Badlands Morphogenesis

Dr. Pravat Kumar Shit
Assistant Professor of Geography, Raja N.L.Khan Women's College (Autonomous), Midnapore

1. INTRODUCTION

Badlands are stunning landscapes of extremely rugged terrain that have enthralled naturalists, geomorphologists and environmental scientists because of the apparent, general devastation seen in these bare or, at most, sparsely vegetated landforms. The etymology of badlands probably originates in the French expression 'mauvaises terres pour traverser', employed in the eighteenth century by early explorers in North Dakota (USA) when faced by 'bad' landscapes, in the sense of their being difficult to cross and agriculturally unproductive (Fairbridge, 1968). Contemporary badland definitions derive from similar perceptions and describe these landscapes as highly dissected areas, comprising hillslopes and divides carved in soft rock outcrops and unconsolidated sediments, with little or no vegetation, that are useless for agriculture (Bryan and Yair, 1982; Gallart et al., 2002; Moreno-de las Heras and Gallart, 2016). The immediacy and speediness of processes and the general dominance of natural phenomena in badland landscapes largely justify the attraction of badland research (Gallart et al., 2013a). Badlands usually have high drainage density, with a variety of slope and gully forms, colluvial expanses and washes generated by the interaction between bedrock weathering, mass-wasting, piping and surface fluvial processes. They offer on a miniaturized short-range spatial scale many of the landforms and processes that shape less extreme landscapes (Campbell, 1989; Howard, 2009). Surface microtopography is frequently very diverse, with surface cracks and micropipes alternating with rills, nearly vertical planes, divides and less acute forms (Yair et al., 1980; Calvo-Cases et al., 1991; Torri and Bryan, 1997; Cantón et al., 2002; Kuhn et al., 2004). Badlands differ from gully systems in the sense that the former include both hillslopes and divides, whereas gullies represent linear erosive forms (Gallart et al., 2002). Gully systems, however, can be closely related to badlands, as gully development may initiate or reactivate badland dynamics (Nogueras et al., 2000; Howard, 2009; Wainwright and Brazier, 2011; Ballesteros-Cánovas et al., 2017). Badlands typically constitute minor sections of the terrain. For example, large-scale regional surveys indicate that badlands cover about 2.5% of desert landscapes in southwestern United States and just a few thousand square kms. in the Mediterranean basin (Clements et al., 1957; Grove and Rackham, 2001). Despite covering in general little terrain, badlands are typically perceived as hot spots of sediment production at broad, regional scales. A recent review of more than 55 badland erosion studies of the Mediterranean basin highlights the frequent occurrence of intensely active erosion processes spanning rates near to and over 100 t ha−1 yr−1 at field measurement scales between 10−4 and 103 ha (Nadal-Romero et al., 2011, 2014a). These large sediment production rates have important geomorphological repercussions for downstream environments. For example, it has been argued that badlands play a key role in organizing the sedimentary structure of rivers and maintaining river deltas (Clotet, 1984; Grove and Rackham, 2001; López-Tarazón et al., 2012). However, care should be taken when specifying badlands' erosive activity because both field observations and modelling studies have revealed rather low denudation rates for large areas at a variety of ancient badland sites, suggesting that these landscapes may have remained relatively stable over millennia (Wise et al., 1982; de Ploey, 1992; Yair et al., 1982; Howard, 1997; Díaz-Hernández and Juliá, 2006). In spite of having common morphological features, badlands can show very diverse geomorphic activity and dynamic behaviour, which highlights the complexity of the processes and interactions involved in the origin and evolution of these intensely dissected landforms (Torri et al., 2000; Gallart et al., 2013a; Yair et al., 2013). A variety of factors have been suggested to play a significant control in badland formation. Climate largely mediates the balance between erosive power and vegetation control in hillslope and badland systems, providing denudational instability (Thornes, 1985; Alexander et al., 1994; Gallart et al., 2002; Bochet et al., 2009). Regional tectonics and lithology exert a major role on badland-formation processes, providing topographic gradient, base-level conditions and bedrock susceptibility to both weathering and erosion (Harvey, 1987; Campbell, 1989; Grove and Rackham, 2001; Kasanin-Grubin, 2013). Extensive badland areas may occur in particular climates, over soft saline bedrocks that impede the development of vegetation or as a consequence of regional base-level change (Calvo-Cases and Harvey, 1996; Howard, 2009; Wainwright and Brazier, 2011), but they can also occur as secondary forms generated by local triggers such as extreme-event deluges or human activity (Gallart and Clotet, 1988; Wainwright, 1996; Ballesteros-Cánovas et al., 2017). Except in the rare cases where it is possible to provide direct evidence or a sound rationale for the relationships between these factors and the triggering of badland initiation mechanisms, usually only established assumptions or indirect signs are applied. This chapter aims to analyze the origin of badlands, exploring the main factors, processes and mechanisms that control badland formation. We tackle this objective by, first, analysing the climate distribution of badland occurrence; second, reviewing dominant badland-shaping processes; and last, studying a set of initiation patterns and major triggering factors involved in the development and evolution of badlands.

2. CLIMATE DISTRIBUTION OF BADLANDS

The oft-dissected geomorphology of badlands and their very frequent lack of vegetation are commonly associated with the incidence of arid and semiarid climate conditions (Bryan and Yair, 1982; Warren, 1984; Campbell, 1989). Badlands, however, can be found under a broad variety of climate conditions over soft bedrock and unconsolidated sediments where vegetation is disturbed or naturally lacking (Grove and Rackham, 2001; Gallart et al., 2002; Howard, 2009). For example, badland occurrence around the Mediterranean basin reveals that these erosive landforms are scattered over a broad array of areas distributed under the full regional range of climate conditions (Fig. 2.1A). Mediterranean badlands can be found in arid regions of Israel and Morocco, in semiarid and dry-subhumid areas of Spain (e.g., extensively in the central Ebro basin and the southeastern dryland regions), southern Italy, eastern Greece, and inland Turkey, and in subhumid and humid conditions in the Pyrenees, Alps, Apennines, and Pindus mountain ranges. Beyond the Mediterranean basin, badlands are common in dryland regions of North America (e.g., the Dinosaur Park, Henry Mountains, and White River badlands; Bryan et al., 1987; Howard, 2009; Benton et al., 2015) and Africa (Feoli et al., 2002; Boardman et al., 2003; Tooth et al., 2013), but also in humid areas of China, Japan, New Zealand, and South America (Lam, 1977; Lin and Oguchi, 2004; Parkner et al., 2007; Higuchi et al., 2013; Hermelin, 2016). Gallart et al. (2002) identified three homogeneous groups of badland types as a function of their climate distribution, general characteristics, and dominant processes:

1. Arid badlands, which typically occur in desert areas with mean annual precipitation (MAP) below 200mm and MAP to potential evapotranspiration (PET) ratio values up to 0.20. The Zin Valley (northern Negev, Israel) and Henry Mountains (Utah, USA) badlands are well-known examples of this group of desert-type badlands in the Mediterranean basin and North America, respectively (Howard, 1997; Yair et al., 2013). Vegetation, largely lacking in these desert systems, plays no relevant role in arid badland dynamics. Arid badland surfaces can, to some extent, be stabilized by (cyanobacteria, algae or lichen) biological crusts, which may modulate water infiltration and erosion rates (Belnap, 2006). Geomorphic processes in desert-type badlands are, however, primarily explained by abiotic control of soil surface spatiotemporal variability (Yair et al., 1982; Kuhn et al., 2004; Howard, 2009). In general, arid badlands are very ancient landscapes apparently stabilized under present climate conditions. For example, the desert-type Zin Valley badlands (Fig. 2.1B, MAP∼90 mm yr−1), originating during a less arid phase in the late Pleistocene, have shown throughout the last 20,000 years of evolution rather low channel incision rates (∼0.25mmyr−1) and, at present, no significant slope retreat despite locally active badland ridge lowering by erosion (Kuhn et al., 2004; Yair et al., 1980, 1982, 2013).

2. Semiarid and dry-subhumid badlands, which are developed in dryland areas with 200–700mm MAP and 0.20–0.65MAP to PET ratio. Interactions between climate, vegetation dynamics, and site geomorphology explain the preferential distribution and higher erosion activity of semiarid and dry-transition badlands on sunny hillslopes, more exposed to solar radiation, where vegetation development is strongly limited by scarce soil moisture and, therefore, cannot exert any control over soil erosion processes (Kirkby et al., 1990; del Prete et al., 1997; García-Fayos et al., 2000; Cantón et al., 2002; Bochet et al., 2009). The Tabernas (southeastern Spain) and Basilicata (southern Italy) badlands (Fig. 2.1C and D, respectively) constitute good examples of this dry-type group of badlands (Piccarreta et al., 2006; CalvoCases et al., 2014). Semiarid and dry-subhumid badlands are, in general, quite old – over 40,000years old in some cases – and may have experienced different (tectonic and/or climatic) phases of incision and stabilization (Wise et al., 1982; Harvey, 1987; Díaz-Hernández and Juliá, 2006; Piccarreta et al., 2011). Despite the general natural origin of most dry badlands, high erosion susceptibility at this precipitation range (Langbein and Schumm, 1958) makes semiarid and dry-subhumid areas sensitive to badland enlargement or initiation by human-induced degradation (Gallart, 1992; Piccarreta et al., 2011; Ballesteros-Cánovas et al., 2017).

3. Subhumid and humid badlands, very frequently mountainous in character, which are distributed in areas with over 700mm MAP and above 0.65MAP to PET ratio. Vegetation growth in these montane erosive landforms is not limited by the availability of water but may be affected by the high erosion activity of these systems and their cold-season harsh thermal conditions, which reduce the vegetative period, particularly on shady hillslopes (Guàrdia et al., 2000; Nadal-Romero et al., 2014b). Unlike dry badlands, subhumid and humid badlands are not clearly asymmetrical and in some cases may even show (reverse) preferential distribution on shady aspect (Regüés et al., 2000; Moreno-de las Heras and Gallart, 2016). Usually much younger than desert and dry badlands, subhumid and humid badlands generally show higher erosion rates caused by deep weathering and high annual precipitation on bare soft bedrock. Average denudation rates of about 10mmyr−1 and over are frequently reported for these relatively young, rainy badland systems (Nadal-Romero and Regüés, 2010; Descroix and Mathys, 2003; Gallart et al., 2013b; Vericat et al., 2014). Good examples of this type of badlands can be found in the main mountain ranges of the Mediterranean basin, for instance, the Araguás, Isabena, and upper Llobregat (Fig. 2.1E) badlands, located in the southern Pyrenees (Nadal-Romero and Regüés, 2010; López-Tarazón et al., 2012; Moreno-de las Heras and Gallart, 2016), or the terres noires badlands of Draix, in the French Alps (Descroix and Mathys, 2003).

3. BADLAND-SHAPING PROCESSES

Detailed badland forms result from the interaction of soft or unconsolidated materials and multiple geomorphological processes, very frequently acting in combination on different temporal and spatial scales. Overall, bedrock weathering, surface fluvial processes (sheet washing, rilling and gullying), subsurface erosion (piping, tunnelling) and mass-wasting processes (creeping, landsliding), sometimes concealed in short-range morphological features (e.g., crusts, cracks, rills, micropipes) on

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Assistant Professor of Geography, Raja N.L.Khan Women's College (Autonomous), Midnapore

and under the badland surfaces, dominate the evolution of the landforms in these erosive systems. Bedrock weathering, frequently neglected in badland papers, represents the first badland-shaping process that exerts a dynamic control on these erosive landforms. Badland erosion generally takes place on a regolith or crust developed by bedrock weathering (Gallart et al., 2002; Kasanin-Grubin and Bryan, 2007). The relevance of the weathering process to badland dynamics is clear even for those badlands that are developed in poorly consolidated mudstones, such as the southern Taiwan badlands of the (Plio-Quaternary) Lower Gutingkeng Fm. (Higuchi et al., 2013). Annual badland denudation in these monsoonal systems (about 12cmyr−1) matches approximately the amount or thickness of regolith formed throughout the dry season by repeated wetting–drying cycles, suggesting that erosion in these very humid (∼2000mmyr−1 precipitation) systems is limited by weathering. For the Mediterranean mountain badlands of Vallcebre (eastern Pyrenees), weathering of moderately preconsolidated (late Cretaceous) mudstones of the Tremp Fm. is dominated by the effects of cold-season freezing–thawing cycles (Fig. 2.2A; Regüés et al., 1995; Pardini et al., 1996; Regüés and Gallart, 2004). Erosion rates in these fairly humid (∼850mmyr−1 precipitation) badland systems may be weatheringor transport-limited depending on the intensity of winter bedrock weathering and further summer incidence of high-intensity erosive rainstorms (Gallart et al., 2013b; Moreno-de las Heras et al., 2018). The importance of freeze–thaw weathering of bedrock (also called gelivation), as occurs in the Vallcebre badlands, was cited early in badland literature (e.g., Schumm, 1956, 1964) and has been broadly recognized as a major badland-shaping process in mountain badlands (Descroix and Mathys, 2003; Nadal-Romero and Regüés, 2010; Moreno-de las Heras and Gallart, 2016). In the semiarid badlands of El Cautivo field station (Tabernas, SE Spain), Cantón et al. (2001) found that weathering of (late Miocene) gypsiferous mudstones of marine origin was primarily caused by swelling–shrinking and salt hydration– crystallization, both induced by wetting–drying cycles (Fig. 2.2B). Fresh mudrock weathering rates were found to be proportional to the number of rainfall events during the sampling periods, averaging from 0.7 to 8mmyr−1. Annual erosion in these dry (∼300mmyr−1 precipitation) badlands is considered to be transport-limited, although badland incision for the area is probably within the range of local weathering rates under present semiarid conditions. Wetting–drying cycles also dominate bedrock weathering in the desert-type, Zin Valley badlands (northern Negev desert, Israel) developed on smectite-rich materials of the (Palaeocene) Taqyia Fm. On the badland divides, the regolith layer is thin (more rapidly after rainfall, limiting in situ regolith development. These variations in regolith thickness, along with the occurrence of desiccation cracks and pipes, cause strong spatial differences and discontinuities in water and sediment transfer on the hillslopes (Yair et al., 1980). Erosion rates in these very dry (∼90mmyr−1 precipitation) desert badlands are largely limited by transport, which provides a physically based explanation for their apparent geomorphological stability under current arid regional climate conditions (Yair et al., 2013). Like bedrock weathering, particle detachment and downslope transport by rain splash are common badland-shaping processes, which have been reported little in published studies. Direct evidence of the importance of rain splash on badland surface particle detachment is seen in how rainfall simulations in small plots (too small to be able to produce rills) can produce very large sediment concentrations on runoff (Solé-Benet et al., 1997; Regüés and Gallart, 2004; Yair et al., 2013). Indirect evidence of rain splash on badland surfaces is also provided by the emergence of micro-hoodoos and pinnacles below the protection of stones or vegetation fragments after rainfall events (Regüés et al., 1995; Desir and Marin, 2003). Overland flow generation is, on the other hand, a set of processes very commonly discussed in the literature. The reason is that, far from simple behaviour, runoff generation and routing in badland surfaces may vary strongly, both temporally and spatially. This depends, among other things, on detailed lithology, rainfall depth from the beginning of the events and regolith conditions (Yair et al., 1980; Hodges and Bryan, 1982; Regüés and Gallart, 2004). High-intensity or copious rainfall are the most common forces triggering runoff at badland sites, but snowmelt may also trigger it in high-altitude/latitude and inner-land continental climates (Hodges and Bryan, 1982; Faulkner, 1994). Surface fluvial processes have been traditionally evoked as the main landformshaping processes of badland settings (de Ploey, 1992; Howard, 1997). As overland flow moves away from the badland divides, runoff depth increases, causing sheet erosion and further rill and gully incision. Sheet wash, rill and gully erosion are undoubtedly major badland-evolution processes. The hydraulics of these surfaceerosion processes may change markedly due to regolith characteristics, causing transport or supply-limited flow conditions (Howard, 2009). Rill flow generation is yet more intricate because rills can also be fed by subsurface flow through cracks and micropipes that may open and close within a rainfall event or at the larger, seasonal scale (Hodges and Bryan, 1982). Furthermore, rills may form and disappear periodically from badland surfaces, particularly during intense bedrock weathering periods (e.g., cold-season gelivation periods) or episodic mudslide events that largely reshape badland surfaces (Fig. 2.2C; Schumm, 1956; Regüés et al., 1995; Desir and Marín, 2013; Gallart et al., 2013b). Linear propagation of the elementary drainage network, frequently in the form of gully erosion, is also considered a necessary fluvial process controlling the development of badland landscapes (Harvey, 1992; Nogueras et al., 2000). In fact, gullying is largely perceived as a decisive badland process, providing these erosive landscapes with the highly dissected shape and drainage density that characterize their landforms (Howard, 2009; Wainwright and Brazier, 2011).

Subsurface erosion, including micropipe development, piping and tunnelling, are very frequent processes in many badlands (Fig. 2.2D), particularly in those erosive systems developed on dispersive geological materials (Bryan and Yair, 1982; Campbell, 1989; García-Ruiz, 2011; Faulkner, 2013). Faulkner (Chapter 6) argues that most badlands showing typical fluvial-like landforms on marine-sourced dispersive lithology were actually initiated through networks of underground pipes and tunnels, which caused the present dissected morphology through pipe/tunnel collapse and further evolution over a rather stable base level. Neglecting the importance of subsurface erosion phenomena in dispersive badland settings may lead to erroneous interpretations of the origin of these landscapes and considerable underestimation of the magnitude of erosion (Faulkner, 2013). In addition to the comprehensive approach to subsurface erosion in marine-sourced bedrock materials, silty colluvial or artificial fillings may also be very sensitive to piping, facilitating the expansion of secondary gullies at sidewalls or gully heads (Harvey, 1982; Benito et al., 1993). Steep hillslope gradients and low breakdown resistance of soft bedrock and/ or weathered regolith can induce a variety of mass-wasting and gravity processes affecting badland dynamics. Detached particles from desiccated regolith surfaces may accumulate in downslope sections by gravity fall covering rills and/or forming gravity talus or fans (Fig. 2.3; Clotet et al., 1988; Gallart et al., 2002). Regolith slow creeping is very frequent on badland hillslopes when swelling–shrinking alternations.

or prolonged wetting periods are active. For example, Schumm (1964) found in badland surfaces affected by intense cold-season bedrock weathering by gelivation a linear relationship between creeping rate and the sine of the topographic gradient. Furthermore, an array of small-to-large wet flows (e.g., mudflows, wet slides and slumps) initiated by regolith failure or liquefaction may have an active role in shaping many badlands (Howard, 2009). These mass-wasting processes, typically prevalent in badland areas under humid climate conditions, are also quite frequent in those

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Assistant Professor of Geography, Raja N.L.Khan Women's College (Autonomous), Midnapore

badlands in drier areas (Calvo-Cases and Harvey, 1996; Godfrey, 1997; Desir and Marin, 2007). In fact, the lack of physical evidence for mass-wasting processes during field visits (e.g., microscar occurrence) does not mean that mass movements do not occur in these systems. Small-scale debris flows have been observed during rainfall experiments (Hodges and Bryan, 1982; Wijdenes and Ergenzinger, 1998) or when using time-lapse video recordings during natural events (Yamakoshi et al., 2009) without causing conspicuous postevent marks.

4. BADLAND-INITIATION PATTERNS AND TRIGGERING FACTORS

Three general badland initiation patterns can be distinguished. The first two patterns correspond to the expansion of hillslope gullies that can themselves be initiated at mid-slope sections, caused by within-slope conditions, or at the slope bottom, through a combination of within-slope and basal conditions (Harvey, 1992). The third pattern corresponds to the disruption of a nonchannelized hillslope by mass movements that open a bare soil or rock scar to weathering and water erosion. The analysis of these phenomena is particularly relevant in subhumid and humid badlands, where these landforms are relatively small and young, which means that their initiation can be physically examined. Conversely, in semiarid and arid areas, badlands are usually very extensive and relatively old, so that their initiation factors are frequently obscured by the action of other drivers that control the long-term evolution of these systems. Mid-slope gullies initiate within hillslopes as 'discontinuous gullies' where the local erosive power of runoff overcomes soil-vegetation resistance, without the need of any change in base-level conditions (Harvey, 1992; Faulkner, 2008). In other words, this may occur when current hillslope processes are out of the equilibrium between climate, hillslope morphology and vegetation conditions. Major climate oscillations associated with geomorphic regime shifts (e.g., from periglacial to fluvial-dominated processes) may provide natural reasons for this framework (Bryan et al., 1987; Harvey, 1992), although human-induced vegetation disturbances are frequently claimed (Le Roux and Sumner, 2012). Even if the conditions for the development of these gullies are established, their initiation is often attributed to some particular disturbances such as intense overland flow caused by an extreme precipitation event. In other cases, as detailed below, man-made structures such as drainage channels or agricultural terraces may cause the formation of hillslope gullies. These mid-slope–initiated gullies may develop on steep hillslopes without the concurrence of any fluvial base-level stress. For example, sidewalls of deep gullies frequently continue to evolve as small badlands after the stabilization of the gully bottom, independently of any role of the base-level conditions in their initiation (Martínez-Casasnovas et al., 2003; Howard, 2009). Similarly, many gullies and badland hillslopes develop micropediments or large depositional forms at their base as a result of contemporary disconnection with regional drainage (Schumm, 1964; Harvey, 2007; Faulkner, 2008; Desir and Marín, 2013). Differently, slope-base gullies are initiated by basal scour of the hillslope, through a combination of headwall recession and incision of the master network (Harvey, 1987). The upwards progression of the gullies usually occurs where the hillslope stability conditions are already weak, although deeply incised channels may also cause the retreat of lateral gullies on gentle hillslopes, alluvial terraces or pediments with colluvial deposits, particularly where local lithology facilitates the development of aggressive piping (Faulkner et al., 2000). The main cause of the development of this kind of slope-based gullies is the incision of the mainstream network. In consequence, not only climate fluctuations or land use changes may play a decisive role, but tectonically induced base-level lowering or other discontinuities in drainage network evolution are also decisive factors (Mather et al., 2002; Harvey et al., 2014). Finally, landsliding has seldom been invoked as an important gully- and badlandinitiating phenomenon. For example, Piccarreta et al. (2006) demonstrated that, in the Basilicata badland area of southern Italy, newly generated gullies are initiated by landsliding on steep clayey slopes. Their results suggest that the speed of the geomorphic processes on these relatively steep landslide scars hampers the development of protective vegetation cover. Similarly, Lin and Oguchi (2004) described the formation of a badland-like morphology triggered by a large deep-seated landslide in a very humid area of the southern Japanese Alps where annual precipitation reaches 2000mm. The landslide, assumed to be over 100 years old, is attributed to rock instabilities driven by the incision of the drainage net (Chigira and Kiho, 1994). The incidence of these three different badland initiation patterns is ultimately determined by multiple interactions of a series of terrain instability and triggering factors, among which are: local lithology, regional to local tectonic drivers, climate oscillations, extreme rainfall events and human actions. The role of these phenomena in gully and badland initiation is extensively discussed with examples in the following subsections.

4.1 BADLAND LITHOLOGY

Lithology, which largely controls bedrock susceptibility to both weathering and erosion processes, is a major badland development factor (Solè et al., 1992; Calvo-Cases and Harvey, 1996; Faulkner, 2013; Kasanin-Grubin, 2013; Moreno-de las Heras and Gallart, 2016). Badlands occur on soft rock and unconsolidated sediments, which frequently contain soluble, cementing minerals. Specific characteristics, such as particle sorting, mineralogy and physical and chemical properties, may play determining roles in badland-forming processes (Gallart et al., 2002).

Fine-grained sediments are common components of badland geological materials that, depending on silt/clay abundance and lamination patterns, are generically termed mudstone, claystone, marl or shale (Campbell, 1989; Howard, 2009; Kasanin-Grubin, 2013). Particle sorting is an important factor affecting the breakdown and erosion susceptibility of the geological materials. Well-packed sediments containing a wide range of particle sizes are very resistant to disintegration, whereas highly-sorted materials with narrower particle-size distributions are more susceptible to breakdown and, consequently, to badland development (Taylor and Smith, 1986). Accordingly, grain-sorting meta-analysis of 14 sites along the Mediterranean basin indicated that, in general, Mediterranean badlands have well-sorted particle-size distributions, frequently dominated by silt particles, with clay as the second most common fraction, and very poor sand content (Gallart et al., 2002). The mineralogy of bedrock constituents is also fundamental to badlandformation processes. A variety of highly to moderately soluble minerals can be found in badland materials, among others halite, gypsum, calcite and dolomite. Dissolution and crystallization of soluble materials in cracks, bioclasts or the bedrock matrix near the surface can actively favour rock fragmentation and disintegration (Calvo-Cases and Harvey, 1996; Cantón et al., 2001; Alonso-Sarria et al., 2011). Furthermore, clay mineralogy is highly relevant to bedrock weathering processes (Kasanin-Grubin, 2013; Chapter 3). The effect of clay minerals on the behaviour of badland materials is highly variable, depending on their swelling capacity. Strongly active swelling occurs on smectite clay (including montmorillonite and bentonite), which upon wetting increase in volume very considerably (Frenkel et al., 1978). The dramatic impact of smectite on badland bedrock dynamics is well illustrated by Benton et al. (2015), who indicated for badland materials of the Chadron Fm. (White River badlands, South Dakota, USA) that 'a fragment of rock placed in a glass of water will break down in a matter of minutes, literally melting in place as the smectite absorbs the water, expands, and sloughs off the side of the rock fragment'. In fact, bedrock weathering by the action of repeated wetting–drying or freeze–thaw cycles is particularly intense on badland materials containing smectite and results in the formation of deep and highly disaggregated 'popcorn' regoliths (Solé et al., 1992; Regüés et al., 1995; Pardini et al., 1996; Kasanin-Grubin and Bryan, 2007). Differently, nonswelling badland materials lacking smectite clays resist bedrock weathering better and typically form thinner regolith mantles characterized by the presence of crusts (Kasanin-Grubin, 2013). Beyond local weathering processes, clay mineralogy may have wider implications affecting the distribution and detailed morphology of badlands. A good example illustrating these broad lithological effects can be found in the Pyrenean badlands of the upper Llobregat basin, NE Spain (Moreno-de las Heras and Gallart, 2016). These montane badlands are developed over two different lithologies differing in swelling capacity and clay mineralogy: swelling (smectite-rich) Garumnian mudstones of late Cretaceous continental origin (Fig. 2.4A) and nonswelling (smectitelacking) Eocene marls of marine origin (Fig. 2.4B). The high bedrock-weathering susceptibility of the swelling mudstones explains the lack of slope-aspect anisotropy of badlands developed over the Garumnian materials. Conversely, higher weathering resistance of the nonswelling Eocene marls precludes badland formation over these marly materials on sunny hillslopes, where cold-season gelivation is less active. Furthermore, the presence of smectite strongly influences the geotechnical stability of the Garumnian mudstones, which show lower friction angles and shear strength than the Eocene marls (Fig. 2.4C), with important implications for badland morphology. In fact, badlands developed on the more stable Eocene marls typically show rough topography characterized by very high slope gradients (30–40 degrees, Fig. 2.4B), whereas badlands show gentler slopes (20–30 degrees, Fig. 2.4A) on the more unstable, smectite-rich Garumnian mudstones. Similar badland morphology variations have been found in the vicinity of the Henry Mountains and White River regions of North America, where badlands developed on the smectite-rich materials of the Morrison and Chadron Fms. typically show rounder forms with lower mean average slope gradients than badlands developed on the smectite-poor materials of the Mancos and Brule Fms. (Schumm, 1956; Howard, 2009; Benton et al., 2015). The presence of high salt content is a common feature of badland geological materials, particularly for those of marine origin. High badland bedrock and/or regolith salinity constrain considerably the chances for seed germination and plant establishment, particularly in drylands where climate conditions (i.e., limited rainfall and high evaporation) strongly limit salt leaching from surface badland materials (García-Fayos et al., 2000; Maccherini et al., 2011; Gallart et al., 2013a). These striking effects of bedrock/regolith salinity on vegetation dynamics may justify, at least in part, the absence of vegetation on several semiarid badland sites (e.g., the extensive Bardenas Reales badlands in the central Ebro basin, NE Spain; Fig. 2.3). Salinity, and particularly the presence of exchangeable sodium in badland materials, has a fundamental relevance for the occurrence of clay dispersion and subsurface erosion phenomena (Benito et al., 1993; Gutierrez et al., 1997; Faulkner, 2013; Chapter 6). High sodium content saturating part of the exchange complex of reactive double-layer clays causes rapid clay swelling and dispersion on wetting, rendering them highly erodible. In clay-rich materials, dispersion can reduce considerable water infiltration, which may prevent extensive pipe enlargement but induce rill development (Bouma, 2006). For materials with moderate or relatively low clay content, however, clay dispersion does not alter infiltration and rapid enlargement of pipes through preferential water flow patterns may occur (Faulkner et al., 2000; Faulkner, 2013). Subsurface erosion in badlands developed on dispersive geological materials can result in the formation of near-surface micropipes, widespread pipe networks, and both gully and channel head cut extension by macropipe collapse or tunnel expansion (López-Bermúdez and Romero-Díaz, 1989; Torri and Bryan, 1997; AlonsoSarria et al., 2011; Faulkner, 2013). Buffering by either gypsum or organic acids released by microalgal crusts mobilizes sodium to lower layers of the materials. This process, frequently accompanied by the relocation of dispersed clays down the regolith/rock profiles, reduces the dispersive potential of the surface materials, thus causing geochemical stabilization of the site (Alexander et al., 1994; Faulkner et al., 2000). Dispersion vulnerability is generally measured by either the exchangeable sodium percentage (ESP, proportion of exchangeable sodium in relation to the cation exchange capacity) or the closely correlated sodium adsorption ratio (SAR1), which can easily be determined from the analysis of soluble (sodium, calcium and magnesium) cation concentrations extracted from soil, regolith or rock saturated paste. Soil and rock materials with ESP or SAR over 13 are susceptible to chemical dispersion. Piping susceptibility diagnosis in badland areas is improved by use of the relationship between the electrical conductivity and SAR of the geological materials, whereas the analysis of SAR/grain size and pH/SAR relations at different depths provides important clues for the study of geochemical stabilization processes (Faulkner et al., 2000). A good example of their application is provided in Piccarreta et al. (2006), who investigated the role of physico-chemical material properties and subsurface erosion on the development of sharp-edge calanchi and small, rounded-edged biancane badlands in southern Italy. They concluded that the dome-shaped biancane landforms in this area can be interpreted as geochemically stabilized end-products of the long-term erosive dynamics of calanchi landforms, which provides geophysical support for classical theories describing badland morphological diversity for this region as an evolutionary sequence of erosive landforms (e.g., Alexander, 1980; del Prete et al., 1994).

4.2 TECTONIC FACTORS

On the broad, regional scale, tectonic factors are fundamental to badland formation (Harvey, 1987; Grove and Rackham, 2001; Wainwright and Brazier, 2011). Tectonics, through terrain uplift, results in exposing to the atmosphere soft and erodible sedimentary materials at gradients that they cannot withstand, favouring rapid erosion and badland formation. In addition, tectonic uplift can impose regional baselevel changes that, if resulting in incision, may propagate up through channel and river networks, ultimately inducing the initiation and/or expansion of large gully and badland systems on susceptible, soft lithology. In fact, badlands typically occur where weak sedimentary settings lie in regions affected by significant past and/or present uplift, very frequently on or near mountain ranges. Paradigmatic examples of tectonic influence on badland initiation can be found in the regions of Basilicata (southern Italy) and Almeria (southeastern Spain), where regional landform dynamics have been largely modulated by highly active tectonics since the Pliocene. The Basilicata badlands are located at the southern foothills of the Apennines (Fig. 2.1A and D), in a region that is strongly influenced by the history of geological interactions between the European and African plates. Calanchi and biancane badlands occupy most of the hilly areas in this region, representing about one-third of the territory. Lithologically, these badlands are developed on thick (totalling up to 900m) series of marine-sourced clays and silty/sandy clays of Plio-Pleistocene age from the 'Argille grigio-azzurre' and Subalpine clay formations of the Sant'Arcangelo basin and Bradanic trough (del Prete et al., 1997; Piccarreta et al., 2006). Middleto-late Pleistocene regional tectonics uplifted these marine sediments at a rate of 0.5–0.9mmyr−1, bringing the Sinni, Agri, Cavone, Basento and Bradano rivers to downcut deep valleys perpendicular to the Taranto Gulf coast, with gully and badland forms rapidly progressing headwaters from the river networks (del Prete et al., 1994; Grove and Rackham, 2001; Bentivenga and Piccarreta, 2016). Badland initiation in southeastern Spain, and particularly in the Almeria region, is also linked to highly active tectonics (Wise et al., 1982; Harvey, 1987; CalvoCases et al., 1991; Alonso-Sarria et al., 2011). The Almeria region (Fig. 2.5A) comprises a series of mountain ranges and in-between sedimentary basins (Harvey et al., 2014). Geologically, the mountain blocks consist of metamorphic materials (of the Nevado-Filábride series and the Alpujárride nappe complex, structured in the Cenozoic by collision of the African and European plates) and, south of the Carboneras fault complex, volcanic rocks of Neogene origin. The intervening basins contain Neogene sedimentary rocks of marine origin, later enriched with continental sedimentary series. Badlands in this region are particularly well represented in the Tabernas, Sorbas and Vera basins, which have sustained significant tectonic activity since the Pliocene. Pliocene to recent average uplift rates in the area have shown the highest values (in excess of 0.17mmyr−1) for the mountain blocks, decreasing to 0.12–0.07mmyr−1 for the Tabernas and Sorbas basins and to 0.02–0.01mmyr−1 for the Vera basin (Mather et al., 2002; Braga et al., 2003; Harvey et al., 2014). High uplift rates and steep tectonically induced gradients have resulted in the formation of the very extensive Tabernas badlands (Fig. 2.1A and C), which show large vertical incisions (up to 200m) on deep-water, late Miocene (Tortonian) mudstones. These spectacular erosive landforms alternate in space along the Tabernas basin with deeply incised channels, remains of old pediments and more stable hillslopes resulting from sequences of dissection and stability phases (Alexander et al., 2008; CalvoCases et al., 2014). Badlands in the Sorbas and Vera basins extend over late Miocene (Tortonian/Messinian) gypsiferous marls and Plio-Pleistocene deposits of sands and silts (Harvey, 1987; Calvo-Cases and Harvey, 1996; Faulkner et al., 2000). Sustained Quaternary uplift in the Almeria region has induced a variety of river capture sequences (main capture sites are shown in Fig. 2.5A), with important consequences for both drainage network evolution and landform development (Harvey, 1987; Harvey et al., 2014). Probably the most dramatic of these sequences, in terms of badland development, took place during the late Pleistocene (about 100ka) in the Sorbas basin (Harvey and Wells, 1987; Harvey et al., 2014). The lower Aguas river, a stream draining towards the eastern coast, incised Messinian marly materials to behead the upper section of the Aguas/Feos drainage system (Fig. 2.5B). As a result, about 70% of the original Sorbas drainage (initially routed southwards through the Rambla de los Feos) was diverted towards the adjacent Vera basin in the east (Mather, 2000). Other consequences of this capture event were a sharp 90-m base-level reduction in master drainage at the capture site and the associated upstream propagation of a wave of incision (Mather et al., 2002; Stokes et al., 2002). Where the wave of channel incision coincided with the presence of erodible marly and silty materials, severe erosion caused the formation of badlands. In particular, rapid base-level change produced badlands near the capture site almost immediately (Fig. 2.5C), whereas in a more distant zone (Rambla de Mocatán, Fig. 2.5B and D) badland development was not initiated until around 80 ka after the capture event (Mather et al., 2002). As well as the large influence of regional tectonics in badland formation, landscape tectonic structure may also influence badland-shaping processes at shorter, local scales. In fact, badland geological materials very frequently contain considerable networks of joints and fissures derived from direct tectonic action (e.g., faults, compression, deformation stress) or from the combined effects of terrain uplift and erosion, causing pressure-release unloading of bedrock materials (Campbell, 1989; Gallart et al., 2002; Wainwright and Brazier, 2011). These fissures and joints affect bedrock strength and also provide preferential pathways for the entry and interaction of atmospheric fluids within the matrix of badland materials, which spatially conditions weathering processes (Solé et al., 1992; Cantón et al., 2001), the activity and distribution of subsurface pipes and tunnels (Torri and Bryan, 1997; Calzolari and Ungaro, 1998; Alonso-Sarria et al., 2011) and the nature and direction of general surface drainage (del Prete et al., 1994; Piccarretta et al., 2006).

4.3 CHANGES AND FLUCTUATIONS IN CLIMATE CONDITIONS

Naturally sculptured badlands have been traditionally seen as characteristic forms of semiarid regions, whereas in wetter climates badland development has been seen as more likely to be the

Dr. Pravat Kumar Shit
Assistant Professor of Geography, Raja N.L.Khan Women's College (Autonomous), Midnapore

result of disturbances of a fragile natural equilibrium by humans (Bryan and Yair, 1982). This conceptual framework is evoked in many publications, where the onset or reactivation of badlands and gully incision is merely attributed to dry drift climate oscillations resulting in reductions of vegetation cover, without leading, in general, to any further discussion of the geomorphic processes that underlie these changes. Indeed, general Quaternary landform evolution explorations for semiarid regions ascribe the phases of pediment and fluvial terrace formation to wet 'pluvial' periods, whereas the phases of drainage incision are attributed to drier periods, regardless of sea level variations (Tricart, 1969; Ballais, 1982; Howard, 1997). Two significant contributions, however, provide physically based evidence on how an increase in climate aridity may compromise badland activity for already dry regions. Yair et al. (2013) indicate that the high infiltration rates of colluvial regolith deposits on downslope badland positions in desert-type systems can typically overcome arid precipitation rates to the extent that the current frequency of effective events does not justify badland development. In addition, Bryan et al. (1987) suggested that aridphase deposition of highly permeable aeolian loess over badland topography may also contribute to stabilizing these landforms. More explicit associations between climate shifts and badland initiation are attributed to periods of base-level change and drainage network incision, with the assumption that drainage incision is an indirect sign of badland growth in areas where soft bedrock lithology and current climate conditions facilitate badland development (e.g., Díaz-Hernandez and Juliá, 2006; Piccarreta et al., 2011; Tooth et al., 2013). As discussed previously, some relief vigour is a necessary but not sufficient cause for badland development, because these landforms may also develop on hillslopes without invoking any base-level change. In addition, it is also frequently assumed that the periods of incision of the main channel network are synchronous with the incision of elementary badland drainage pathways and hillslope gullies, but the increased sediment production caused by badland erosion may also cause aggradation in the mainstream channels (Wells and Gutierrez, 1982; Bryan et al., 1987). The periods of channel incision in the low orders of the drainage networks have been generally related to climate phases with fewer low-intensity rains and more intense storms (Leopold, 1976; Piccarreta et al., 2011). Nogueras et al. (2000) showed that the effects of increased frequency and duration of dry periods are more important than the intensity of individual storms because healthy grass carpets at the bottom of temporary streams provide good protection against linear erosion, whereas aggradation of the valley bottoms was compatible with erosion in the badland interfluves by high-intensity rainfall. Howard (2009) showed that a short-term reduction of vegetation cover by, for example, a dry-spell disturbance could trigger accelerated erosion in valley bottoms as well as on hillslopes, which may continue much longer than the duration of the initial disturbance. This author also suggested that hillslope gullies and channel incisions may form and recover without the action of any external forcing, either spatially distributed as discontinuous gullies are (Leopold and Miller, 1956) or analogous to the colluvial infilling and excavation cycles described by Dietrich and Dunne (1978). Faulkner (2008) also insisted on the possibility that intrinsic changes of drainage network connectivity may cause large changes in the operation of badland development processes, even in the absence of large-scale climate variations. Moving beyond the study of badland development and stabilization processes, some authors have paid particular attention to the analysis of likely change patterns in badland evolution in the context of regional climate variations. Calvo-Cases and Harvey (1996), comparing badland processes in diverse areas in SE Spain, claimed that a regional increase in precipitation may simplify the set of badland-shaping processes, ultimately reinforcing the relevance of rilling to badland evolution, whereas a decrease in precipitation would strengthen the role of cracking and desiccationrelated processes, probably increasing the complexity of process interactions involving mass movements. Yair et al. (2013)

speculated for the Zin Valley badlands of Israel that, if climate were to change, these fairly stable landforms would respond in a nonlinear way to regional increases in precipitation. In fact, these desert systems show a very important resilience to climate change, but sharp increases of precipitation over a critical regional threshold (at ∼300mmyr−1 annual rainfall) could induce the erosion of the downslope colluvial deposits that at present disconnect runoff within the badland systems, increasing channel incision in the master channel network and further rejuvenating the stabilized landforms.

4.4 EXTREME RAINFALL EVENTS

Cause–effect relationships between the occurrence of extreme rainfall events and both gully and badland initiation are not infrequently evoked for subhumid and humid areas, although the connection of these processes is usually difficult to verify in the field (Grove and Rackham, 2001; Vanwalleghem et al., 2005; Chiverrell et al., 2007). Harvey (1992) suggested that large gullies in the Howgill Fells (northwestern England) were the result of the coalescence of discontinuous gullies induced by vegetation changes due to human activity, whereas small gullies were triggered by basal stream scour, particularly during ∼2-year return period floods. In a humid temperate area of Australia, Prosser and Soufi (1998) suggested that moderate rainfall events of 80–100mmday−1 with a recurrence interval of 1.4–2years can initiate gully erosion in hillslopes recently cleared of native forest, whereas a rainfall event of 200mm in a day with a recurrence of 10–15years induced massive gully development in all spatial conditions.

In montane areas, extreme rainfall events very frequently trigger shallow soil slips that may give rise to debris flow and gully initiation. For example, Rice and Foggin (1971) described a widespread development of slips in the San Dimas Experimental Forest (California, USA) as a consequence of two extreme rainy periods in 1966 and 1969. These authors stated that the final result of most of these slips was the formation of large gullies. Further inspection of 1994 aerial photographs, however, indicates that these scars have disappeared from the landscape due to vegetation and/ or soil recovery. Similar long-term natural recovery of flood-originated gullies can be interpreted for the effects of a 100-year recurrence event that occurred in 1949 in the Central Appalachians (USA). Hack and Goodlett (1960) reported nearly 100 shallow soil slips that produced large chutes in the hillslopes. Most of these scars still appeared remarkably fresh as incised gullies 6years after the flood event, although 1997 aerial photographs demonstrate that these incisions have, for now, recovered naturally. In spite of the important strength of the natural colonization processes that characterize the, in general, highly resilient subhumid and humid landscapes, soil slips and large landslide scars generated by extreme floods may in some cases persist or progress in time in the form of highly active erosive landforms. In 1988, the highly humid (∼2400mmyr−1 precipitation) and tectonically active East Coast region of New Zealand's North Island received an event of 535mm of rain in 2 days that caused large generalized scars on steep hillslopes covered by indigenous forests on mudstone lithology (Parkner et al., 2007). Inventories made by these authors 9years after the flood event found initial stages of natural colonization by vegetation on many scars. Exploration of 2016 aerial photographs for this area, however, shows that most of the scars are still active or have progressed in the form of mass-wasting–dominated badlands, while some others have clearly recovered vegetation (Fig. 2.6A). Similarly, in November 1982, an extreme flood event occurred in the Catalan Pyrenees (NE Spain), accounting for up to 560mm rainfall in 2 days, which caused over 100 shallow soil slips in a 1500 km2 humid region, with a marked slip preference for deforested grasslands (Gallart and Clotet, 1988). Although the evolution of these slips has not yet been thoroughly inspected, a rapid exploration in this mountain region indicates no currently conspicuous active scars derived from the 1982 flood, except for two cases developed over smectite-rich mudstones near Vallcebre village (Fig. 2.6B). Ten years after the flood, these two scars showed clear evidence of intensively active surface erosion (33mmyr−1 denudation; Llorens et al., 1997). At present, they continue expanding slowly as small badlands, which is probably facilitated by the montane (about 1500m altitude) harsh cold-season conditions of the site and the high weathering/erosion susceptibility of the local mudstone lithology (Moreno-de las Heras and Gallart, 2016). Overall, these four landscape evolution cases distinguish two contrasting landform responses after extreme flood disturbance. The long-term evanescent nature of the flood scars in the aforementioned San Dimas and Appalachian examples matches the classical cycles of revegetation and prolonged colluvial infilling of hollows following rapid excavation by mass movements during extreme events proposed by Dietrich and Dunne (1978). However, the flood scars for the New Zealand and Pyrenean examples do not evolve towards vegetation and soil recovery but follow an alternative path to intensified erosion (Gallart et al., 2002), ultimately controlled by a variety of regional to local physiographic conditions favourable to badland development (e.g., high relief, soft bedrock, active tectonics, harsh winter thermal conditions).

4.5 HUMAN IMPACT

Human activities can induce drastic alterations in the geomorphological stability and structure of landscapes, resulting in land degradation, very frequently accompanied by the onset of intense erosion (Hooke et al., 2012). In the most extreme cases, human-induced degradation may give rise to the formation of large gully systems, badland extension or the initiation of badland and badland-like landscapes. However, landform evolution modelling by de Ploey (1992) in the Mediterranean basin – where humans have played an active role transforming landscapes for several millennia – concluded that badlands in this region are in general quite old, with a minimum age between 2,700 and 40,000years. These results indicate a full Holocene or late Pleistocene age for, at least, the most extensive Mediterranean badland areas, suggesting a natural origin that opposes alternative scenarios of massive badland initiation in the region caused by forest clearance and soil degradation during and after Roman times (Grove and Rackham, 2001). More likely, historical-to-present human activities, probably combined with other factors, have contributed to the rejuvenation of naturally initiated, extensive badland systems or the initiation of more local badland and badland-like landscapes. A classic case of human influence on badland processes is offered by the humid badland and gully systems developed on blue and black marls of late Cretaceous and Jurassic age throughout the southern French Alps. Different gullying phases have excavated these erosive systems (Descroix and Gautier, 2002). The oldest phase dates back over 10,000 years and probably originates in post-glacial climate changes. The latest gullying phase took place in historical times and has been associated with deforestation and regional agro-pastoral expansion pulses during and after the Little Ice Age (Ballais, 1997). Climate fluctuations during this historical period may have also modulated gully incision (Descroix and Gautier, 2002). This fairly recent incision phase had a determining influence on rejuvenating and/or generating new gullies in the region, and particularly for the renowned terres noires badlands of Draix (Fig. 2.7A). Similar historical phases of demographic pressure, combined with parallel changes in climate conditions, have been pointed out to have contributed to the formation of new gullies and/or rejuvenation of badland systems in the tectonically active Italian regions of Tuscany and Bassilicata (Calzolari and Ungaro, 1998; Torri et al., 2000; Piccarreta et al., 2011). Dryland agricultural practices on erodible silty or marly materials may exacerbate both gullying and pipe-formation processes, favouring the initiation or expansion of

Dr. Pravat Kumar Shit
Assistant Professor of Geography, Raja N.L.Khan Women's College (Autonomous), Midnapore

highly erosive systems. For example, Gallart (1992) describes widespread phenomena of rapid gully incision and pipe development in the dryland agricultural region of the Conca d'Odena (inner Catalonia, NE Spain) as a consequence of surface levelling and runoff concentration into ditches, which resulted in the expansion of deep gully systems carved by the Anoia River on weak grey marls of Eocene age (Fig. 2.7B and C). Similarly, the abandonment of traditional agricultural structures (e.g., drainage channels, irrigation ditches, terraces) on vulnerable lithology may locally intensify surface and subsurface erosion. Traditional agricultural terraces, for instance, may have disastrous consequences on highly dispersive (sodic) materials, where aggressive piping and further gully development by tunnel collapse can lead locally to the genesis of badland-like landscapes (Fig. 2.7D). Piping may precede terrace abandonment in these farming systems, where the presence of reworked soils of dispersive nature and high hydraulic gradients at the terrace risers may exacerbate subsurface erosion (Faulkner et al., 2003; García-Ruiz, 2011). These extreme effects of piping on terraced landscapes have locally favoured the rejuvenation and/or extension of several badland areas developed on dispersive marl and mudstone materials in SE Spain (Mather et al., 2002; Romero Diaz et al., 2007; Alonso-Sarria et al., 2011; Calvo-Cases et al., 2011). Human activities involving large earth movements are major vectors of landscape change. Landscape scars generated by old quarries can evolve into highly erosive gullied landforms. For example, combined application of detailed geomorphological analysis, tree-ring examination and exploration of historical socio-economic and geographical archives in the region of Segovia (central Spain) revealed the formation of an array of local badland sites in areas affected by traditional quarrying of fine silica sand on poorly compacted sand and sandstone formations during and after the 18th century (e.g., the Barranca de los Pinos badlands; Fig. 2.7E), which at present show highly active gully retreat (Díez-Herrero and Martín-Duque, 2005; BallesterosCánovas et al., 2017). More impressively, the lack or failure of land reclamation on artificial surfaces generated by contemporary, industrial surface mining (e.g., large tailings, waste dumps and spoil banks) can lead to the generation of badland-like landscapes characterized by the rapid development of very dynamic gully systems (Moreno-de las Heras et al., 2011; Martín-Duque et al., 2015). A disastrous case of mining reclamation failure is well illustrated by Martín-Moreno et al. (2018), who describe the rapid formation of badland-like landforms on highly unstable sidehill dumps by severe gullying – accounting for over 350 tha^{-1} yr^{-1} erosion – in a mountain kaolin mine of the Alto Tajo region in central-eastern Spain (Fig. 2.7F). At the broader, regional scale, intense erosion and landform instability at this location is blamed for significant changes in the sedimentary structure of rivers. In fact, these catastrophic landform disorders, though local in extension, can induce large geomorphological impacts of comparable intensity to those caused by tectonics and climate fluctuations.