ardennes (Van Balen et al., 2000) and Rhenish Massif (Meyer and Stets, 2002) as well as a continent-scale onset of more rapid incision since the middle Pleistocene in Europe (Gibbard and Lewin, 2009).

Base level and knickpoints

Local base level fall influence on terrace genesis deserves consideration apart from the base level fall of a region driven by regional rock uplift. A local base level fall can be cause by an process, tectonic or geomorphic, that introduces a step into a long profile, called a knickpoint, that migrates upstream and declines in relief in response to specific bed shear stress conditions directly above and below the step (Gardner, 1983). The result of migrating knickpoints are terraces parallel to the steady state profile that are born at the knickpoint and increase in age downstream proportional to the rate of knickpoint migration. The process triggers a wave of incision that propagates through a basin (Garcia, 2004) or orogen (Safran, et al., 2005) and may trigger surface – crustal process feedbacks. The time scale of this process (Whipple, 2001) can play out over the late Cenozoic as has been proposed for the Black Hills of South Dakota (Zaprowski et al., 2001), or can be much more rapid as has been documented for the headwaters of the Yellow River (Harkins et al., 2007) and Red River (Schoenbohm et al., 2004).

Terraces and Climate

The weight of observational data and modeling results strongly favors Quaternary climatic change and associated hydrologic responses as the main driver of vertical incision unsteadiness in rivers, as recorded by river terraces. This is not a new idea but rather represents a theme that has evolved for over a century (reviewed in Vandenburghe, 2003). The fact that river process and form changed as a result of unsteady hydrology was a well accepted concept by the middle part of the 20th century (e.g. Leopold et al., 1964). The specific lateral and vertical incision responses to discharge and sediment, first discussed in Schumm (1969) have emerged as the key driving forces that have guided subsequent studies.

Periodic forcing of terrace formation

The most well-accepted idea regarding climatic influences on terraces links their formation to glacial-interglacial-driven changes in watershed hydrology and sediment flux. This idea is very appealing because it provides an explanation for unsteady geomorphic processes and a model for temporal correlation by comparison to the now mature marine oxygen isotopic record of Quaternary climate (**Fig. 5**). Orbitally-driven changes in solar insolation are well-represented in the marine paleoclimatic record and of these, the 100 k.y. Milankovitch cycle has been the dominant modulation for the past ~1 Ma causing major advances and retreats of high-latitude and high altitude ice sheets. Synchronous fluvial response at the continental scale to this forcing is an idea that has influenced subsequent studies for the past 40 years (Vita Finzi, 1969; 1976).

Among the most complete and best studied record of fluvial response to late Cenozoic (Pliocene and Quaternary) climate this comes from the Maas River of northern Europe (van den Berg, 1996; van den Berg and van Hoof, 2001) where 21 paired and unpaired fill terraces spanning 2 million years of incision are preserved. The Maas has incised ~100 m over this time for an average incision rate of ~50 m/m.y. Ages for the Maas terraces include numeric and relative dating methods that have been compiled by several independent studies. Generally speaking, the terrace alluvium aggraded during periods of cold, glacial climate, whereas river incision occurred during times of warm interglacials or interstadials. Many rivers located in the temperate mid-latitudes are known to have followed similar aggradation – incision responses to glacial and interglacial climates respectively; however, as will be illustrated below, this general model is modified by geography, relief, and rock-type.

Growing refinement of numeric techniques for dating river terraces has revealed synchroneity in periodic terrace formation for rivers draining wide regions as well as the degree of diachroneity in the formation of an individual terrace. One of the first studies to document diachroneity using radio carbon ages demonstrated that a thick fill terrace in Cajon Creek, California aggraded over a ~9,000 year period in the Late Pleistocene and Early Holocene (Weldon, 1986). The aggradation began in the headwaters of the drainage and progressed downstream, resulting in an alluvial fill with a lens-shaped longitudinal profile. The strath and tread of this fill are separated by 150 m of alluvium in the middle reaches of the stream. That separation decays to less than 10 m at the upstream and downstream extent of the terrace. Similar downstream younging of fill alluvium is documented for terraces along rivers in southern Spain (Fuller et al., 1998) and the U.S. Gulf Coast (Blum and Valastro, 1994). The amount of time corresponding to aggradation at a given reach illustrates how much of the Quaternary and its attendant sediment production processes and hydrology has been characterized by glacial-like, rather than interglacial-like conditions. For example, an individual alluvial fill for east-drainage rivers in central Italy are known to have basal radiocarbon ages of 36,000 yrs and tread radiocarbon ages of 15,000 years indicating 21,000 yrs of filling (Coltorti et al., 1991, Wegmann and Pazzaglia, 2009).

At least 13 major alluvial episodes of variable amplitude and duration corresponding to 13 fill terraces are recognized in the Mediterranean basin (Macklin et al., 2002). Latitude may play a role in determining the onset of an alluvial phase, but given the diverse size, tectonic, hydrologic, and rock-type characteristics of the studied basins, the degree of synchroneity in these terrace alluvial records is remarkable. It is notable that the rivers in all of the above cited studies are currently situated in Mediterranean climates and likely experienced similar corresponding Mediterranean glacial climates. Modern (interglacial) Mediterranean climate is characterized by hot, dry summers, cool moist winters and a dominance of trees, shrubs, and vine vegetative cover on the landscape (Watts et al., 2000). In contrast, glacial Mediterranean climate is characterized by cold, dry conditions with a dominance of herbal vegetative cover.

The dominant vegetative cover, its role in regolith production, and its response to changing climate appears to be the key controls on sediment yield and hydrologic unsteadiness in Quaternary river basins (Bull, 1991; Vandenberghe, 2003). Beyond the Mediterranean basin there are several studies that demonstrate how sediment flux responds to Quaternary environmental and vegetative changes. The timing of the response occurs in phase or shifted in phase to the environmental change depending on latitude, altitude, and presence or absence of glaciers (**Fig. 18**). Generally speaking, sediment flux to the channel peaks in high latitude, high altitude, and glaciated basins, like those northern Europe (van den Berg and van Hoof, 2001), in the lofty mountains of New Zealand (Bull and Knuepfer, 1987), or the northern Rocky Mountains (Ritter et al., 1993), during peak glacial conditions. Channels exceed capacity during peak glacial conditions resulting in valley aggradation and alluvial fills. In these settings, interglacial periods like the Holocene are marked by low sediment yields, channel incision, and creation of fill terraces. The sediment is both produced and transported off of hillslopes when they are not mantled with vegetation and periglacial and glacial erosion processes are maximized.

In contrast, hillslopes in semi-arid, Mediterranean, or lower-latitude environments liberate their sediment at the climate and environmental transition from glacial to interglacial conditions, such as the late Pleistocene-Holocene transition (Bull, 1991; Blum et al., 1994; Ritter et al., 2000). The sediment is produced on hillslopes when they are vegetated during the effectively more moist glacial climates, and then rapidly stripped as the climate becomes progressively drier and perhaps more seasonal. Hillslopes in arid or hyper-arid environments respond similarly to times of environmental change, but the hillslope sediment flux has been documented to be both more variable, as shown by watersheds in the Mojave desert (Bull, 1991) and delayed until well into the interglacial period, as shown by watersheds in the hyper-arid Sinai (Bull and Schick, 1979).

Some terrace studies highlight responses to periodic forcing that cannot easily be ascribed to climatic or geomorphic drivers, but rather seem to indicate that there may large-scale complex responses and lag-times involving sediment storage and transport (Schumm and Rea, 1995). Notable among these is the detailed stratigraphy, anchored with numeric ages, of the Colorado River and its tributaries through the Grand Canyon reach (Anders et al., 2005; Pederson et al., 2006; **Fig. 19**). As in many terrace investigations in the American west, the Colorado River stratigraphy shows major terrace forming periods coincident with major glacial, interstadial, and interglacial climates of the past 200 k.y. However, the thickest and most extensive terrace represents ~ 40 m of aggradation between 90 and 120 ka, during MIS 5d-b (the M4 terrace in **Fig. 19**). Climatic unsteadiness is known from this time, but it is typically taken to be less than that associated with MIS 6, 4, and 2 which represent major glacial episodes (Winograd et al., 1992, 2006) well recorded in the Colorado River headwaters in the Rocky Mountains. It is possible that the weathering-limited nature of hillslopes local to the Colorado Plateau have a peculiar response to climate, fixed by their particular lithology, and produce more terrace-forming sediment during interglacials or interstadials. Alternatively, it is possible that the hypsometry of canyon country promotes a peculiar complex response involving long lag times of how a disturbed natural system responds to a climatic forcing. In this case, sediment produced by the MIS 6 glaciation is not available for transport and axial stream alluviation until 40 ka later during the specific hydro-climatic conditions of MIS 5d-b.

Evidence for and effects of unsteady sediment yields

The unsteadiness in hillslope sediment flux in response to environmental change is a conceptual model, well based in field observations of terrace alluvium, but directly testable using cosmogenic techniques. There are at least two excellent recent studies that appear to strongly support the model. Cosmogenic analysis of terrace alluvium can determine paleo-erosion rates two ways (Gosse and Phillips, 2001). If the age of a terrace is independently known using some other numeric method, then the amount of cosmogenic ^{10}Be in quartz can be interpreted as being proportional to the cosmogenic dosage that quartz received when it was in the hillslope environment. High concentrations of ^{10}Be indicate a long residence time, and hence slow hillslope erosion whereas low concentrations of ^{10}Be indicate short residence times and rapid erosion. In the case where the age of the terrace alluvium is not independently known, two cosmogenic isotopes with different decay rates, typically ^{10}Be and ^{26}Al can be used to simultaneously determine alluvium age and paleo-erosion rate. This two-isotope method can only be applied for deeply-buried alluvium (>10 m) that has been effectively shielded from any further cosmogenic dosage since deposition.

The paleoerosion rate of the alluvium preserved in the Maas and related terrace alluvium of the Allier and Dore rivers has been reconstructed for the past 1.3 my (Schaller et al., 2002; 2004). The studies reveal basin wide erosion rates during terrace aggradation in the late and middle Pleistocene of ~60-80 m/my. which are 3 times greater than rates obtained from Holocene terrace aluvium and modern gauged sediment fluxes. Early Pleistocene terraces older than 0.7 m.y. indicate paleoerosion rates ~25-30 m/my. The relatively slow Holocene erosion rates are interpreted as being characteristic for an interglacial climate when vegetation is able to hold sediment on hillslopes and physical processes like freeze-thaw are minimized. The faster Pleistocene erosion rates are interpreted as being consistent with cold-climate physical processes and the lack of any stabilizing vegetative cover on the hillslopes. The increase in Pleistocene erosion rates at 0.7 Ma is consistent with the onset of 100-k.y. dominant modulation of glacial-interglacial climates. It is also coincident, at least for this part of northern Europe, with the onset of slow, epeirogenic rock uplift.

Similarly, the paleoerosion rates measured from late Pleistocene strath terrace alluvium in northern California are twice as large as those obtained from comparable Holocene strath terrace alluvium and the modern channel alluvium (Fuller et al., 2009). The Holocene rate of vertical incision is three times as fast as the rate of Holocene erosion, a conclusion also reached by terrace studies in the Clearwater basin in Washington State (Wegmann and Pazzaglia, 2002; Belmont et al., 2007). These results are consistent with the idea that vertical incision rates are unsteady and temporally restricted to the relatively short Holocene-like interglacial climates during the Quaternary. Northern California is known to have been wetter in the late Pleistocene during the time of terrace alluvium aggradation (Adam and West, 1983) and it is hypothesized that the wetter conditions caused greater frequency of mass movements at that time. Larger late Pleistocene sediment yields are consistent with a slowing of the vertical rate of incision, lateral widening of the valley bottom, and deposition of terrace alluvium.

Renewed vertical incision in a valley bottom that has been widened and mantled with alluvium requires a triggering mechanism that may be linked to an intrinsic threshold such as channel gradient (Patton and Schumm, 1975) or an extrinsic threshold driven by a changing climate and basin hydrology. Such a threshold response is demonstrated by rapid vertical incision when channel bed alluvium is mined for gravel aggregate. The alluvium sets the channel gradient and with its removal, particularly the coarse size fraction, that gradient is too steep for the remaining, fine-grained bed load resulting in incision and reduction of the channel gradient. Streams draining the foothills of the northern Apennines in Italy experienced extensive gravel bed mining in the decades following the second world war and display dramatic evidence of this incision process. The incision was initially localized at the gravel extraction sites on the Po Plain, generating a knickpoint that continues to rapidly migrate upstream through poorly-consolidated bedrock. The resulted is 10's of kilometers of steep-walled, arroyo-like channels up to 10-m deep. In this example, the mining is a proxy for the natural removal of channel bed alluvium, or breaching of an armored bed perhaps by larger or more flashy discharges, or an oversteepening of the channel following a long period of aggradation. Both explanations appeal to unsteadiness in discharge and sediment yield that on longer time frames drive long periods of valley bottom widening and alluviation punctuated by rapid, but short periods of vertical incision.

The prevailing rock-type that underlies a watershed determines how it will respond hydrologically and geomorphically to glacial-interglacial climate change (Bull and Schick, 1979; Kelson and Wells, 1989) and reflected in the resulting terrace type and stratigraphy (Kelson, 1986; Wegmann and Pazzaglia, 2009; **Fig. 20**). The same rivers draining the northern Apennines described above in the gravel bed mining example are flanked by up to nine distinct strath terraces. Only 100 km further south along the central Italian Adriatic coast, similar-sized rivers are flanked by four-five thick fill terraces. The tectonic and climatic setting for these rivers is very similar, but notably, the drainage basins are underlain by different rocks. The northern Apennine basins are underlain mostly by siliciclastics including a friable Tertiary sandstone whereas the central Apennine basins, particularly their headwaters, are underlain by carbonates. Glacial climate conditions associated with MIS 3 and 2 triggered a phase of alluviation in the central Apennine valleys that persisted for 20 k.y. creating a single thick alluvial fill that was exposed as a fill terrace terrace following post 19 k.y. incision. Over the same time period and presumably in response to the same climatic conditions, three strath terraces were carved in the northern Apennine basins. Erosion processes in the northern basins are dominated by shallow mass movements, the frequency of which is sensitive to moisture. In contrast, erosion in the central basins is dominated by freeze-thaw physical weathering of carbonate bedrock, producing cm-sized chips that form ubiquitous stratified slope deposits throughout the landscape (Coltorti and Dramis, 1995). Apparently the central Italian rivers were overwhelmed with this periglacial debris throughout MIS 3 and 2, resulting in continuous aggradation. In contrast, the northern Italian rivers oscillated between three periods of relatively drier climate, few landslides, bedload undercapacity, and vertical incision and relatively wetter climate, frequent landslides, bedload capacity, and the carving of wide straths.

Similar conclusions are reached for the channel and terrace long profiles of streams draining rugged mountains in northern New Mexico (Kelson, 1986; **Fig. 20b**). Here, streams experience higher bankfull discharge and discharge per unit area at all flow stages for watersheds of comparable size underlain by crystalline rocks in comparison to watersheds underlain by sedimentary rocks (Kelson and Wells, 1989). Rivers draining the crystalline rocks have numerous strath and fill terraces that are arranged sub-parallel to a straight long profile incising at the same rate as the axial drainage (Rio Grande) determining regional base level. Rivers draining the sedimentary rocks, in contrast, have few terraces with variable gradients above convex channel long profiles unable to maintain incision with the regional base level. Rock-type controlled watershed hydrology and hillslope erosion rate apparently determine the stratigraphic fidelity of the resulting terraces.

Holocene alluvial and bedrock valleys

The impact of glacial-interglacial climate change has left Holocene rivers with a legacy of sediment and valley forms that can be divided into two major types: (1) river valleys with few if any Holocene terraces and channels that flow in alluvium, commonly gravelly sand, that is several meters to 10’s of meters thick, atop an older, bedrock-carved valley and (2) river valleys with several bedrock-floored strath terraces and channels that flow directly on bedrock, transporting a thin alluvial cover (**Fig 3**). At first glance it may appear that valleys with Holocene strath terraces are associated with regions undergoing rapid rock uplift such that incision rates are high and there is accommodation space to carve Holocene terraces (e.g. Lave and Avuouc, 2000). Upon closer examination there are many examples from tectonically active settings where valleys do not have Holocene strath terraces or examples of both terraced and unterraced valleys occur in close proximity to one another. Such examples demand a careful examination of how terrace genesis responds in phase or shifted in phase with the suspected unsteadiness in climatic driven discharge and sediment yield.

The Rio Grande and its tributaries in the American Southwest drain an actively deforming continental rift that regionally stands high (Roy et al., 2004) and has experienced some of the highest rates of river incision in the western United States since the middle Pleistocene of approximately 200 m/my (Kelson, 1986; Formento-Trigilio and Pazzaglia, 1997; Dethier, 2001; Wisniewski and Pazzaglia, 2002). The Rio Grande is a braided, sand-bedload stream that flows atop the poorly-consolidated rift-basin fill Santa Fe Group of fluvial volcaniclastic, alluvial fan, and eolian origin. The stratigraphy of the Rio Grande valley is well known from detailed mapping (Lambert, 1968; Connell, 2008) and numerous water well logs (Hawley and Kennedy, 2004). Holocene alluvium in the Rio Grande floodplain and channel is approximately 5 m thick and composed of gravelly sand and sand with a rift provenance. It sits unconformably atop a coarse sandy gravel paleo-braided stream alluvium approximately 30 m thick with a southern Rocky Mountain provenance that unconformably overlies the Santa Fe Group bedrock. The contact with the bedrock is irregular in the axis of the Rio Grande valley, but more planar and strath-like towards the valley margins. Despite the difference in grain size between the Holocene sand and underlying pre-Holocene gravel, the longitudinal gradient of the gravel-Santa Fe Group contact is parallel to the long profile of the modern Rio Grande (Hawley and Kennedy, 2004).

The Rio Grande Valley is flanked by four major fill terraces with Santa Fe Group basal contacts, stratigraphy, grain size, and provenance characteristics identical to the alluvium buried in the axial valley (Connell, 2007). The stratigraphically youngest terrace is a late Pleistocene minor fill terrace 3-6 m thick with a tread that stands 15 m above the Rio Grande channel. The next older terrace is a major fill terrace 52 m thick with a tread that stands some 48 m above the

Rio Grande. This thick fill is known to date to MIS 6 because it is interbedded with basalts dated to 156±29 k.y. (Peate et al., 1996). The favored interpretation is that the Rio Grande has cycled through several aggradation – incision periods, producing alluvial fills of variable size including the buried axial valley alluvium. The working model for this system are major fills deposited during the major glacial periods such as MIS 2 and 6, minor fills deposited during lesser glacial periods such as MIS 4, and incision occurring during the interglacials or interstadials. Given that the youngest fill remains buried in a bedrock-cut inner valley and is overlain by Holocene alluvium, it is thought to be LGM or MIS 2 age, coincident with Pinedale glaciations in the Rocky Mountains. Subsequently, the minor fill would be estimated to be ~ 60 k.y. old, coincident with MIS 4 and the older, major fill coincident with the pre-LGM or Bull Lake glaciation in the Rocky Mountains. Importantly, the modern interglacial (Holocene) valley remains unincised and rather, filled with alluvium that aggraded during the LGM. This buried axial alluvium will presumably be incised and exposed as a major fill terrace sometime in the future as the Rio Grande responds to the next glacial-interglacial climate cycle.

The precise timing of aggradation and incision in response to glacial-interglacial climates in this system are revealed in the terrace stratigraphy that are best exposed and documented for Rio Grande tributaries (Rogers, 1996; Rogers and Smart, 1996; Formento-Trigilio, 1997; Formento-Trigilio and Pazzaglia, 1998; Pazzaglia, 2005; **Fig. 21**). For example, seven major terraces are preserved in the middle portion of the Jemez valley, in the vicinity of its confluence with the Rio Guadalupe. The highest and oldest of these alluvial deposits (Qg) overlies a paleo-valley bottom of high local relief cut into upper Paleozoic sedimentary rocks. This alluvium is composed of well stratified, well rounded, interbedded, volcanic provenance gravel and sand, with lesser amounts of Paleozoic sedimentary rocks. Deposit thickness varies from 1 to 15 m and is exposed locally beneath the upper (1.2 Ma) and lower (1.6 Ma) Bandelier Tuff. Poorly consolidated ash at the base of the Bandelier ignimbrites contains climbing ripples, indicating its deposition in the ancestral Rio Jemez channel that at this time was an alluvial stream draining a predominantly volcanic upland.

Four major locally paired fill terraces (Qt1 to Qt4) are inset below Qg. Terraces Qt1 through Qt3 have planar straths, but Qt4 has a high-relief strath and clearly represents alluvial burial of paleotopography. All four fill terraces share a common stratigraphy, sedimentology, and provenance (**Fig. 4b**). The deposits are approximately 10-25 m thick and composed of mixed volcanic, plutonic, and sedimentary rock including limestone and chert, and granitic clasts. Texture and bedding in the deposits follow a distinct pattern of a basal, coarse-grained channel facies gravel, conformably overlain by a cross-bedded overbank and crevase-splay facies gravelly sand and silt, unconformably overlain by a second channel facies of coarse-grained gravel. This stratigraphy reflects a change in the overall channel geometry and pattern from braided at the beginning and end of valley alluviation to meandering in the intervening phase. Numeric age control is not precise enough to know how long a valley-filling phase lasted.

Fill terrace Qt4 is interpreted as the now exposed last interglacial valley alluvium. Terrestrial gastropods in the Qt4 deposit are radiocarbon dead but have calibrated amino acid racemization ages of 160±50 k.y. (Rogers and Smart, 1996), terraces stratigraphically older than Qt4 are dated to be older than 180 ka (Rogers and Smart, 1996; Rogers et al., 1996; Goff and Shevenell, 1987), and the next younger inset terrace (Qt5) contains clasts of volcanic rocks known to be erupted 60 ka that are lacking in the Qt4 alluvium. All lines of evidence indicate that the Qt4 alluvium was deposited and then exposed as a fill terrace between 180 and 60 ka, a time period spanning one major glacial-interglacial cycle of MIS 6 and 5. The modern Jemez River below the Rio Guadalupe confluence is predominantly an alluvial river, flowing atop 25 meters of mixed provenance sediment that unconformably overlies Paleozoic, Mesozoic, and Cenozoic bedrock. This late Pleistocene and Holocene alluvium resembles terraces Qt4 in its texture and stratigraphy and like the buried alluvium of the axial Rio Grande valley, likely represents the MIS 2-1, LGM-Holocene glacial interglacial cycle.

In the context of the unsteady discharge and hillslope sediment flux model, deposition and incision of the Qt4 alluvium may have proceeded the following way (**Fig. 22**). Descent into the full glacial conditions of MIS 6 caused a cooling of the climate, increased vegetative cover, a decrease in hillslope erosion rate, and increase in mean annual discharge because of greater mean effective precipitation (less evaporation). The response of the Jemez channel to these changes was to incise through the Qt3 alluvium and into bedrock, carving the bedrock valley floor and walls of the Qt4 alluvium. At full glacial conditions, the new base level of erosion is attained. The Jemez River is a braided stream flowing in coarse alluvium several meters thick. A change to interglacial conditions at the transition from MIS 6 to 5 sees a warming of the climate, a decrease in vegetative cover, and more seasonal, intense precipitation. The hillslopes respond by liberating regolith produced and stored during the cooler, glacial period. Fine-grained sandy soil is striped first, followed by coarser partially-weathered bedrock resulting in an unroofing sequence in the terrace deposit of sand deposited by a meandering channel topped by braided channel gravel. The ensuing interglacial is characterized by a channel with low mean annual discharge because of the warm, arid climate, and hillslopes stripped of vegetation. The result is an underfit Jemez channel flowing atop the Qt4 alluvial fill, unable to incise because of its low discharge and the armoring characteristics of the upper gravel unit. The channel remains atop the Qt4 alluvium more or less through MIS 5, or at least 5e, until descent once again into the next glacial cycle drive incision of the alluvium to expose it as a fill terrace, and prepare the valley to accept the next big pulse of sediment related to the LGM-Holocene glacial-interglacial cycle.

Terrace genesis in valleys that have Holocene strath terraces are thought to similarly respond to the glacial-interglacial cycle, but that response is phase shifted as a function of latitude or altitude. In modern drier, warmer, or mid-latitude settings, glacial climates are commonly associated with more effective precipitation, greater vegetative cover, formation of soils, and slowing of hillslope erosion. In these settings, like the Rio Jemez, it is the transition to an interglacial climate the sparks an increase in hillslope erosion and increased sediment yields in the channel. In contrast, in modern wetter, colder, or higher latitude settings, glacial climates are associated with less effective precipitation, colder, tundra vegetation, intense physical weathering, no soil formation, and an increase in hillslope erosion. Even for drainages lacking glaciers in their headwaters, these conditions cause peak sediment loads to the channel to track in phase with peak glacial conditions, causing valleys to fill with alluvium in the case of over capacity bedload conditions or carve straths in the case of capacity conditions during the glacial phase, and incise creating terraces in the interglacial phase (Formento-Trigilio et al., 2002).

Examples of this phase shifted behavior are well documented for rivers draining active orogens in wet climates (Bull, 1991; Pazzaglia and Brandon, 2001). In Italy, the phase-shift can be observed to happen as a function of latitude, given that the Italian peninsula spans almost 12 degress of latitude. Rivers in southern Italy have a terrace and axial valley stratigraphy virtually identical to that describe for the Rio Grande and its tributaries (Boenzi et al., 2008). However, in northern Italy, the rivers are on bedrock and flanked by Holocene strath terraces (Wegmann and Pazzaglia, 2009). The rates of rock uplift for basins throughout Italy are not uniform, but they vary less than a factor of 2, averaging about 200 m/my (Ferranti et al., 2006; Cyr and Granger, 2008). A better explanation for the latitudinal change in Holocene valley morphology for these

Italian rivers is that erosion rates and sediment yield in rivers in the cooler, better vegetated north respond in phase with the glacial-interglacial cycle, and in shifted phase in the drier, warmer south. The summary conclusion is that for similar tectonic settings, the fluvial incision that creates a strath or fill terrace tends occurs during the interglacial climates for high latitude, cool, well-vegetated settings, but during the transition to full glacial climate for the mid latitude, warm, arid, sparsely vegetated setting.

The role of discharge unsteadiness

The preceding discussion has focused primarily on the role of unsteady sediment yields in response to climate forcing. Here, the role of discharge unsteadiness in response to climate forcing is explored in terms of its role in setting the sediment entrainment and threshold criteria that lead to terrace formation (Sugai, 1993; **Figs. 6**). Rivers draining the volcanic arc in Japan are steep, sediment-rich, contain several tephrachronology-dated and correlated fill and strath terraces, and intercept a proximal western Pacific moisture source. Instrumented watersheds, such as the 287 km^2 Usui River basin, have stage and discharge measurements at up to six locations along the river channel. From these, bed shear stress can be calculated for geomorphically significant flows, which during the late Holocene are dominated by typhoons. Calculated bed shear stresses are higher in the headwaters than in the lower reaches, but they everywhere predict that geomorphically significant flows are able to entrain the grain size distribution of sediment currently being delivered to the channel. This prediction is supported by observation that the channel is on bedrock and in a phase of active vertical incision.

A similar analysis of paleo-bed shear stress was completed using the grain size distributions of the terraces and all discharge and stage height data except the typhoon events. These results predict that the critical tractive forces are not large enough to entrain the grain size distribution and should result in alluvium aggrading in the valley bottom to steepen the transport gradient. This prediction is supported by the presence of fill terraces in the headwaters that transition to strath terraces downstream. Hence, the prevailing discharge characteristics of Pleistocene glacial climates, when no typhoons track towards Japan puts the Usui River into an overcapacity setting resulting in alluviation or wide strath cutting. In contrast, the prevailing discharge characteristics during interglacial climates when typhoons frequent strike Japan puts the Usui River into an undercapacity condition, resulting in vertical incision.

Process linkage and the integrated model

The Holocene terraces of the northeast part of Yellowstone National Park, a largely pristine landscape, synthesize the process linkage between climate, vegetation, hillslope erosion (sediment yield), and discharge (Meyer et al., 1995; **Fig 23**). In this author's opinion, the ensuing model is the best example of the linked processes that result in terrace formation. Strath terraces of Soda Butte Creek record unsteady incision into late Pleistocene glacial deposit "bedrock". Holocene climate for the northern Rockies region is dominated by dry summers where the moisture arrives as intense, convective, isolated storms and moist winters where moisture arrives as snow delivered by regional weather fronts. Droughts occur when winter snowpacks are thin and the lack of snow melt runoff to streams leads to low base flow during the summer months. Forest fires are more numerous during a drought which leads to destabilized hillslopes and delivery of sediment to the valley bottoms by debris flows during summer convective storms. This sediment accumulates in debris fans along the flanks of the river valley. Meanwhile, the axial channel, Soda Butte Creek in this case, narrows and vertically incises in response to the intense and flashy convective storm-driven discharge.

Interspersed with the periods of drought are wet periods that coincide to years when there is ample winter snow pack to provide a stable, base flow and maintain soil moisture through the summer months. High soil moisture suppresses the summer convective storms. Stable, discharge during these wet periods favors channel widening and a meandering pattern, a response that is positively reinforced as the fines from debris fans are recruited by the widening valley bottom.

These climatic characteristics predict that discharge and hillslope sediment flux are out of phase which combine to alternate times of strath carving and valley widening during relative moist, stable discharges and strath incision, terrace formation, and channel lowering during relative dry, flashy discharges. Observations support these predictions. Calibrated radiocarbon ages of the debris fans show that they accumulate during times when there are few or any ages for strath terraces. In contrast, the strath terraces are dated to times of little to no debris fan aggradation. Over the last 3000 years for when there exists the best numeric age constraints, there are three strath terraces interspersed with three incision events. Approximately 70% of the last 3000 years is dominated by the strath forming process, with vertical incision restricted to the shorter time periods coincident with droughts, fires, and debris flow activity.

Summary and future research directions

The large and growing literature on river terraces and the processes that govern their formation have converged on several common themes despite the geographic and temporal diversity of terrace research. Terraces are nearly ubiquitous features of river valleys that attest to a fundamental unsteadiness in the rate of vertical incision. The most common sources of that unsteadiness are vegetative, geomorphic, and hydrologic responses to climate, which for the Quaternary, are characterized by 100-k.y. glacial-interglacial cycles. The precise response depends much on watershed substrate and the climatic, tectonic, and base level setting. Fill terraces indicate a response that drives the sediment transport in the channel to range broadly from under capacity to over capacity conditions. In contrast strath terraces indicate a response driving sediment transport across a more narrow range of at capacity to slightly under capacity conditions.

Continued growth and application of numeric dating methods will further permit terrace studies to quantify the timing of terrace formation and better explore the watershed scale responses to environmental change. Characterizing paleo-erosion rates and precisely when hillslopes liberate sediment during a glacial-interglacial cycle are two examples of how terrace studies will continue to directly influence watershed-scale process investigations. More sophisticated numeric models coupled with field studies dedicated to better understanding the strath formation process remain on the horizon for the next generation of terrace research. As a synthetic and globally applicable genesis model emerges from these studies, fluvial terraces should remain core to a wide range of geologic investigations probing fundamental tectonic, climatic, and paleoclimatic topics.

References

Adam, D.P., and West, G.J., 1983, Temperature and precipitation estimates through the last glacial cycle from Clear Lake, California, pollen data: Science, 219, 168–170, doi: 10.1126/science.219.4581.168.

Anders, M. D., Pederson, J. L., Rittenour, T. M., Sharp, W. D., Gosse, J. C., Karlstrom, K.E., Crossey, L. J., Goble, R. J., Stockli, L., and Yang, G., 2005, Pleistocene geomorphology and geochronology of eastern Grand Canyon: linkages of landscape components during climate changes: Quaternary Science Reviews, 24, 2428-2448.

Baker, V. R. and Ritter, D. F., 1975, Competence of rivers to transport coarse bedload material: Geological Society of America Bulletin, 86, 975-978.

Baker, V. R., 1974, Paleohydraulic interpretation of Quaternary alluvium near Golden, Colorado: Quaternary Research, 4, 94-112.

Belmont, P., Pazzaglia, F. J., and Gosse, J. C., 2007, Cosmogenic ^{10}Be as a tracer for hillslope and channel sediment dynamics in the Clearwater River, western Washington State: Earth and Planetary Science Letters, 264, 123-135.

Bishop, P., Young, R. W., and McDougall, I., 1985, Stream profile change and longterm landscape evolution; early Miocene and modern rivers of the East Australian highland crest, central New South Wales, Australia: Journal of Geology, 93, 455-474.

Blum, M. D., 1993, Genesis and architecture of incised valley fill sequences; a late Quaternary example from the Colorado River, Gulf Coastal Plain of Texas, in Weimer, P., Posamentier, H. W. eds., Siliciclastic sequence stratigraphy; recent developments and applications: AAPG Memoir 58, 259-283.

Blum, M. D., and Valastro, S. Jr., 1994, Late Quaternary sedimentation, lower Colorado River, Gulf Coastal Plain of Texas; with Suppl. Data 9420: Geological Society of America Bulletin, 106, 1002-1016.

Boenzi, F., Caldara, M., Capolongo, D., Dellino, P., Piccarreta, M., Simone, O., 2008, Late Pleistocene-Holocene landscape evolution in Fossa Bradanica, Basilicata (southern Italy): Geomorphology, 102, 297-306.

Bogaart, P. W. and van Balen, R. T., 2001, Numerical modeling of the response of alluvial rivers to Quaternary climate change: Global and Planetary Change, 27, 147-163.

Boll, J., Thewessen, T. J. M., Meijer, E. L., and Kroonenberg, S. B. 1988, A simulation of the development of river terraces: Zeitschrift fuer Geomorphologie, 32, 31-45.

Bridgland, D. R., 2000, River terrace systems in North-west Europe; an archive of environmental change, uplift and early human occupation: Quaternary Science Reviews, 19, 1293-1303.

Bucher, W. H., 1932, “Strath” as a geomorphic term: Science, 75, 130-131.

Bull, W. B. and Knuepfer, P. L. K., 1987, Adjustments by the Charwell River, New Zealand, to uplift and climatic changes: Geomorphology, 1, 15-32.

Bull, W. B. and Schick, A. P., 1979, Impact of climatic change on an arid watershed; Nahal Yael, southern Israel: Quaternary Research, 11, 153-171.

Bull, W. B., 1979, Threshold of critical power in streams, Geological Society of America Bulletin, 90, 453-464.

Bull, W. B., 1991, Geomorphic responses to climate change: Oxford University Press, Oxford, 326 p.

Bull, W. B., 2007, Tectonic geomorphology of mountains: Blackwell Publishing, Malden, MA, 316 p.

Campbell, M. R., 1929, The river system, a study in the use of technical geographic terms: Journal of Geography, 28, 123-128.

Cande, S.C., Kent, D.V., 1995, Revised calibration of the geomagnetic polarity timescale for the Late Cretaceous and Cenozoic: Journal of Geophysical Research Solid Earth, 100, p. 6093-6095.

Carney, F., 1907, A form of outwash drift: American Journal of Science, 23, 336-341.

Carroll, A. R., Doebbert, A. C., Booth, A. L., Chamberlain, C. P., Rhodes-Carson, M. K., Smith, M. E., Johnson, C. M., and Beard, B. L., 2008, Capture of high-altitude precipitation by a low-altitude Eocene lake, Western U.S.: Geology, 36, 791-794.

Chatanantavet, P. and Parker, G., 2009, Physically based modeling of bedrock incision by abrasion, plucking, and macroabrasion: Journal of Geophysical Research Earth Surface, 114, doi:10.1029/2008JF001044.

Coltorti, M., Consoli, M., Dramis, F., Gentili, B. and Pambianchi, G., 1991. Evoluzione geomorfologica delle piane alluvionali delle Marche centro-meridionali. Geografia Fisica e Dinamica Quaternaria, 14, 87-100.

Connell, S. C., 2008, Geologic map of the Albuquerque-Rio Rancho metropolitan area and vicinity, Bernalillo and Sandoval counties, New Mexico: New Mexico Bureau of

Geology and Mineral Resources, Geologic Map 78, 1:200,000.

Connell, S. D., Love, D. W., and Dunbar, N. W., 2007, Geomorphology and stratigraphy of inset fluvial deposits along the Rio Grande valley in the central Albuquerque Basin, New Mexico: New Mexico Geology, 29, 13-31.

Cragun, W. S., 2007, Quaternary evolution of the Colorado River at Lees Ferry, Arizona [M.S. Thesis]: Utah State University, Logan, UT, United States, 179 p.

Crosby, B. T., Whipple, K. X., Gasparini, N. M., and Wobus, C. W., 2007, Formation of fluvial hanging valleys; theory and simulation: Journal of Geophysical Research, 112, 10.1029/2006JF000566.

Cyr, A. J. and Granger, D. E., 2008, Dynamic equilibrium among erosion, river incision, and coastal uplift in the Northern and Central Apennines, Italy: Geology, 36, 103-106.

Dana, J. D., 1881, On the relation of the so-called "kames" of the Connecticut River valley to the terrace formation: American Journal of Science, 22, 451-468.

de Wit, M. C. J., Marshall, T. R., and Partridge, T. C., 2000, Fluvial deposits and drainage evolution in Partridge, T. C. and Maud, R. R., eds., The Cenozoic of Southern Africa: Oxford Monographs on Geology and Geophysics, 40, 55-72.

Deacon, J. and Lancaster, N., 1988, Late Quaternary palaeoenvironments of Southern Africa, Oxford Science Publishing, New York, 225 p.

Dethier, D. P., 2001, Pleistocene incision rates in the Western United States calibrated using Lava Creek B Tephra: Geology, 29, 783-786.

Duvall, A., Kirby, E., and Burbank, D., 2004, Tectonic and lithologic controls on bedrock channel profiles and processes in coastal California, Journal of Geophysical Research – Earth Surface, 109, 10.1029/2003JF000086.

Ferranti, L., Antonioli, F., Mauz, B., Amorosi, A., Dai Pra, G., Mastronuzzi, G., Monaco, C., Orru, P., Pappalardo, M., Radtke, U., Renda, P., Romano, P., Sanso, P., and Verrubbi, V., 2006, Markers of the last interglacial sea-level high stand along the coast of Italy; tectonic implications: Quaternary International, 145-146, 30-54.

Formento-Trigilio, M. L., Burbank, D. W., Nicol, A., Shulmeister, J., and Rieser, U., 2002, River response to an active fold-and-thrust belt in a convergent margin setting, North Island, New Zealand: Geomorphology, 49, 125-152.

Formento-Trigilio, M. L. and Pazzaglia, F.J., 1998, Tectonic geomorphology of Sierra

Nacimiento; traditional and new techniques in assessing long-term landscape evolution of the southern Rocky Mountains: Journal of Geology, v. 106, p. 433-453.

Formento-Trigilio, M.L., 1997, Tectonic Geomorphology and long-term landscape evolution of Sierra Nacimiento, northern New Mexico [M.S. thesis]: Albuquerque, University of New Mexico, 201 p.

Frankel, K. L. and Pazzaglia, F. J., 2006, Mountain fronts, base level fall, and landscape evolution: Insights from the southern Rocky Mountains, in Willett, S. D., Hovius, N., Brandon, M. T., and Fisher, D. eds., Tectonics, climate, and landscape evolution: Geological Society of America Special Paper 398, 419-434.

Frankel, K. L., Pazzaglia, F. J., and Vaughn, J. D., 2007, Knickpoint evolution in a vertically bedded substrate, upstream-dipping terraces, and Atlantic slope bedrock channels: Geological Society of America Bulletin, 119, 476-486.

Fuller, T. K., Perg, L. A., Willenbring, J. K., and Lepper, K., 2009, Field evidence for climate-driven changes in sediment supply leading to strath terrace formation: Geology, 37, 467-470.

Garcia, A. F., 2006, Thresholds of strath genesis deduced from landscape response to stream piracy by Pancho Rico Creek in the Coast Ranges of Central California: American Journal of Science, 306, 655-681.

Garcia, A. F., Zhu, Z., Ku, T. L., Chadwick, O. A., and Chacon M., J., 2004, An incision wave in the geologic record, Alpujarran Corridor, southern Spain (Almeria): Geomorphology, 60, 37-72.

Gardner, T. W., 1983, Experimental study of knickpoint and longitudinal profile evolution in cohesive, homogeneous material: Geological Society of America Bulletin, 94, 664-672.

Gardner, T. W., Jorgensen, D. W., Shuman, C., and Lemieux, C. R., 1987, Geomorphic and tectonic process rates; effects of measured time interval: Geology, 15, 259-261.

Gardner, T. W., Verdonck, D., Pinter, N. M., Slingerland, R. L., Furlong, K. P., Bullard, T. F., and Wells, S. G., 1992, Quaternary uplift astride the aseismic Cocos Ridge, Pacific Coast, Costa Rica: Geological Society of America Bulletin, 104, 219-232.

Gibbard, P. L. and Lewin, J., 2009, River incision and terrace formation in the late Cenozoic of Europe: Tectonophysics, 474, 41-55.

Gilbert, G. K., 1877, Report on the geology of the Henry Mountains [Utah]: Publication of the

Powell Survey, U.S. Government Printing Office, Washington D.C., 160 p.

Goff, F. and Shevenell, L., 1987, Travertine deposits of Soda Dam, New Mexico and their implications for the age and evolution of the Valles caldera hydrothermal system: Geological Society of America Bulletin, v. 99, p. 292-302.

Gosse, J. C. and Phillips, F. M., 2001, Terrestrial in situ cosmogenic nuclides; theory and application: Quaternary Science Reviews, 20, 1475-1560.

Gunnell, Y., Gallagher, K., Carter, A., Widdowson, M., and Hurford, A. J., 2003, Denudation history of the continental margin of western peninsular India since the early Mesozoic; reconciling apatite fission-track data with geomorphology: Earth and Planetary Science Letters, 215, 187-201.

Hack, J. T., 1960, Studies of longitudinal stream profiles in Virginia and Maryland: U. S. Geological Survey Professional Paper 294B, 45-97.

Hancock, G. S. and Anderson, R. S., 2002, Numerical modeling of fluvial strath-terrace formation in response to oscillating climate: Geological Society of America Bulletin, 114, 1131-1142.

Haug, G., and Tiedemann, 1998, Effect of the formation of the Isthmus of Panama on Atlantic Ocean thermohaline circulation: Nature, 395, 673-676.

Harkins, N., Kirby, E., Heimsath, A., Robinson, R., and Reiser, U., 2007, Transient fluvial incision in the headwaters of the Yellow River, northeastern Tibet, China: Journal of Geophysical Research, 112, 10.1029/2006JF000570.

Hartshorn, K., Hovius, N., Dade, W. B., and Slingerland, R. L., 2002, Climate-driven bedrock incision in an active mountain belt: Science, 297, 2036-2038.

Hawley, J.W., and Kennedy, J.F., 2004, Creation of a digital hydrogeologic framework model of the Mesilla Basin and southern Jornada del Muerto Basin: N. M. Water Resources Research Institute, NMSU; prepared for Lower Rio Grande Water Users Organization; Technical Completion Report 332, 105 p. With CD ROM including 2005 Addendum extending model into Rincon Valley and adjacent areas. *(p. 48, 51-53, 73; plates on CD-ROM include longitudinal section of inner-valley fill from Caballo Dam to El Paso)* http://wrri.nmsu.edu/publish/

Hitchcock, E., 1824, Remarks on the geology of the district adjoining the Erie Canal, *in* Amos, E., A geological and agricultural survey of the district adjoining the Erie Canal in the

State of New York; taken under the direction of the Hon. Stephen Van Rensselaer; Part I, Containing a description of the rock formations together with a geological profile extending from the Atlantic to Lake Erie, Packard and van Benthuysen, Albany, NY, 158-163.

Howard, J. L., Amos, D. F., and Daniels, W. L., 1993, Alluvial soil chronosequence in the inner coastal plain, central Virginia: Quaternary Research, 39, 201-213.

Kelson, K. I. and Wells, S. G., 1989, Geologic influences on fluvial hydrology and bedload transport in small mountainous watersheds, northern New Mexico, USA: Earth Surface Processes and Landforms, 14, 671-690.

Kelson, K. I., 1986, Long-term tributary adjustments to base-level lowering, northern Rio Grande Rift, New Mexico [M.S. Thesis]: Albuquerque, University of New Mexico, 210 p.

Kirby, E. and Whipple, K., 2001, Quantifying differential rock-uplift rates via stream profile analysis: Geology, 29, 415-418.

Knox, J. C., 1975, Concept of the graded stream in, Melhorn, W. N. and Flemal, R. C., eds., Theories of landform development: Procedings of the 6th annual Binghamton Symposium, Publications in Geomorphology, State University of New York, Binghamton, NY, 169-198.

Lambert, P. W., 1968, Quaternary stratigraphy of the Albuquerque area, New Mexico [Ph.D. Thesis]: Albuquerque, University of New Mexico, 329 p.

Lave, J. and Avouac, J. P., 2000, Active folding of fluvial terraces across the Siwaliks Hills, Himalayas of central Nepal: Journal of Geophysical Research: 105, 5735-5770

Leopold, L. B. and Bull, W. B., 1979, Base level, aggradation, and grade: Proceedings of the American Philosophical Society, 123, 168-202.

Leopold, L.B., Wolman, M.G., and Miller, J. P., 1964, Fluvial processes in geomorphology: San Francisco, Calif., W. H. Freeman and Company, 522 p.

Lesley, J. P., 1878, On terrace levels in Pennsylvania: American Journal of Science, 16, 68-69.

Mackin, J. H., 1937, Erosional history of the Big Horn Basin, Wyoming: Geological Society of America Bulletin, 48, 813-893.

Mackin, J. H., 1948, Concept of the graded river: Geological Society of America Bulletin, 59, 463-511.

Maddy, D., 1997, Uplift-driven valley incision and river terrace formation in southern England: Journal of Quaternary Science, 12, 539-545.

Maddy, D., Bridgland, D., and Westaway, R., 2001, Uplift-driven valley incision and climate-controlled river terrace development in the Thames Valley, UK: Quaternary International, 79, 23-36.

Matmon, A., Bierman, P., and Enzel, Y., 2002, Pattern and tempo of great escarpment erosion: Geology, 30, 1135-1138.

McMillan, M. E., Heller, P. L., and Wing, S. L., 2006, History and causes of post-Laramide relief in the Rocky Mountain orogenic plateau: Geological Society of America Bulletin, 118, 393-405.

Meyer, G. A., Wells, S. G. ; Balling, R. C., Jr., Jull, A. J. T., 1992, Response of alluvial systems to fire and climate change in Yellowstone National Park: Nature, 357, 147-150.

Meyer, G. A., Wells, S. G., and Jull, A. J. T., 1995, Fire and alluvial chronology in Yellowstone National Park; climatic and intrinsic controls on Holocene geomorphic processes: Geological Society of America Bulletin, 107, 1211-1230.

Meyer, W. and Stets, J., 2002, Pleistocene to Recent tectonics in the Rhenish Massif (Germany): Rift tectonics and syngenetic sedimentation; the Cenozoic Lower Rhine Basin and related structures: Netherlands Journal of Geosciences, 81, 217-221.

Miller, S. R., Slingerland, R. L., and Kirby, E., 2007, Characteristics of steady state fluvial topography above fault-bend folds: Journal of Geophysical Research, 112, DOI: 10.1029/2007JF000772.

Mills, H. H., 2000, Apparent increasing rates of stream incision in the Eastern United States during the late Cenozoic: Geology, 28, 955-957.

Molin, P., Pazzaglia, F. J., and Dramis, Francesco, 2004, Geomorphic expression of active tectonics in a rapidly-deforming forearc, Sila Massif, Calabria, southern Italy: American Journal of Science, 304, 559-589.

Montgomery, D. R., 2004, Observations on the role of lithology in strath terrace formation and bedrock channel width: American Journal of Science, 304, 454-476.

Moucha, R., Forte, A. M., Mitrovica, J. X., Rowley, D. B., Quere, S., Simmons, N. A., and Grand, S. P., 2008, Dynamic topography and long-term sea-level variations; there is no such thing as a stable continental platform: Earth and Planetary Science Letters, 271, 101-108.

Nelson, N. P., 1983, The formation of a terrace: American Geologist, 12, 125-126.

Nesci, O. and Savelli, D., 1990, Valley terraces in the Northern Marche Apennines (Central Italy): cyclic deposition and erosion: Giornale di Geologia, 52, 89-195.

Nesci, O. and Savelli, D., 2003, Diverging drainage in the Marche Apennines (central Italy): Quaternary International, 101-102, 203-209.

Ohmori, H., 1991, Change in the mathematical function type describing the longitudinal profile of a river through an evolutionary process: Journal of Geology, 99, 97-110.

Pan B., Burbank, D., Wang Y., Wu G., Li J., and Guan Q., 2003, A 900 k.y. record of strath terrace formation during glacial-interglacial transitions in Northwest China: Geology, 31, 957-960.

Paola, C. and Mohrig, D., 1996, Paleohydraulics revisited: paleoslope estimation in coarse-grained braided rivers: Basin Research, 8, 243-254.

Partridge, T. C. and Maud, R. R., 1987, Geomorphic evolution of Southern Africa since the Mesozoic: Transactions of the Geological Society of South Africa, 90, 179-208.

Patton, P. C. and Schumm, S. A., 1975, Gully Erosion, Northwestern Colorado; A Threshold Phenomenon: Geology, 3, 88-89.

Pazzaglia, F. J. and Brandon, M. T., 1996, Macrogeomorphic evolution of the post-Triassic Appalachian Mountains determined by deconvolution of the offshore based sedimentary record: Basin Research, 8, 255-278.

Pazzaglia, F. J. and Brandon, M. T., 2001, A fluvial record of long-term steady-state uplift and erosion across the Cascadia forearc high, western Washington State: American Journal of Science, 301, 385-431.

Pazzaglia, F. J. and Gardner, T. W., 1993, Fluvial terraces of the lower Susquehanna River: Geomorphology, 8, 83-113.

Pazzaglia, F. J., Thackray, G. D., Brandon, M. T., Wegmann, K. W., Gosse, J., McDonald, E., Garcia, A. F., and Prothero, D., 2003, Tectonic geomorphology and the record of Quaternary plate boundary deformation in the Olympic Mountains, *in* Swanson, T. W., ed., Western Cordillera and adjacent areas: Boulder, Colorado, Geological Society of America Field Guide 4, p. 37-67.

Pazzaglia, F. J., Braun, D. D., Pavich, M., Potter, N., Jr, Merritts, D., Walter, R., and Germanoski, D., 2006, Rivers, glaciers, landscape evolution, and active tectonics of the Central Appalachians, Pennsylvania and Maryland: Geological Society of America Field Guide 8, 169-197.

Pazzaglia, F. J., Gardner, T. W., and Merrits, D. J., 1998, Bedrock fluvial incision and longitudinal profile development over geologic time scales determined by fluvial terraces *in*, Tinkler, K. J.; Wohl, E. E., eds., Rivers over rock; fluvial processes in bedrock channels: Geophysical Monograph, 107, 207-235.

Peate, D. W., Chen, J. H., Wasserburg, G. J., and Papanastassiou, D. A., 1996, ^{238}U- ^{230}Th dating of a geomagnetic excursion in Quaternary basalts of the Albuquerque volcanoes field, New Mexico (USA): Geophysical Research Letters, 23, 2271-2274.

Pederson, J. L., Anders, M.D., Rittenhour, T. M., Sharp, W. D., Gosse, J. C., and Karlstrom, K. E., 2006, Using fill terraces to understand incision rates and evolution of the Colorado River in eastern Grand Canyon, Arizona: Journal of Geophysical Research, 111, doi:10.1029/2004JF000201.

Pelletier, J. D., 2008, Quantitative modeling of Earth surface processes: Cambridge University Press, Cambridge, 295 p.

Penck, A. and Brückner, E., 1909, Die Alpen im Eiszeitalter: Tauchnitz, Leipzig, 1199 p.

Pierce, K. L. and Scott, W. E., 1982, Pleistocene alluvial-gravel in eastern Idaho: U. S. Geological Survey Professional Paper 1375, 63-64.

Reusser, L. J., Bierman, P. R., Pavich, M. J., Zen, E-an, Larsen, J., Finkel, R. C., 2004, Rapid late Pleistocene incision of Atlantic passive-margin river gorges: Science, 305, 499-502.

Reusser, L. Bierman, P., Pavich, M., Larsen, J., Finkel, R., 2006, An episode of rapid bedrock channel incision during the last glacial cycle, measured with (super 10) Be: American Journal of Science, 306, 69-102.

Riihimaki, C. A., Anderson, R. S., and Safran, E. B., 2007, Impact of rock uplift on rates of late Cenozoic Rocky Mountain river incision: Beyond steady state; the dynamics of transient landscapes: Journal of Geophysical Research, 112, 10.1029/2006JF000557.

Ritter, D. F. and Miles, C., 1973, Problems of stream terrace correlation and reconstruction of geomorphic history caused by colluviums: Geografiska Annaler. Series A: Physical Geography, 2, 85-91.

Ritter, D. F., 1967, Terrace development along the front of the Beartooth Mountains, southern

Montana: Geological Society of America Bulletin, 78, 467-483.

Ritter, D. F., Kochel, R. C., and Miller, J. R., 2002, Process Geomorphology: Waveland Press, Long Grove, Illinois, 560 p.

Ritter, J. B., Miller, J. R., and Husek-Wulforst, J., 2000, Environmental controls on the evolution of alluvial fans in Buena Vista Valley, north central Nevada, during late Quaternary time: Geomorphology, 36, 63-87.

Ritter, J. B., Miller, J. R., Enzel, Y., Howes, S. D., Nadon, G., Grubb, M. D., Hoover, K. A., Olsen, T., Reneau, S. L., Sack, D., Summa, C. L., Taylor, I., Touysinhthiphonexay, K. C. N., Yodis, E. G., Schneider, N. P., Ritter, D. F., and Wells, S. G., 1993, Quaternary evolution of Cedar Creek alluvial fan, Montana: Geomorphology, 8, 287-304.

Roe, G. H., Montgomery, D. R., and Hallet, B., 2002, Effects of orographic precipitation variations on the concavity of steady-state river profiles: Geology 30, 143-146.

Rogers, J. B. and Smartt, R. A., 1996, Climatic influences on Quaternary alluvial stratigraphy and terrace formation in the Jemez river valley, New Mexico: New Mexico Geological Society, Guidebook 47, p. 347-356.

Rogers, J. B., 1996, The fluvial landscape evolution of San Diego Canyon, Jemez Mountains, New Mexico [M.S. Thesis]: Albuquerque, University of New Mexico, 123 p.

Rogers, M. A., Kues, B. A., Goff, F., Pazzaglia, F. J., Woodward, L. A., Lucas, S. G., and Gardner, J. A., 1996, First-day road log, from Bernalillo to San Ysidro, southern Nacimiento Mountains, Guadalupe Box, Jemez Springs, Valles Caldera, and Los Alamos: New Mexico Geological Society, Guidebook 47, p. 1-40.

Roy, M., Kelley, S., Pazzaglia, F. J., Cather, S., and House, M. A., 2004, Middle Tertiary buoyancy modification and its relationship to rock exhumation, cooling, and subsequent extension at the eastern margin of the Colorado Plateau: Geology, 32, 925-928.

Safran, E. B., Bierman, P. R., Aalto, R., Dunne, T., Whipple, K. X., and Caffee, M. W., 2005, Erosion rates driven by channel network incision in the Bolivian Andes: Earth Surface Processes and Landforms, 30, 1007-1024.

Sak, P. B., Fisher, D. M., and Gardner, T. W., 2004, Effects of subducting seafloor roughness on upper plate vertical tectonism; Osa Peninsula, Costa Rica: Tectonics, 23, 10.1029/2002TC001474.

Schaller, M., von Blanckenburg, F., Veldkamp, A., Tebbens, L. A., Hovius, N., and Kubik, P. W., 2002, A 30,000 yr record of erosion rates from cosmogenic (super 10) Be in middle European river terraces: Earth and Planetary Science Letters, 204, 307-320.

Schaller, M., von Blanckenburg, F., Hovius, N., Veldkamp, A., van den Berg, Meindert W., and Kubik, P. W., 2004, Paleoerosion rates from cosmogenic ^{10}Be in a 1.3 Ma terrace sequence; response of the River Meuse to changes in climate and rock uplift: Journal of Geology, 112, 127-144.

Schoenbohm, L. M., Whipple, K. X., Burchfiel, B.C., and Chen, L., 2004, Geomorphic constraints on surface uplift, exhumation, and plateau growth in the Red River region, Yunnan Province, China: Geological Society of America Bulletin, 116, 895-909.

Schumm, S. A., 1969, River metamorphosis: Proceedings of the American Society of Civil Engineers, 95, 255-273.

Schumm, S. A., 1977, The fluvial system: John Wiley and Sons, New York, 338.

Schumm, S. A. and Rea, D. K., 1995, Sediment yield from disturbed earth systems: Geology, 23, 391-394.

Schumm, S. A., Mosley, M. P., and Weaver, W. E., 1987, Experimental fluvial geomorphology: John Wiley and Sons, New York, 413 p.

Sinha, S. K. and Parker, G., 1996, Causes of concavity in longitudinal profiles of rivers: Water Resources Research, 32, 1417-1428.

Small, E. E., Anderson, R. S., Repka, J. L., and Finkel, R., 1997, Erosion rates of alpine bedrock summit surfaces deduced from in situ ^{10}Be and ^{26}Al: Earth and Planetary Science Letters, 150, 413-425.

Snyder, N. P. , Whipple, K. X., Tucker, G. E., and Merritts, D. J., 2000, Landscape response to tectonic forcing; digital elevation model analysis of stream profiles in the Mendocino triple junction region, Northern California: Geological Society of America Bulletin, 112, 1250-1263.

Spagnolo, M. and Pazzaglia, F. J., 2005, Testing the geological influences on the evolution of river profiles; a case from the Northern Apennines (Italy): Geografia Fisica e Dinamica Quaternaria (Testo Stampato), 28, 103-113.

Sugai, T., 1993, River terrace development by concurrent fluvial processes and climatic changes: Geomorphology, 6, 243-252.

Tebbens, L. A., Veldkamp, A., van Dijke, J. J., and Schoorl, J. M., 2001, Modeling longitudinal-profile development in response to late Quaternary tectonics, climate and sea-level changes; the River Meuse: Global and Planetary Change, 27, 165-186.

Tucker, G. E. and Slingerland, R., 1997, Drainage basin responses to climate change: Water Resources Research, 33, 2031-2047.

van Balen, R. T., Houtgast, R. F., van der Wateren, F. M., Vandenberghe, J., and Bogaart, P. W., 2000, Sediment budget and tectonic evolution of the Meuse catchment in the Ardennes and the Roer Valley rift System: Global and Planetary Change, 27, 113-129.

van den Berg, M. W. and van Hoof, T., 2001, The Maas Terrace sequence at Maastricht, SE Netherlands; evidence for 200 m of late Neogene and Quaternary surface uplift *in*, Maddy, D., Macklin, M. G., Woodward, J. C., eds., River basin sediment systems; archives of environmental change: A.A. Balkema Publishers, Netherlands, 45-86.

van den Berg, M. W.,1996, Fluvial sequences of the Maas: Ph.D. dissertation, Landbouwuniversiteit, 181 p.

Vandenberghe, J., 2003, Climate forcing of fluvial system development; an evolution of ideas: Quaternary Science Reviews, 22, 2053-2060.

Vandenberghe, J. and Maddy, D., 2000, The significance of fluvial archives in geomorphology: Geomorphology, 33, 127-130.

Veldkamp, A. and van Dijke, J. J., 2000, Simulating internal and external controls on fluvial terrace stratigraphy; a qualitative comparison with the Maas record: Quaternary fluvial archives: Geomorphology, 33, 225-236.

Veldkamp, A. and Vermeulen, S. E. J. W., 1989, River terrace formation, modelling, and 3-D graphical simulation: Earth Surface Processes and Landforms, 14, 641-654.

Veldkamp, A., 1992, A 3-D model of Quaternary terrace development, simulations of terrace stratigraphy and valley asymmetry; a case study for the Allier terraces (Limagne, France): Earth Surface Processes and Landforms, 17, 487-500.

Vita-Finzi, C., 1969, The Mediterranean Valleys; geological changes in historical times: Cambridge, Cambridge University Press, 140 p.

Vita-Finzi, C., 1976, Diachronism in Old World alluvial sequences: Nature, 263, 218–219.

Wegmann, K. and Pazzaglia, F. J., 2009, Late Quaternary fluvial terraces of the Romagna and Marche Apennines, Italy: Quaternary Science Reviews, 28, 137-165.

Wegmann, K. W. and Pazzaglia, F. J., 2002, Holocene strath terraces, climate change, and active tectonics; the Clearwater River basin, Olympic Peninsula, Washington State: Geological Society of America Bulletin, 114, 731-744.

Wells, S. G., McFadden, L. D., and Dohrenwend, J. C., 1987, Influence of late Quaternary climatic changes on geomorphic and pedogenic processes on a desert piedmont, eastern Mojave Desert, California: Quaternary Research, 27, 130-146.

Whipple, K. X., 2001, Fluvial landscape response time; how plausible is steady-state denudation? : American Journal of Science, 301, 313-325.

Whipple, K. X., 2001, Fluvial landscape response time; how plausible is steady-state denudation?: American Journal of Science, 301, 313-325.

Whipple, K. X., Hancock, G. S., and Anderson, R. S., 2000, River incision into bedrock; mechanics and relative efficacy of plucking, abrasion, and cavitation: Geological Society of America Bulletin, 112, 490-503.

Winograd, I. J., Coplen, T. B., Landwehr, J. M., Riggs, A. C., Ludwig, K. R., Szabo, B. J., Kolesar, P. T., Revesz, K. M., 1992, Continuous 500,000-year climate records from vein calcite in Devils Hole, Nevada: Science, 258, 255–260.

Winograd, I. J., Landwehr, J. M., Coplen, T. B., Sharp, W. D., Riggs, A. C., Ludwig, K. R., and Kolesar, P. T., 2006, Devils Hole, Nevada, $\delta^{18}O$ record extended to the mid-Holocene: Quaternary Research, 66, 202-212.

Wisniewski, P. A. and Pazzaglia, F. J., 2002, Epeirogenic controls on Canadian River incision and landscape evolution, Great Plains of northeastern New Mexico: Journal of Geology, 110, 437-456.

Wobus, C., Tucker, G., and Anderson, R. S., in press, Cenozoic incision of the North American High Plains: Tectonic rejuvenation or climate change?: Geology.

Wohl, E. E. and Ikeda, H., 1997, Experimental simulation of channel incision into a cohesive substrate at varying gradients: Geology, 25, 295-298.

Womack, W. R. and Schumm, S. A., 1977, Terraces of Douglas Creek, northwestern Colorado; an example of episodic erosion: Geology, 5, 72-76.

Young, R. and McDougall, I., 1993, Long-term landscape evolution; early Miocene and modern rivers in southern New South Wales, Australia: Journal of Geology, 101, 35-49.

Zaprowski, B. J., Evenson, E. B., Pazzaglia, F. J., and Epstein, J. B., 2001, Knickzone propagation in the Black Hills and northern High Plains; a different perspective on the late Cenozoic exhumation of the Laramide Rocky Mountains: Geology, 29, 547-550.

Zaprowski, B. J., Pazzaglia, F. J., and Evenson, E. B., 2005, Climatic influences on profile concavity and river incision: Journal of Geophysical Research, 110, DOI: 10.1029/2004JF000138.

Zhang P., Molnar, P., and Downs, W. R., 2001, Increased sedimentation rates and grain sizes 2-4 Myr ago due to the influence of climate change on erosion rates: Nature, 410, 891-897.

Figure Captions

Figure 1. Photographs of terraces. (a) Gray gravel caps on Red Permian bedrock, Jemez river valley, New Mexico; (b) Glacial outwash terraces, Snake River overlook, Grand Teton National Park, WY; (c) two yellow-colored late Pleistocene terraces cut into gray Miocene bedrock, Marzabotto, Italy; (d) Holocene strath terrace cut on top of middle Pleistocene beach sands, Stirone Regional Park, Italy; (e) stacked alluvial gravels separated by a red paleosol in a composite fill terrace, Taro River valley, Italy; (f) middle Holocene strath terrace, Clearwater River , Olympic Peninsula, Washington State; (g) strath terrace of Primus Creek, Alaska; (h) strath terrace illustrating basal unconformity (strath), alluvium, and intact soil of the terrace tread, Reno River, Fortuna, Italy.

Figure 2. Sketches showing (a) a strath beneath a pediment, (b) cross section of a valley with fill terraces, and (c) cross section of a valley with strath terraces (Gilbert, 1877).

Figure 3. Sketch illustrating paired and unpaired fill, fill-cut, and strath terraces. Post-depositional modification of the treads is shown as alluvial fans. Nomenclature follows standard lithostratigraphic convention of beginning numbering with the oldest deposits. Numbers correspond to straths or former valley bottoms. Letters denote multiple treads being shared by one strath. Qg refers to non-stratified upland gravels mantling the drainage divide, but out of the context of the river valley.

Figure 4. (a) Photo and line drawing of a strath facies model for a middle Holocene strath terrace, Clearwater River, WA (Pazzaglia et al., 2003). Letters are soil profile horizon designations. (b) Fill terrace facies model photo and line drawing of Jemez River fill terraces (Rogers and Smith, 1996). The tephra is widespread throughout the American Rocky Mountain west and indicates a period of widespread aggradation ~ 600 ka.

Figure 5. Marine oxygen isotope curve from the equatorial Pacific ODP site 846 (Haug and Tiedemann, 1998; Cande and Kent, 1995). Numbers refer to stages, odd numbers are warm interglacials, even numbers are cold glacials. The upper shaded region corresponds to eccentricity-dominated 100-ky glacial-interglacial cycles.

Figure 6. Longitudinal profile of the Usui River, Japan with long profiles of five major fill and strath terraces (modified from Sugai, 1993). Younger fill terraces are inset

progressively further upstream from the previous fill. Fill terrace treads are steeper and less concave than their corresponding straths.

Figure 7. Line drawing superimposed on latest Pleistocene (11 ka) terrace of the Reno River at Marzabotto, Italy. The alluvium is composed of well-sorted and stratified axial channel gravels for the Reno River, colluvium derived from the adjacent hillslope, and poorly-sorted, local provenance sand a gravel delivered by a tributary.

Figure 8. (a) Longitudinal profile of a tributary of Jordan Creek, draining mixed sedimentary rocks in the Lehigh Valley, Pennsylvania. Raw profile extracted from a 10-m DEM shown as a broad gray line. The smooth profile using a lowess filter is superimposed on top as a thin, solid black line. Growth of drainage area as a function of distance from the divide is shown as a blue line. The slope length index (Hack, 1957) is shown as the red line. Plotted in slope-area space, the long profile can be resolved into two components, the upper of which is graded with normal concavity (olive green), the lower of which is convex (green). b. Photo of the Great Falls knickpoint on the Potomac River, MD-VA.

Figure 9. Common geometries of terrace long profiles as a result of (a) steady-state long profile and uniform incision, (b) increase in profile concavity, and (c) upstream propagation of a knickpoint.

Figure 10. Plots showing the dependence of profile concavity, measured as θ and SCI (see Zaprowski et al., 2005) on the climate proxies of (a) precipitation intensity and (b) peak annual discharge (modified from Zaprowski et al., 2005).

Figure 11. Response of the lower part of a river profile to base level fall across a shelf or landscape of different gradients. (a) Shelf gradient is lower than the river profile gradient, resulting in a large horizontal translation of the coast and fluvial aggradation. (b) Shelf has the same gradient as the river with no corresponding incision or aggradation. (c) Shelf is stepper than the river profile resulting in incision and upstream propagation of a knickpoint.

Figure 12. Photographs of modern straths cut across sedimentary bedrock in the (a) Reno River near Marzabotto, Italy, and (b) Clearwater River, Olympic Mountains, WA. Note the strath at the base of the terrace in (a).

Figure 13. Strath terrace results from flume experiments (modified from Frankel et al., 2007). (a)

channel incises through simulated bedrock canyon in response to upstream knickpoint migration subsequent to the base-level fall. (b) a knickpoint, plunge pool, and strath terrace. (c) A pulse of sediment covers the bottom of the bedrock gorge. The sediment is deposited in the canyon when incision occurs upstream as a result of oversteepening of the broad channel convexity associated with the step-pools in (b). (d) Fluvial features preserved in the bedrock reach at the conclusion of the experiment. In all photographs, arrows point to strath terraces and upstream-dipping terraces formed in bedrock. Dashed white line highlights an upstream-dipping terrace that converges with the active channel bottom. form as the stream continues to incise through the simulated bedrock reach in response to the base-level fall. (e) Sketch of strath terraces formed in simulated alluvium and bedrock following a base level fall and passage of a knickpoint (Gardner, 1983).

Figure 14. River genesis of terrace as a result of a drainage capture that instantaneously changed discharge and sediment caliber for the trunk channel (from Garcia, 2006). (a) Long profile of Pancho Rico Creek showing the point of capture and change in profile shape in the past ~20 ky. (b) Comparison of the pre- and post-capture Pancho Rico valley showing production of terraces. The ratios refer to the pre-/post-capture upstream drainage area from the point of the cross section.

Figure 15. Results of a numeric modeling study of strath terrace genesis (modified from Hancock and Anderson, 2002). (a) lateral and vertical incision rates under the forcing of changing sediment supply and discharge; (b) terraces produced under the changing sediment supply experiment.

Figure 16. Results from the Wegmann and Pazzaglia (2002) study of Holocene strath terrace genesis in the Clearwater River, Olympic Mountains, WA. (a) Map of terraces in the middle portion of the Clearwater valley. (b) Cross section showing terraces along profile X-X’ on (a). (c) Incision rate of Pleistocene (triangles) and Holocene terraces (circles) showing that the Holocene incision rates are 2-3 times faster. (d) Stacked probability diagram of 38 calibrated radiocarbon dates and their associated 2-sigma errors. These data are interpreted as supporting for relatively long periods of time with little to no vertical incision when the straths are cut interspersed with brief periods of time of rapid vertical incision when the straths become strath terraces.

Figure 17. Strath terraces in the (a) Holtwood reach of the lower Susquehanna River, Pennsylvania (modified from Reusser et al., 2006; Pazzaglia et al.,2006). (b) The incision rates obtained from these terraces indicates that there must be brief periods of time when the river vertically incises at 0.5 mm/yr, which is several orders of magnitude faster than the long term rate of 11 m/my.

Figure 18. Timing of aggradation and degradation cycles revealed by terraces in different climatic settings (modified from Bull, 1991). (a) Seaward Kaikoura Range, New Zealand (Bull, 1991); (b) Musone River, Marche, Italy (Wegmann and Pazzaglia, 2009); (c) Colorado River in the eastern Grand Canyon (Anders et at., 2005); (d) San Gabriel Mountains, California (Bull, 1991); (e) Mojave Desert, California (Bull, 1991); (f) southern Israel and Sinai (Bull, 1991).

Figure 19. Stratigraphy of the Marble Canyon area in the Grand Canyon, AZ, modified from Cragun (2007). The ages of the M4 terrace, volumetrically the largest, indicates a major alluviation episode not typically associated with an important glacial-interglacial climate change. All numbers are OSL or radiocarbon ages in ky.

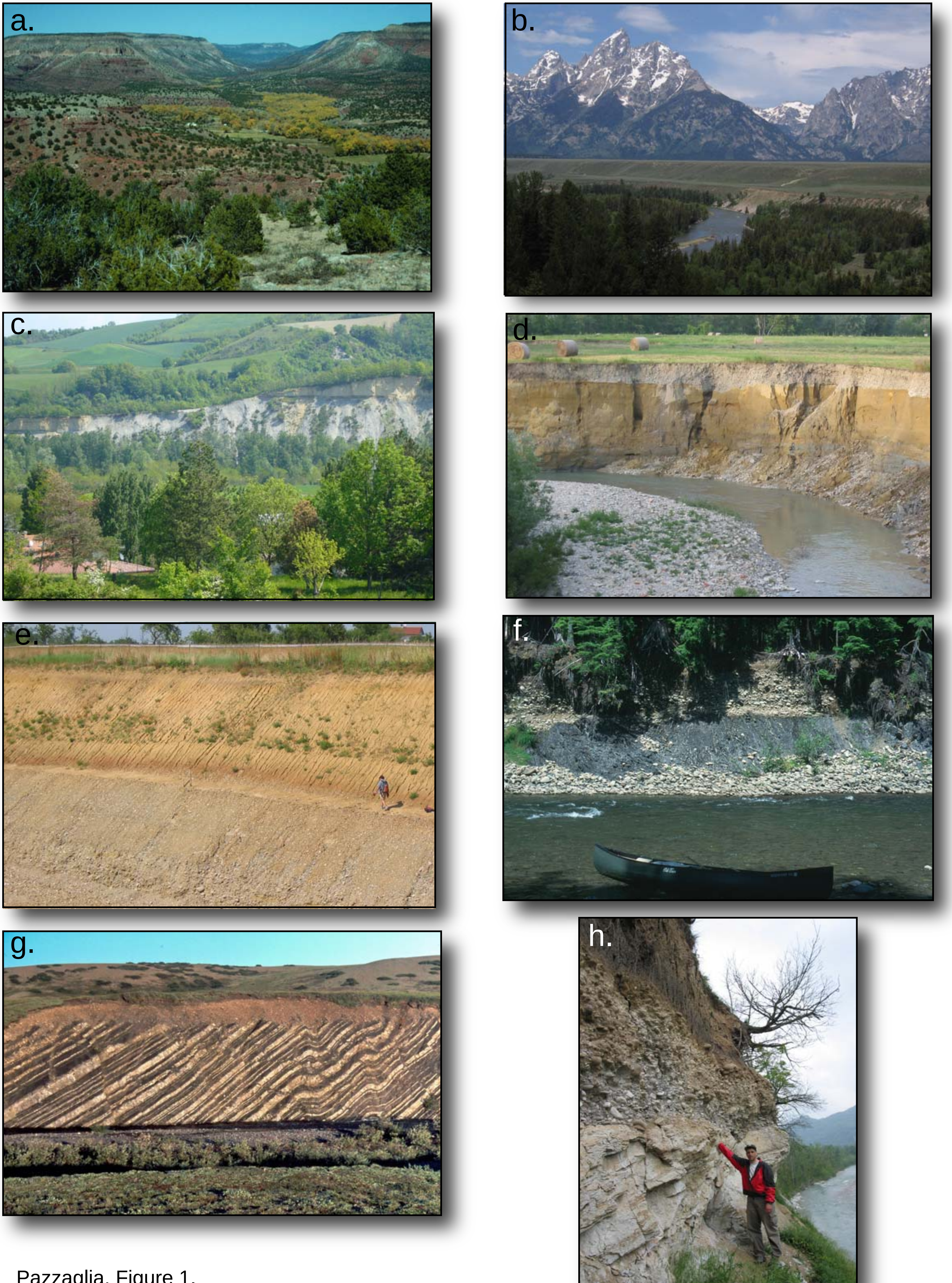
Figure 20. Comparison of diverse terrace sequences in comparable climatic settings as a result of different basin hydrology linked to rock-type. (a) strath terraces of the Bidente and fill terraces of the Musone rivers in north-central Italy. These rivers have very similar tectonic and climatic settings but differ in draining basins underlain by siliciclastics and carbonate rocks respectively (modified from Wegmann and Pazzaglia, 2009). (b) long profiles, terraces, and discharge characteristics of the of the Rio Pueblo de Taos and Rio Hondo in northern New Mexico. Both streams are tributary to the Rio Grande which has carved a deep gorge, resulting in regional base level fall. Despite being much smaller in drainage area, Rio Hondo has maintained grade with the Rio Grande because its basin is underlain by relatively impermeable crystalline rocks that generate greater mean annual discharge and bankfull discharge, in comparison to the more porous sedimentary rocks that underlie the Rio Pueblo di Taos watershed.

Figure 21. The case of an alluvial valley lacking Holocene terraces, the Jemez River, NM (modified from Formento-Trigilio and Pazzaglia, 1997 and Frankel and Pazzaglia, 2005). (a) Terraces of the Rio Jemez valley at Cañon, NM. (b) Long profiles showing long-term pattern of incision and incision rate. The channel from valley distance

35,000 m flows on the buried LGM alluvium depicted in (a).

Figure 22. Terrace genesis history for the Rio Jemez illustrating the aggradation (thin black line encapsulating alluvial deposits), lateral incision and creation of a wide strath (alluvial deposits above thick gray line), and vertical incision (thick gray line) over the past 200 ky with the MIS of Figure 5 scaled for reference. The fluvial response depicted here is representative of the response for semi-arid, non glaciated settings where cool glacial climates cause greater vegetation of hillslopes, reducing sediment flux and the transition from glacial to interglacial conditions causing major pulses of aggradation (thick black lines beneath the alluvial deposits). Terrace alluvium ages are explained in the text. Qt6 points to the current elevation of the Rio Jemez channel. Detailed stratigraphy of a Qt4 or LGM-type alluvium is described in Figure 4.

Figure 23. Model linking climate, hillslopes, fires, and fluvial responses in generating terraces in northeastern Yellowstone National Park, WY-MT. Sketch shows the geomorphic setting and processes during (a) periods of wet winters and stable year-long base flow and (b) periods of dry years, flashy river discharge, and frequent fires (modified from Meyer et al., 1992)



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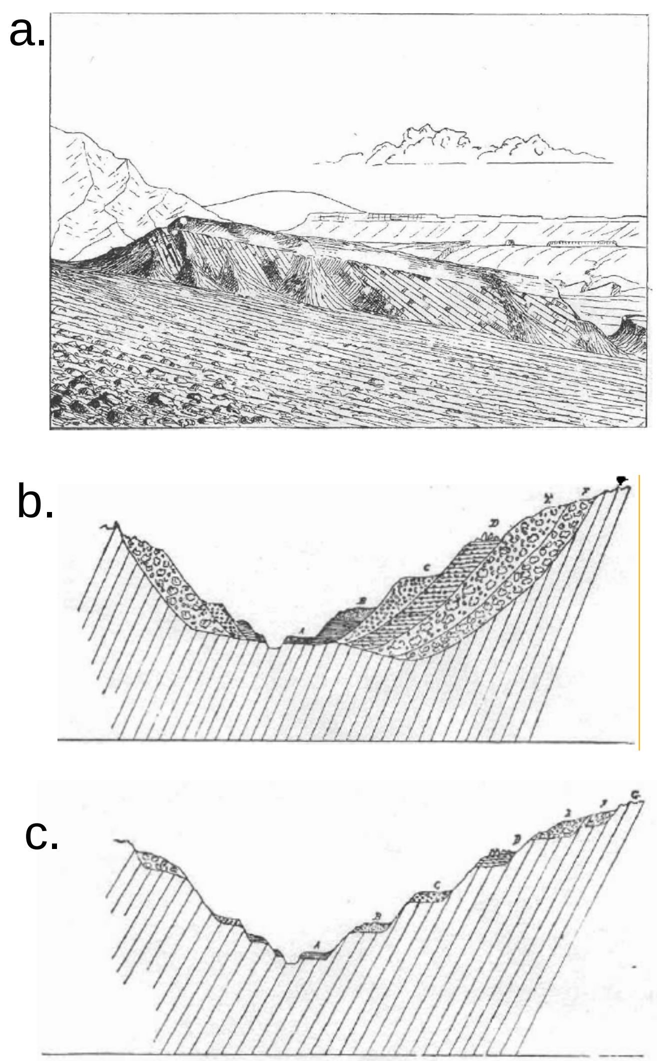

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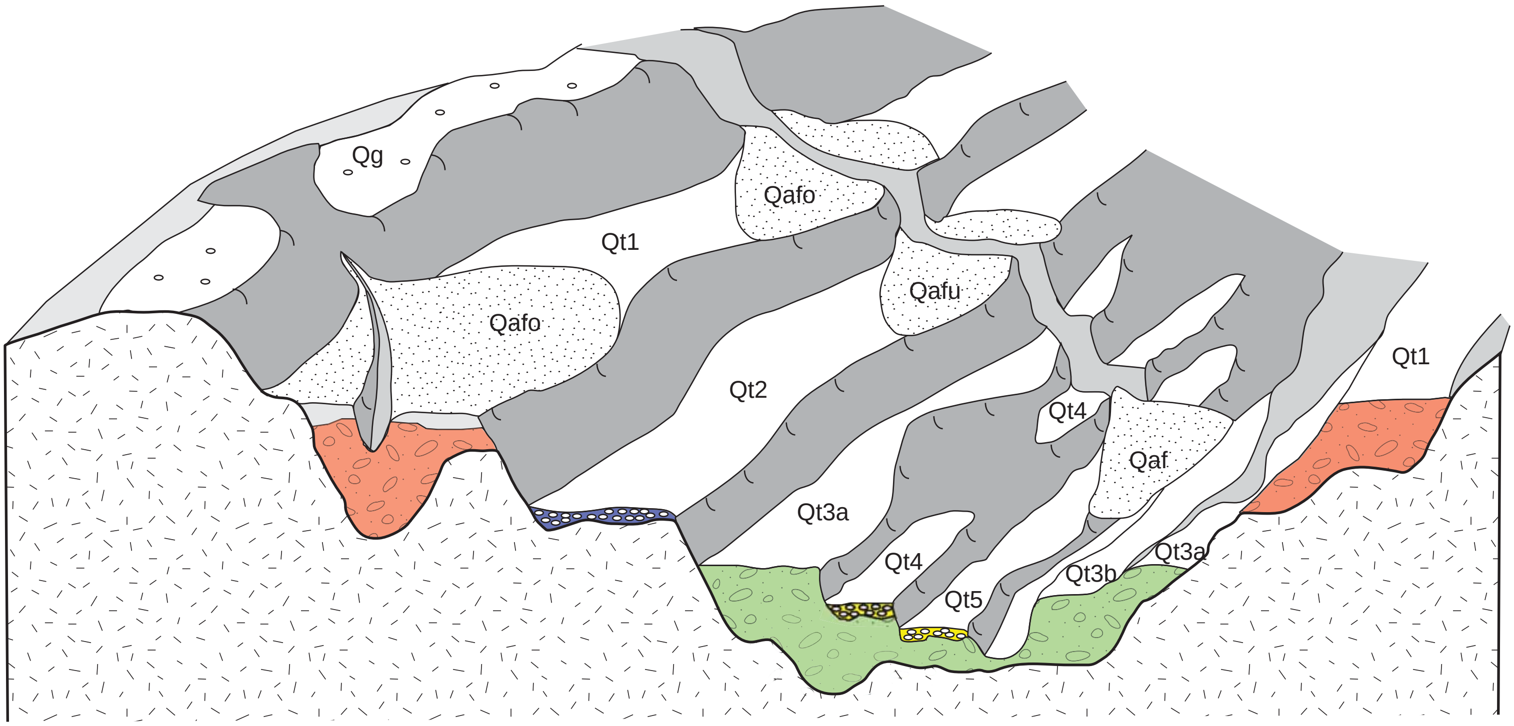

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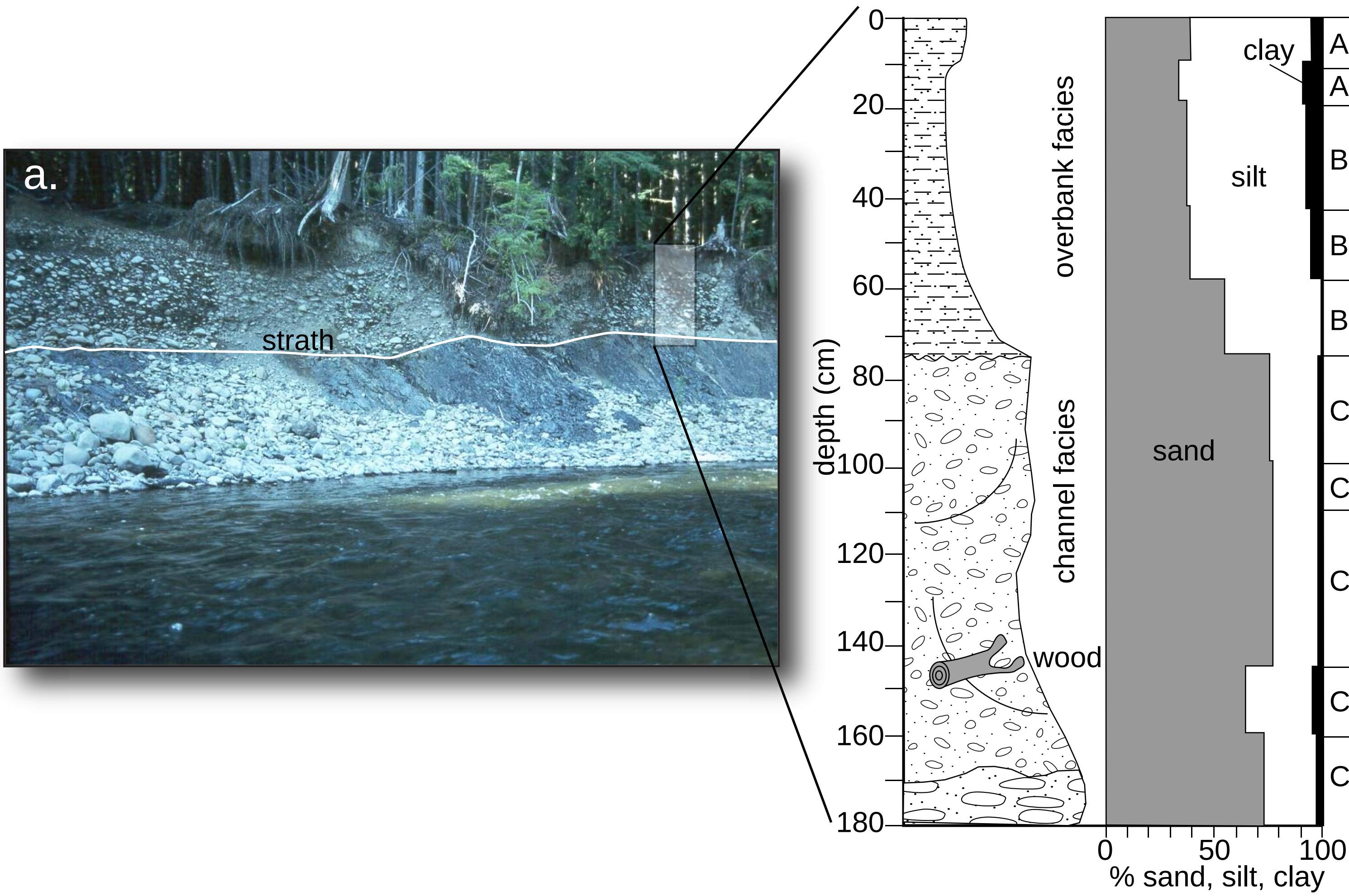

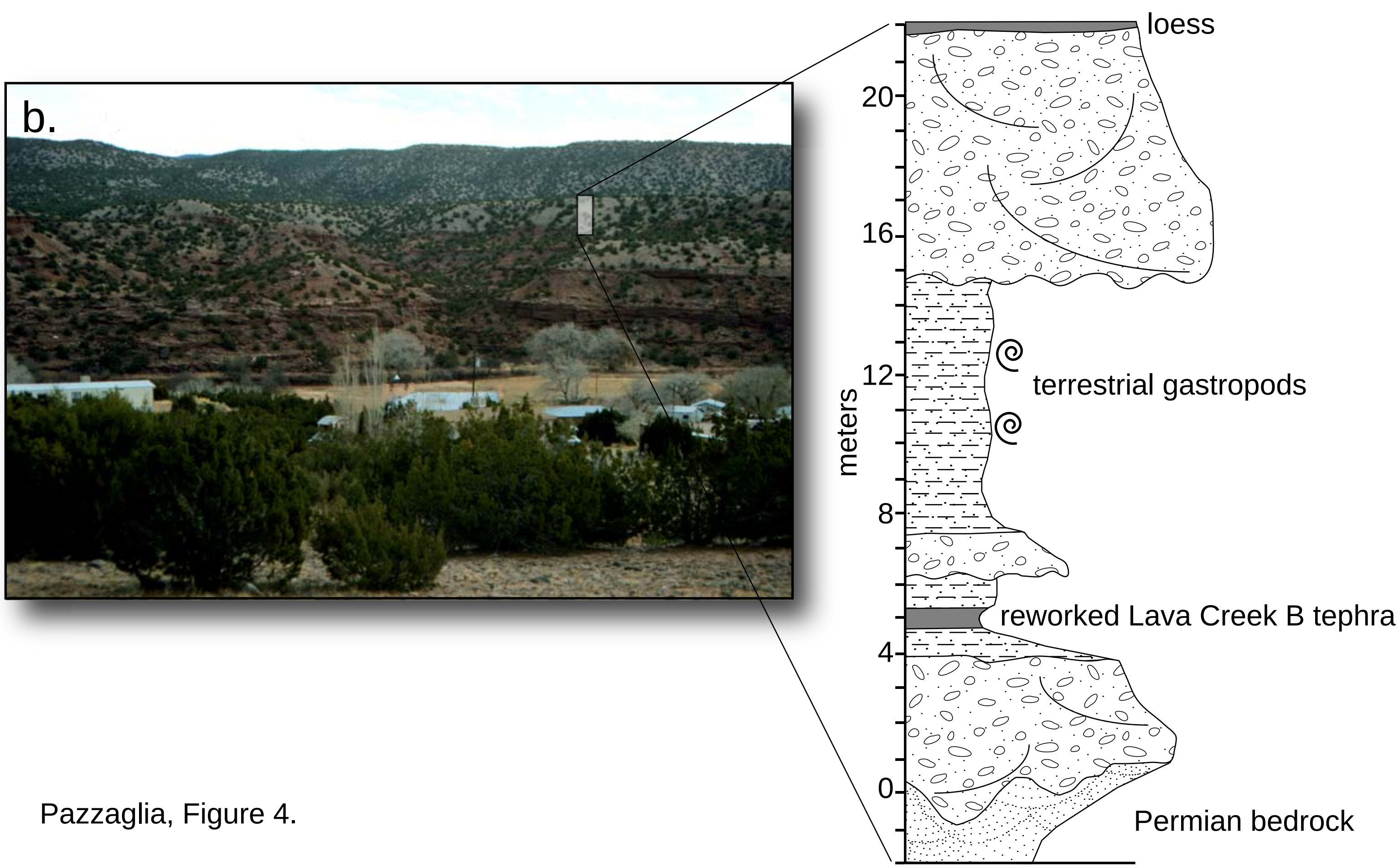

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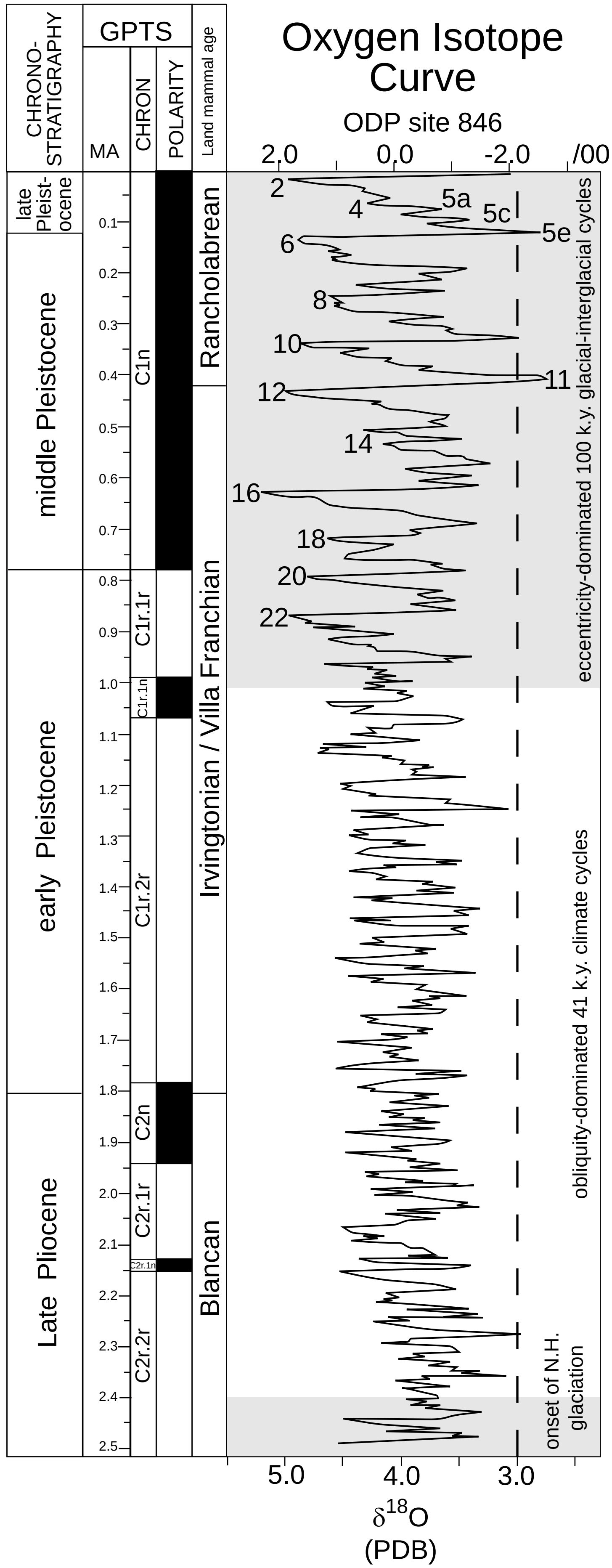

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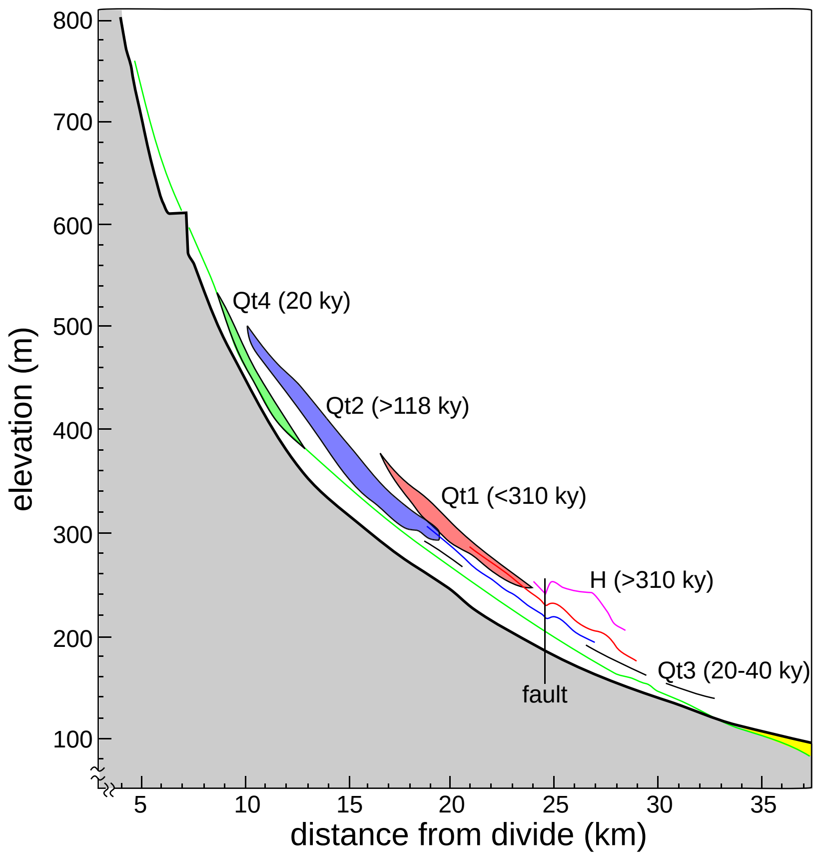
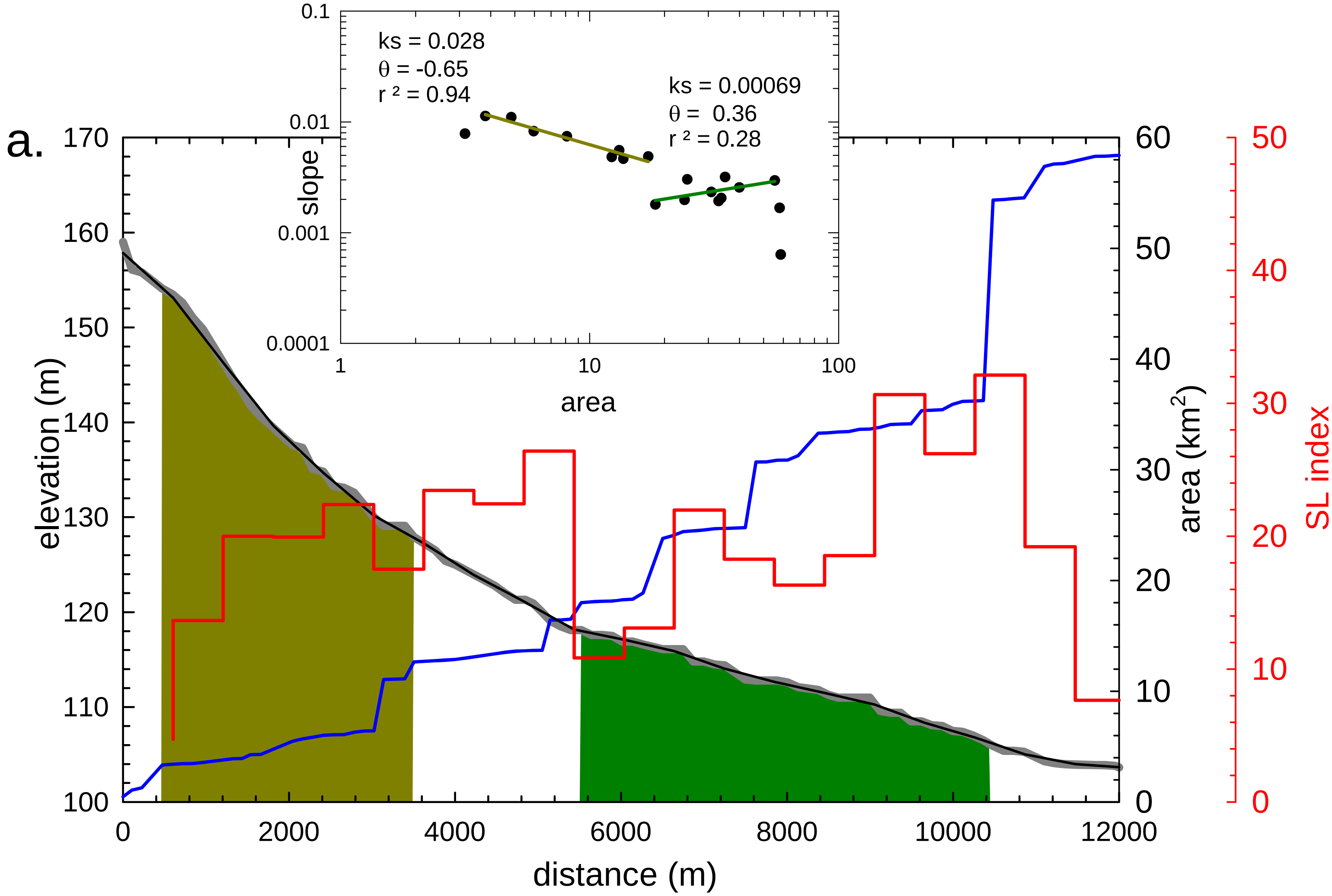


Pazzaglia, Figure 6.

Pazzaglia, Figure 7.

a.



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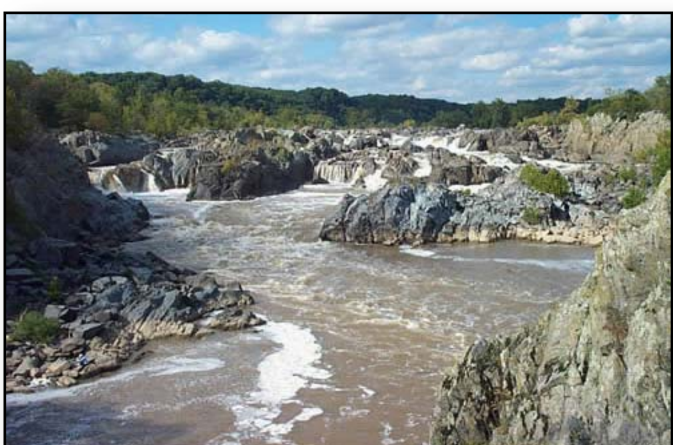

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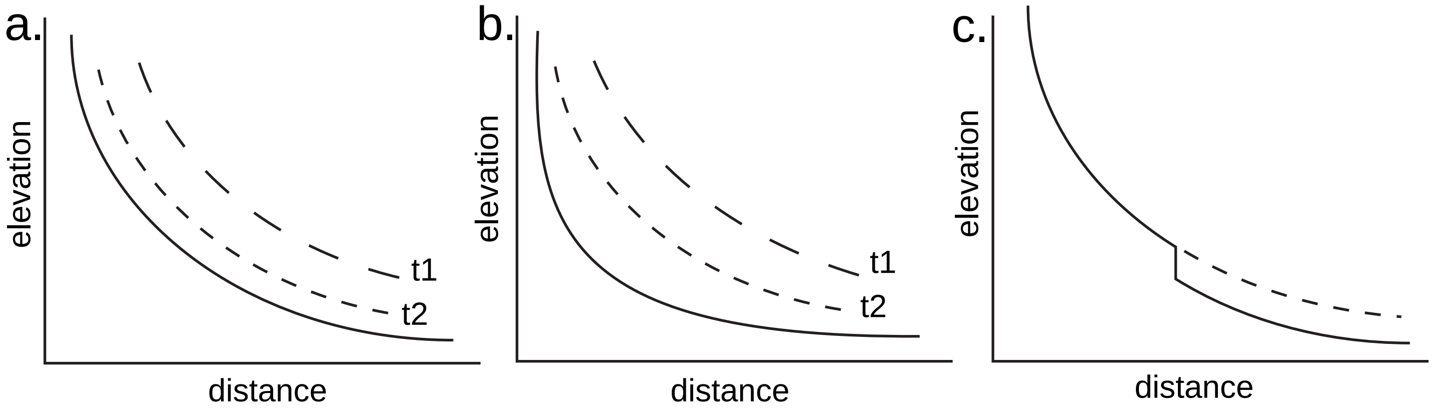

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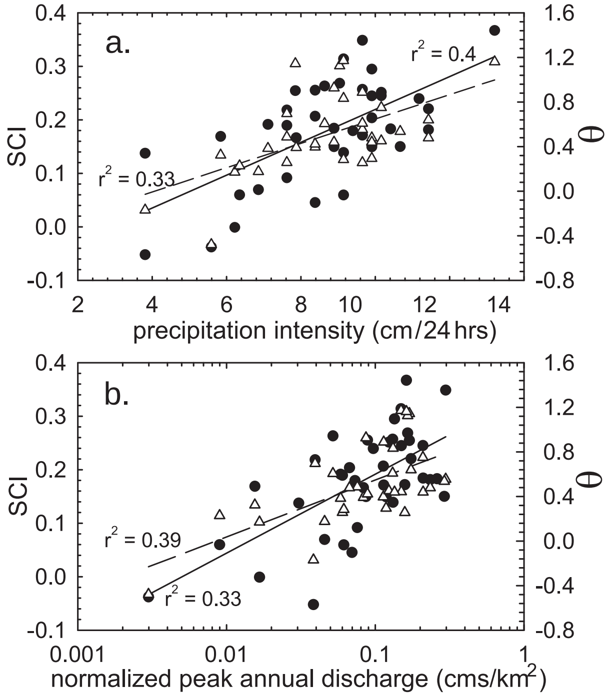

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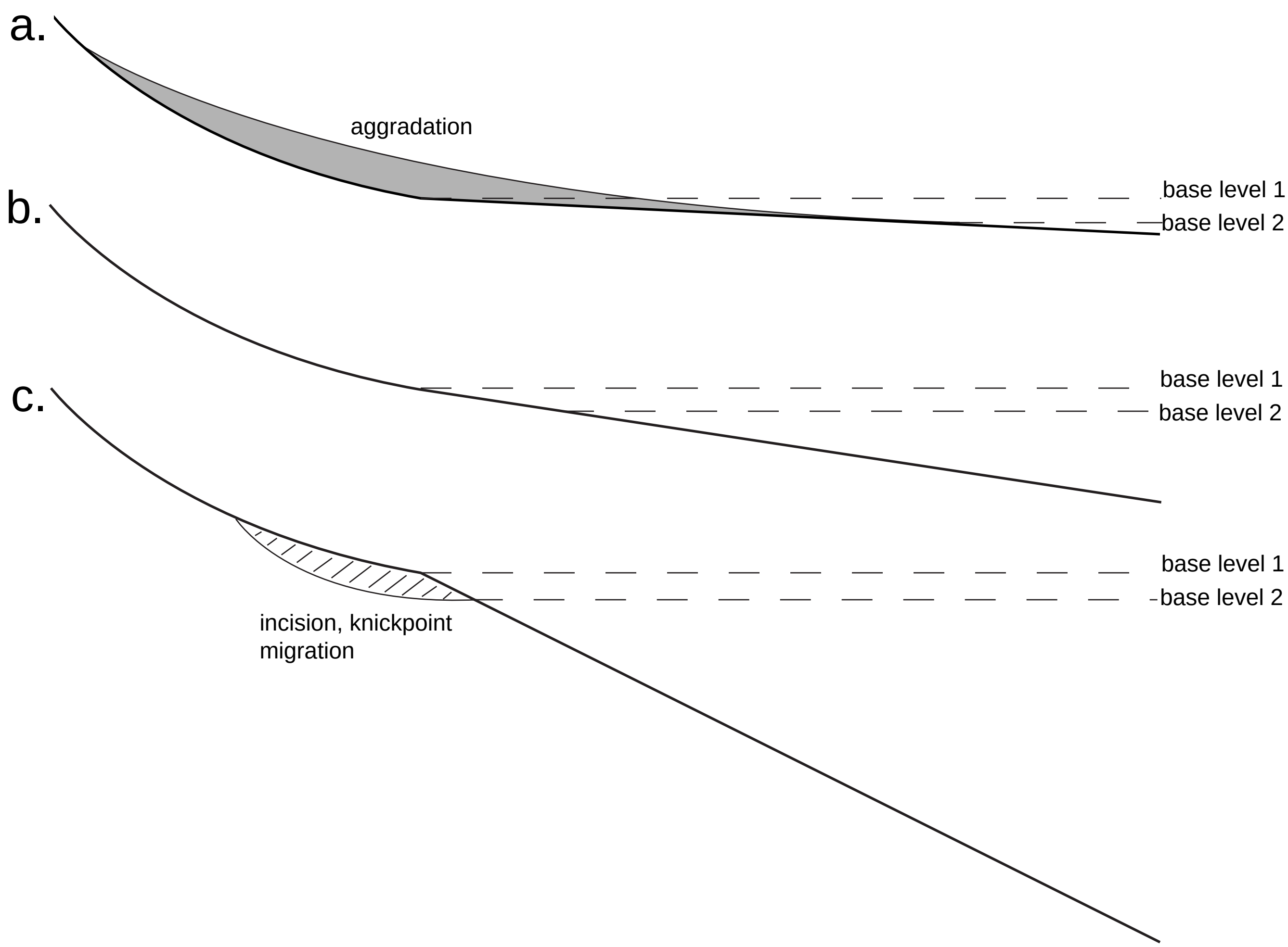

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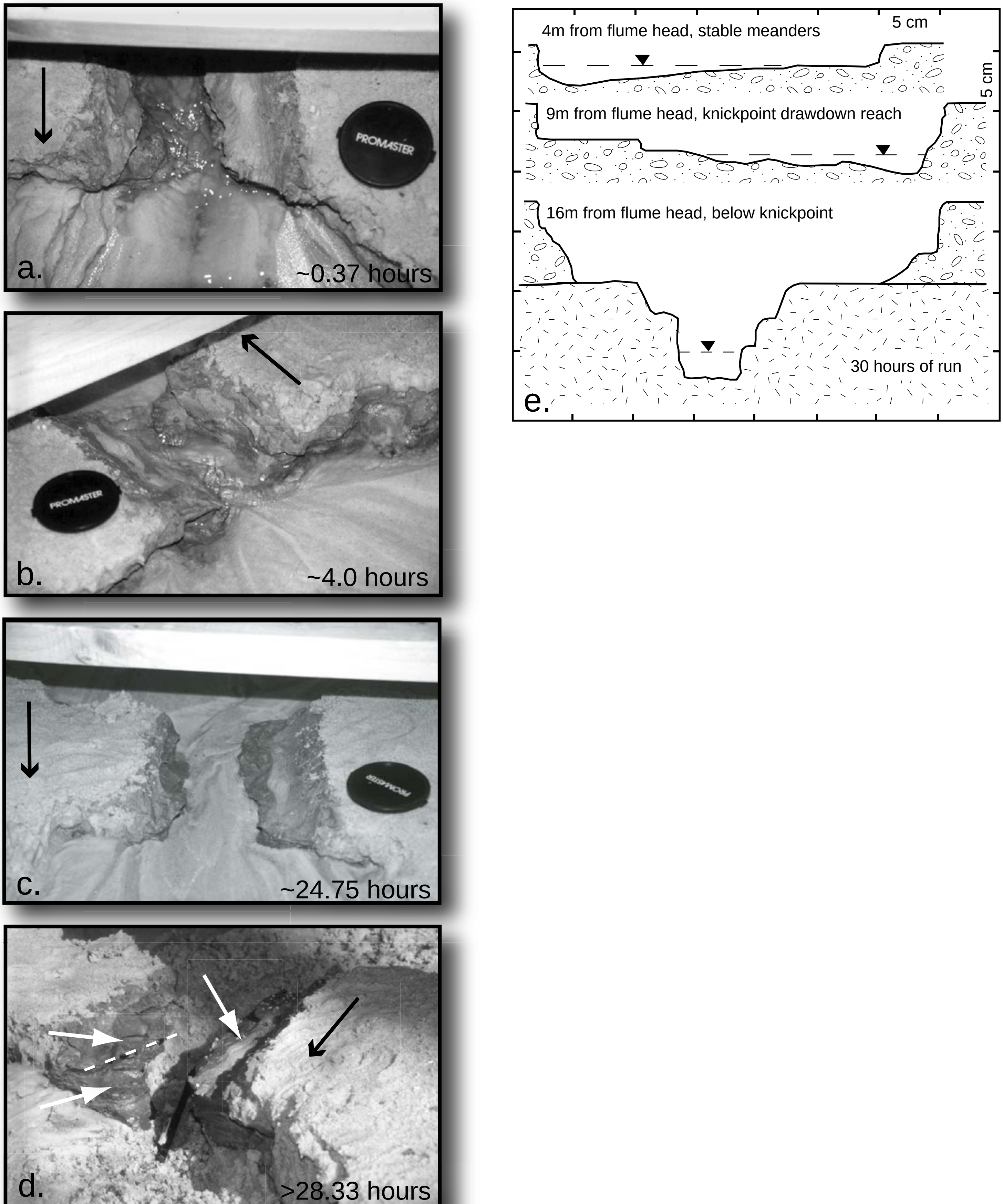
Pazzaglia, Figure 12.



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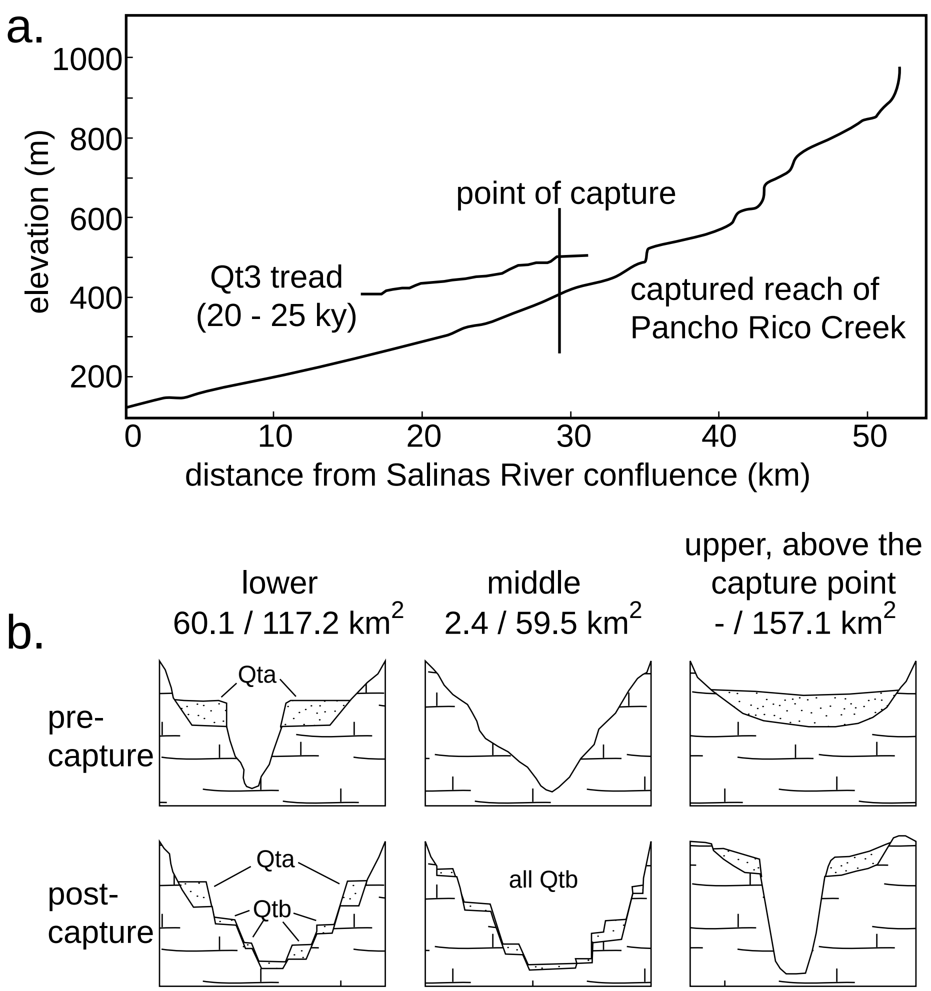

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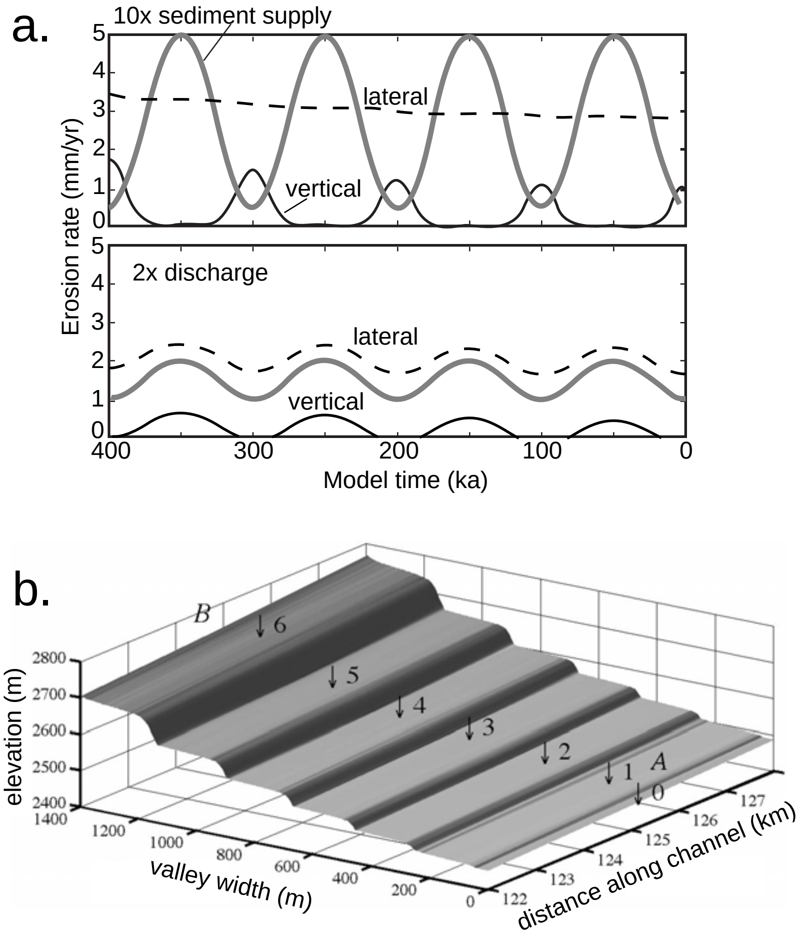

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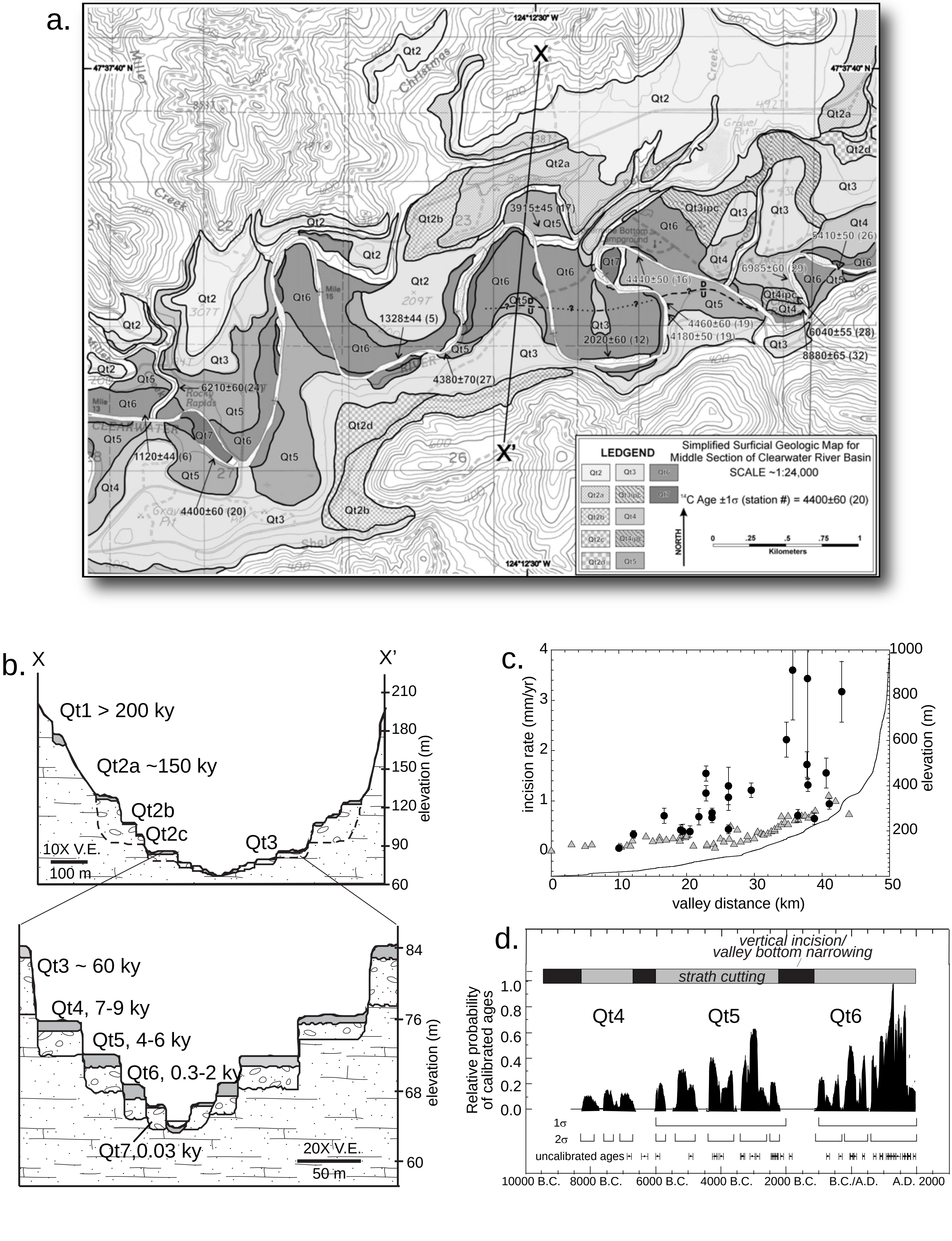

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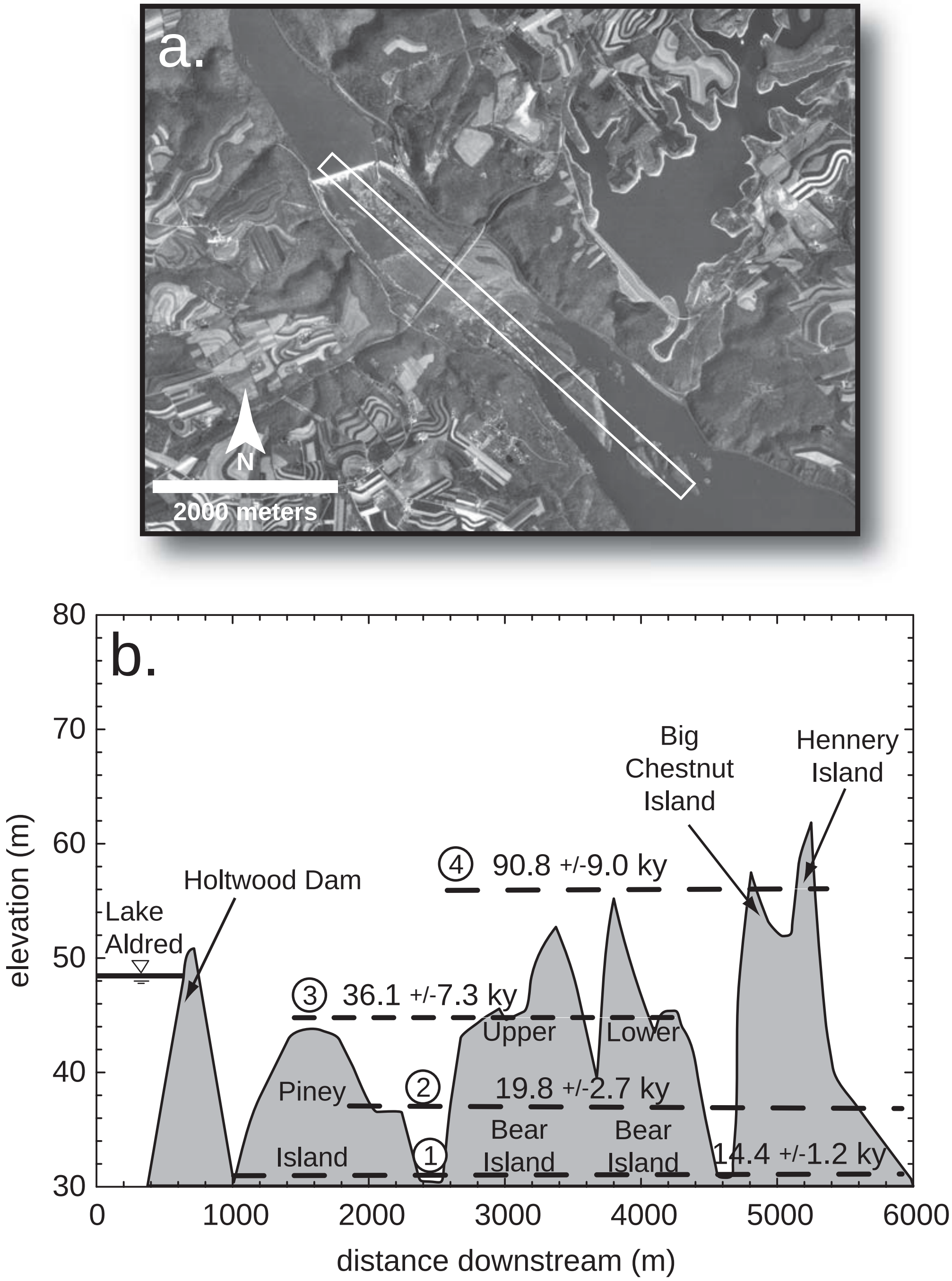

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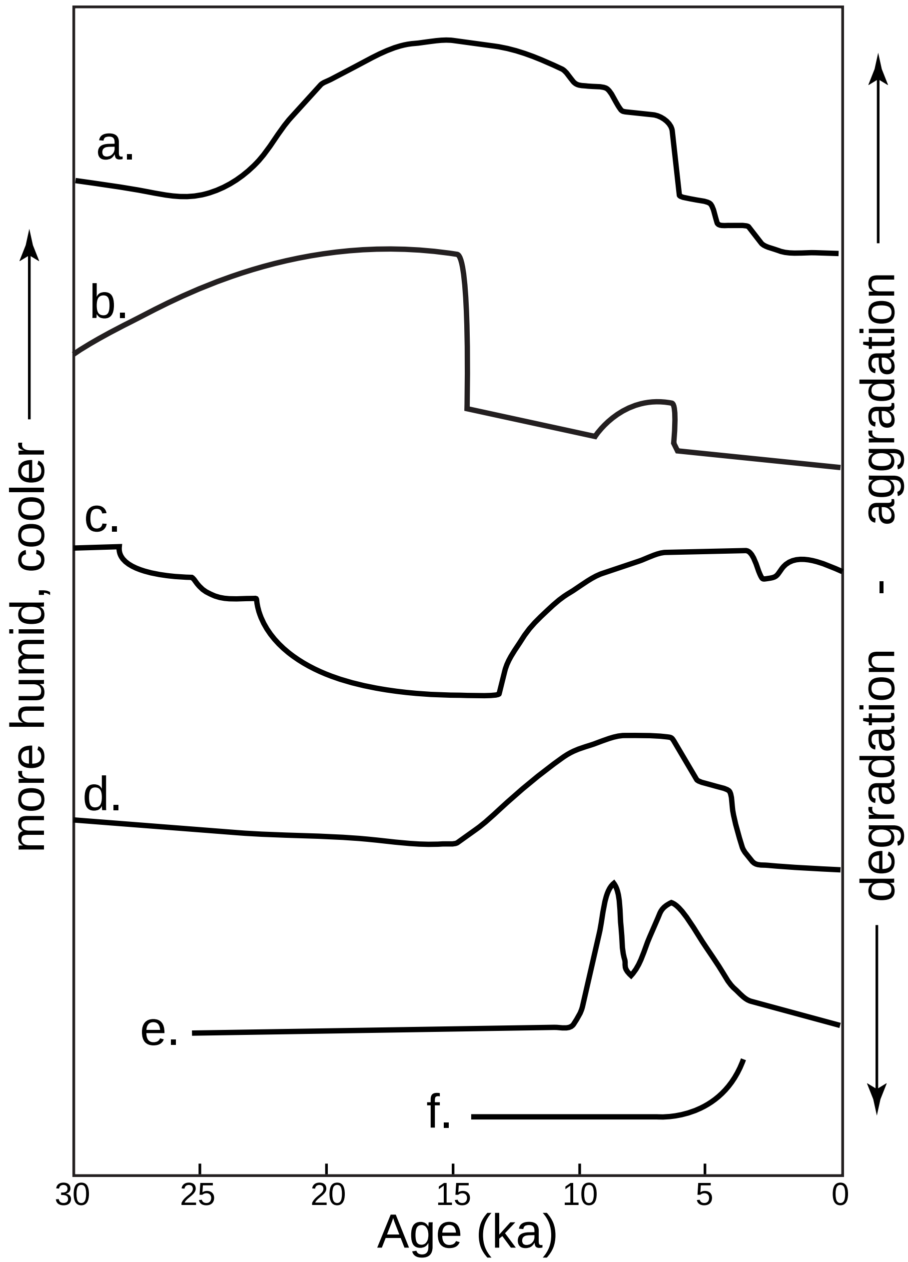

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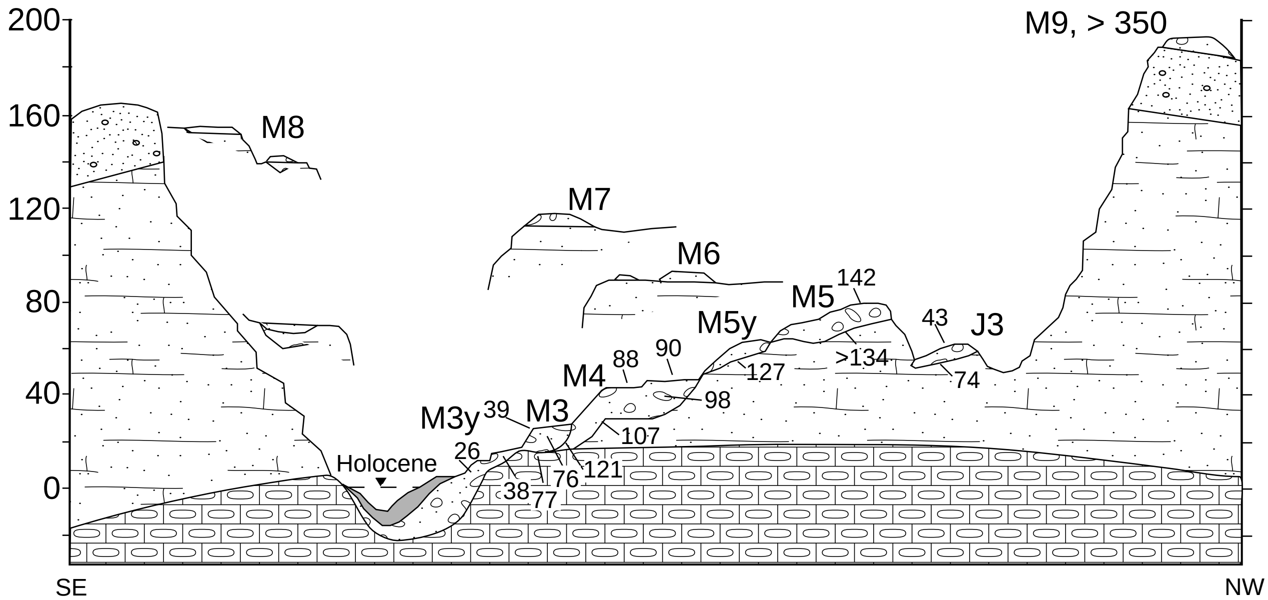

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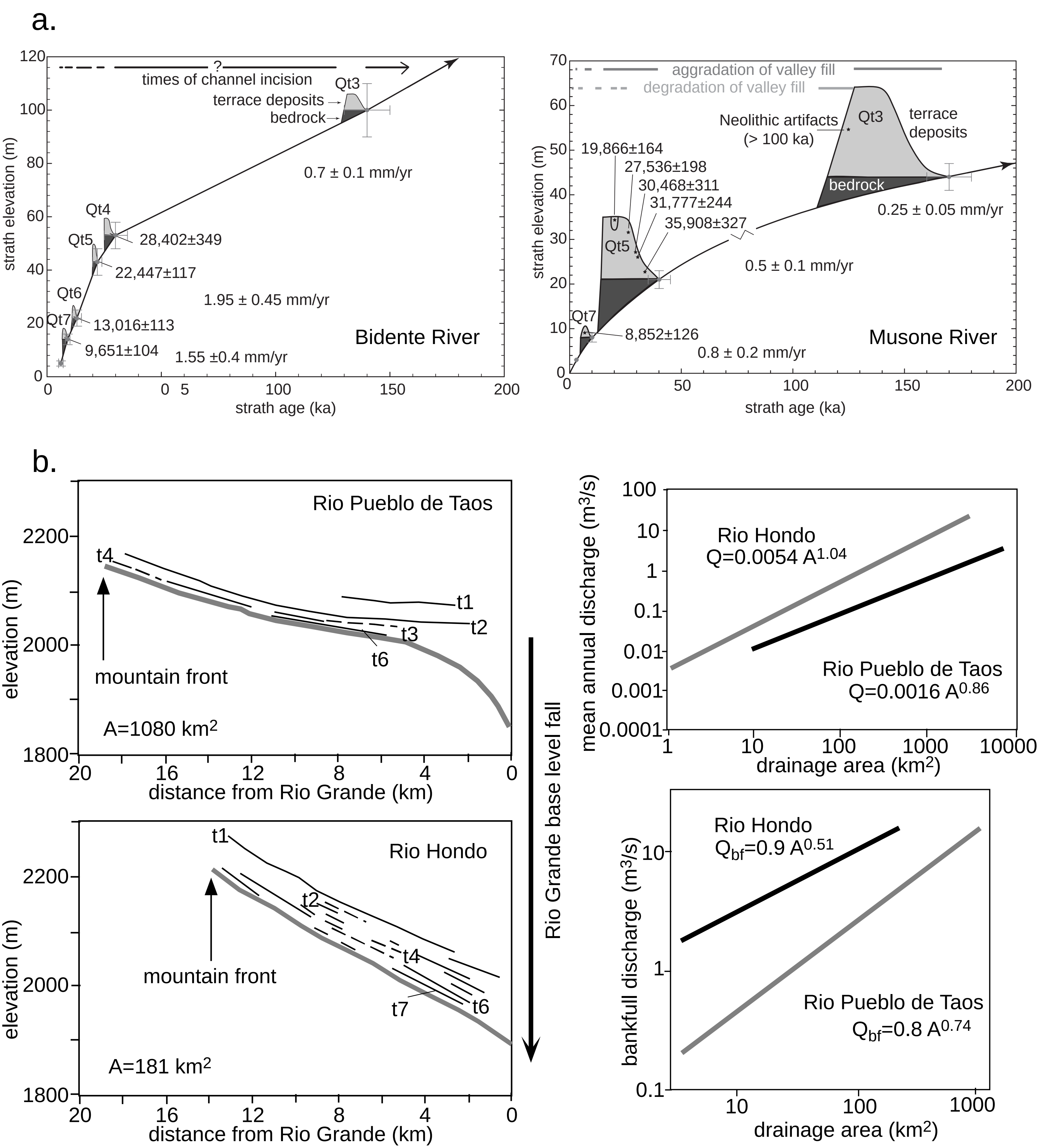

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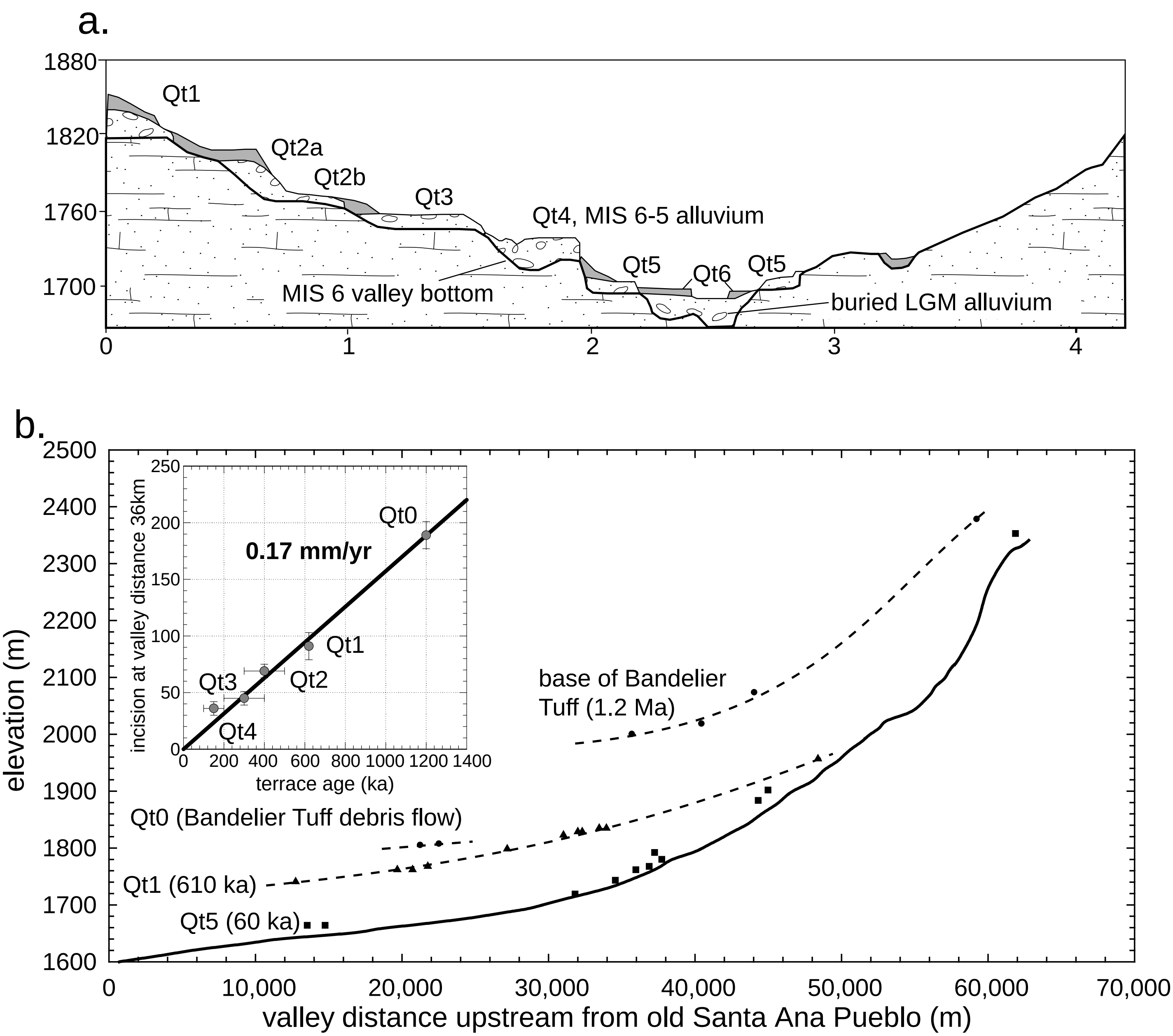

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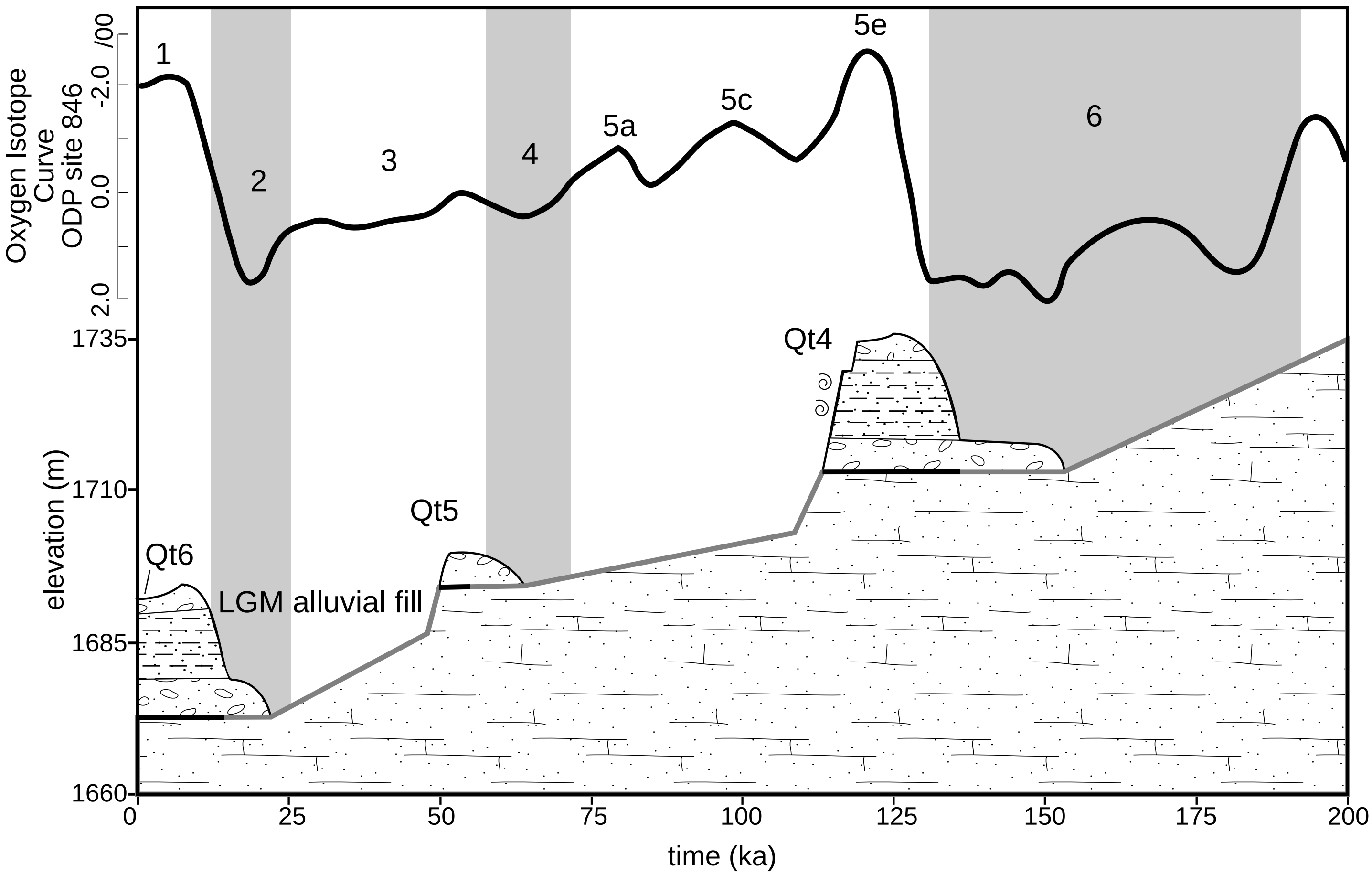

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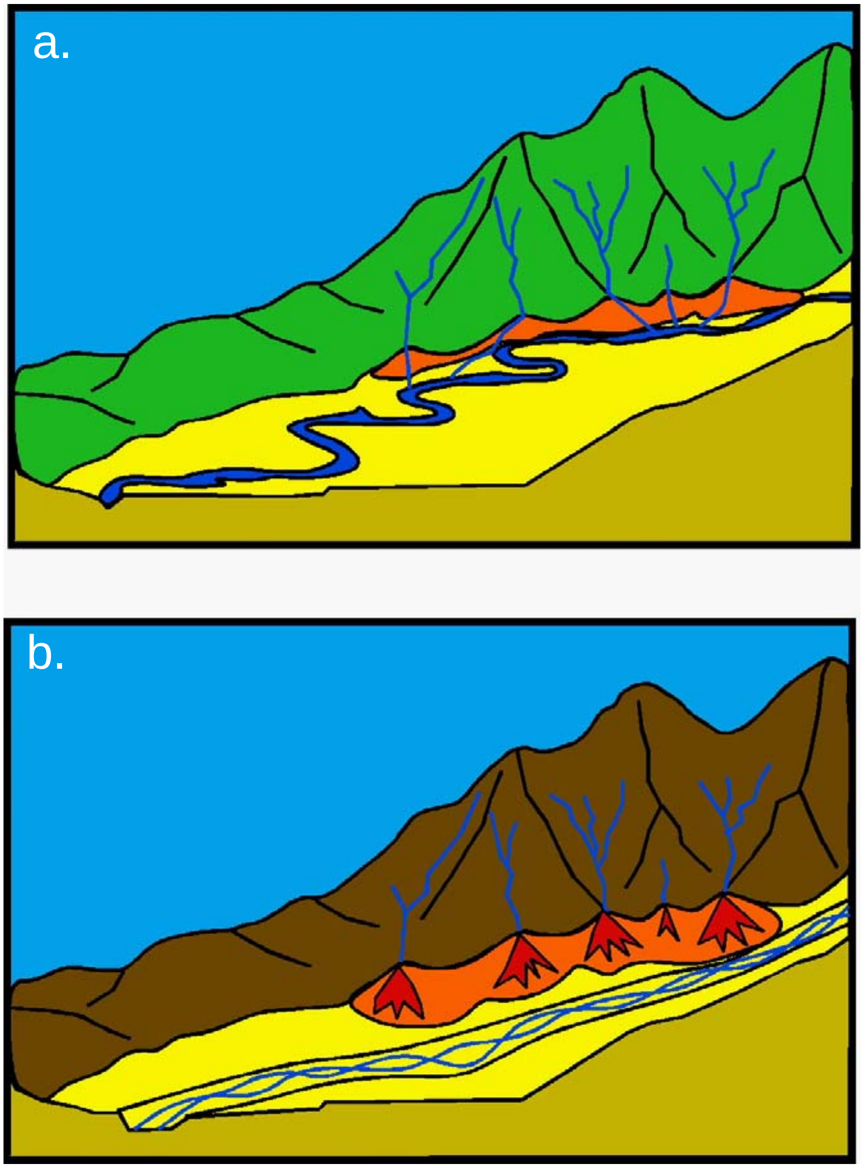

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