Chapter 22

Hillslopes as Evidence of Climatic Change

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Introduction

Geomorphic systems disclose great differences in their sensitivity to climatic change, and the various relief units carry the imprints of past processes to dissimilar degrees. Fluvial systems are highly susceptible to climatically induced changes in process. Hillslopes, on the other hand, are generally regarded as being rather resistant to such changes. In addition to the sensitivity of relief units to climatic change, another point of crucial geomorphic interest is how long the legacies of past processes are preserved in the form elements. Unfortunately, sensitivity to change and the length of time of preservation of past changes are usually inversely correlated. This means that we have either a detailed record of short-term changes for a limited period of time or a relatively inaccurate record of only major changes for a longer timespan. Where a detailed sedimentary record of past climatic changes has survived, related landform records may not be complete, because climatic changes are generally more frequent than landform changes.

The critical factors rendering a process change geomorphologically significant are the intensity and magnitude of the climatic change as well as its direction and the duration of the new climatic regime. These factors are counteracted by the resistance of the geomorphic system to external change. It is this resistance to change which predominantly controls the relaxation time of a system. Relief units with very short relaxation times will only document legacies of former processes for the very recent past, and relief units with very long relaxation times will only document long-lasting and very intensive changes.

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Resistance to Change

With special regard to hillslopes the resistance of a relief unit to process change is dependent on a number of internal system variables relating to form, process, and lithological characteristics (Littmann and Schmidt 1989, Selby 1993).

(a) Slope angle: A high slope angle leads to a low resistance to gravitational and water erosion on the slope units compared with pediment and planation surfaces or lithologically controlled stripped plateau surfaces. On free faces no legacies of past processes are preserved.
(b) Lines of process concentration: If the relief unit is affected by lines of process concentration (e.g. valley floors, erosional gullies, slope rills) which channel water and material throughput, it becomes more susceptible to change. If a scarp is breached by a major river or if a river runs at its foot (Fig. 22.1), resistance to change is reduced by the closer connection to base level fluctuations and flood effects.
(c) Lithological characteristics of the surface material and substrate: This type of control is especially important in arid regions, where weathering and erosion processes operate in a highly selective manner in contrast to more humid regions which have greater moisture availability and an increase of chemical weathering. Outcrops of resistant

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Fig. 22.1 Valley-side scarp in the Canyonlands section of the Colorado Plateau, Utah. The Colorado River and a tributary flow at the base of the scarp. The caprock consists of the sandstones of the Glen Canyon Group, and the lower slope is formed in the redbeds of the Chinle Formation (slope type 3). The lower slope is directly affected by fluvial processes and base level fluctuations. No legacies of past slope processes have been preserved

bedrock layers on slopes, hillslope surfaces consisting of consolidated talus or gravels, and old pediment surfaces show a high resistance to erosion, whereas unconsolidated hillslope debris, particularly when consisting of fine-grained colluvial deposits, is easily rearranged by changing process regimes.

Taken overall, the effectiveness of a past geomorphic process E associated with a past climate in shaping a hillslope is proportional to the product of the duration of the former process T_1 and its intensity I_1 divided by the resistance of the hillslope unit R.

$$E \propto \frac{T_1 I_1}{R} \tag{22.1}$$

On the other hand, the degree of preservation P of a past process on a hillslope is dependent on its resistance to change R divided by the duration T_2 and intensity I_2 of the present process.

$$P \propto \frac{R}{T_2 I_2} \tag{22.2}$$

If the resistance of a hillslope unit is considered to remain constant over time, the response of the system is mainly controlled by the duration and intensity of the processes involved. Only if the product of the duration and intensity of a past process is greater than the product of the duration and intensity of subsequent processes will the effects of the past process remain visible on the landform. With this general background, we will examine the suitability of desert hillslopes for the elucidation of past climatic changes.

Indicator Potential of Hillslopes

The form of present-day slopes has evolved mainly over the past million years or less and the climatic changes over this period have been on a scale sufficient to have had a considerable effect on surface processes, particularly in the subtropical desert margins, which are most susceptible to environmental changes and tend to states of instability (Schmidt and Meitz 1996). The ages of hillslopes strongly depend on the strength of the material of which they are composed and the related denudational activity (Parsons 1988, p. 121). Consequently, the ages of hillslopes are decisively controlled by their resistance to change (see Equations 22.1 and 22.2). When hillslopes or parts of them are old enough to have experienced the climatic changes of the Pleistocene, it is necessary to investigate

(a) whether the climatic changes were strong and long enough to shift form–process relationships from one state of equilibrium to a significantly different one (a change in principle), or
(b) whether they only enhanced or attenuated existing form-process relationships (a change in degree), or
(c) whether they were too weak and short-lived to have caused any effect on the general trend of hillslope development (no morphological change).

Only in the first case will truly independent climatogenetic landform elements in Büdel's (1982) sense have come into existence.

Slope systems are controlled by lateral backwearing, sometimes in combination with downwearing and slope angle reduction where processes of slope decline are involved. On bipartite cuesta scarps parallel retreat is the dominant mechanism (first described by Powell 1875). Slope denudation implies an inherent mechanism of surface destruction along the entire slope reach. This surface-destruction mechanism is a major drawback to the preservation of forms created by past climates.

The ability of hillslopes to bear clear testimony to distinct climatic phases is complicated by the fact that process change may not simply be expressed in a characteristic form assemblage of a slope system. The work of varying processes can also result in a slope configuration, which represents a typical expression of a particular sequence of different climatic periods. The morphology of these slopes is due neither to present-day nor to past processes alone but has been shaped by a temporal succession of processes. Descriptions of hillslopes characterized by this type of development have already been presented by Oberlander (1972) for the Mojave Desert and by Moss (1977) for southern Arizona. The term ‘morphogenetic sequence’ has been proposed by Mensching (1974) for the evolution of landforms, which owe their present shape to a combination of different processes. He described a number of examples from North African arid and semi-arid areas.

As will be shown below, most legacies of former slope processes are found on bipartite slopes, where a hard caprock overlies an easily erodible formation. These hillslopes are generally called cuesta scarps. Homolithic slopes do not tend to furnish similar information. The indicator potential of the bipartite slopes can be explained by their specific composition. They normally consist of an upper steep slope in the resistant caprock and a lower, moderately inclined slope in less resistant beds (see Fig. 22.5). There has been much confusion in the terminology of the different form elements of cuesta scarps (cf. Oberlander 1997). Terms such as subtalus slope (Koons 1955), footslope (Oberlander 1977), rampart (Howard and Kochel 1988) or substrate ramp (Oberlander 1997) have been utilized for the part of the scarp developed in soft rocks below the caprock. The more neutral term lower scarp slope (Schmidt 1987a, Schmidt and Meitz 2000a,b) will be used in this chapter. The caprock provides resistant talus which protects portions of the lower scarp slope from the slope’s general fate of being destroyed by the backwearing processes. Mainly for this reason most studies dealing with evidence of past processes on desert hillslopes have chosen cuesta scarps as their research subject; as will this chapter. This type of scarp may serve as a useful indicator for major climatic fluctuations (Gerson 1982, Schmidt 1989b, 1996, Gutiérrez et al. 2006).

The resistance to change and consequently the inertia is greater for hillslope systems than for rivers, floodplains, and alluvial fans. Fluvial systems are able to store a detailed record, but this record also includes some ‘noisy’ information. For the Colorado Plateau in the south-western United States the Holocene and late Pleistocene depositional and erosional history of the river systems is well known from the alluvial record (e.g. Hack 1942, Euler et al. 1979, Wells et al. 1982, Graf et al. 1987, Hereford 1987). There is, however, no comparable detailed evidence of past climatic oscillations on cuesta scarps (Schmidt and Meitz 2000a,b). Apparently the relaxation times are longer and thresholds for formative change are higher for slopes. The base-level fluctuations were neither long enough nor strong enough to be propagated upslope (Schmidt 1988, Littmann and Schmidt 1989). On the hillslopes only former climates of long duration and characterized by highly effective processes have left their imprint (see Equation 22.1). This makes these hillslopes a valuable indicator for the low frequency climatic changes of the Pleistocene. But not all desert cuesta scarps bear the imprint of former processes. A strong influence is exerted by the lithological and structural attributes of the caprock and substrate and other controlling factors (Gerson and Grossman 1987, Schmidt 1989b, Schmidt and Meitz 1996). These factors will be discussed in a subsequent section.

Stability or Activity in Dry Periods

The question whether desert landforms remain more stable in dry (interpluvial) than in more humid (pluvial) conditions has for long been a matter of controversy. For desert washes (arroyos) in the American Southwest the discussion about the influence of different hydrologic regimes including human interference has been summarized by Cooke and Reeves (1976), Graf (1983) and Karlstrom (1988). Conclusions on this important subject remain very mixed. For cuesta scarps in the same region this controversy has initially been highlighted in the papers of Ahnert (1960) and Schumm and Chorley (1966). Ahnert (1960) argued that erosional processes, particularly slump block production and sapping processes supported by subterranean wash, were much more active during the pluvials, and that present-day landform modifications are only of minor importance. He derived a model of four main scarp profiles from the rock sequence in Monument Valley, which were formed by varying combinations of pluvial and interpluvial processes (Ahnert 1960). The greater moisture availability in the humid periods may well have enhanced groundwater flow and seepage above impervious beds. Howard and Kochel (1988) and Laity and Malin (1985) concluded from their detailed field observations in Navajo Sandstone alcoves that present-day sapping activity and alcove development is much less than during a past pluvial climate. With respect to talus production, Schumm and Chorley (1966) maintained that the frequently observed lack of basal talus on the lower scarp slopes is not a consequence of slope stability but a result of effective weathering and a high rate of talus removal. Ahnert's (1960) climato-cyclic interpretation contrasts sharply with this equilibrium view of Schumm and Chorley (1966). Schmidt and Meitz (2000a,b) in their comprehensive study of scarp slope types on the Colorado Plateau conclude that changes in humidity may induce opposing tendencies of reactions in different slope types corresponding to their intrinsic lithological attributes. This means that there is no general trend of activation or stabilization with changing rainfall input (see below).

Presently inactive landslide material, slump blocks, and major rockfall accumulations on cuesta slopes have frequently been interpreted as indicators of decreasing geomorphological activity from the last humid period to the present dry phase. Figure 22.2 shows the distribution of mean annual precipitation on the central Colorado Plateau, the area to which most of the discussion in this section refers. The gravitational deposits have been regarded as relict landforms by many field geologists, as depositional features inherited from past pluvial processes. Symptomatic of this kind of interpretation with regard to landslides is Hunt et al. (1953, p. 171) statement: 'They appear to have survived from a more humid climate when conditions were more conducive for weathering and recession of cliffs.' Landslides and large-scale rotational slumps as evidence of humid phase activity have been described from the Vermilion Cliffs and Echo Cliffs, Arizona (Strahler 1940, Phoenix 1963), from the Red House Cliffs, Utah (Mullens 1960), and the White Canyon area, Utah (Thaden et al. 1964). All of these sites lie in areas with a present-day preciptation of about 200 mm and are characterized by massive sandstones (Glen Canyon Group), overlying swelling bentonitic Chinle shale. These landslides are enormous in size, being up to 2 km long and 800 m wide (Thaden et al. 1964, p. 75). The absence of youthful slumps is taken as an indicator of present stability. With similar arguments Reiche (1937) reported landslides from the Black Mesa area, Arizona, where Mesa Verde Sandstone overlies Mancos Shale. Reiche, however, also reported two major young slumping events (Toreva Blocks) from 1870 and 1927.

Schumm and Chorley (1966) collected eyewitness accounts of active scarp retreat processes from the National Parks and National Monuments Services on the Colorado Plateau. These accounts show that rockfalls of different magnitudes have frequently occurred in recent years. This conclusion is supported by descriptions of field geologists for different parts of the Colorado Plateau. Ford et al. (1976) gave a detailed account of rockfalls in the Grand Canyon area. Cliff retreat is also a presently active process in the Montezuma Canyon area, southeastern Utah, with isolated rockfalls occurring occasionally in the Dakota and Salt Wash sandstones (Huff and Lesure 1965). In the Slick Rock district, southwestern Colorado a number of slides were observed during fieldwork in the years 1953–1956. The slides involved the Burro Canyon formation and the Salt Wash sandstone and occurred after heavy autumn rain (Shawe et al. 1968). Apparently present-day intensive precipitation events are well capable of triggering mass movements.

Schumm and Chorley (1964) described a major rockfall event in Mesa Verde sandstones in the Chaco Canyon National Monument, New Mexico (Fig. 22.4), dating from 1941. They concluded (1966, p. 19): 'Although such contemporary occurrences of this magnitude may be rare, the fact that slumps have occurred during the past 100 years illustrates the need for caution in attributing all occurrences of an erosional phenomena [*sic*] to past climates greatly different from the present'. This quotation implies that the morphological features of cuesta scarps can be to some extent explained by presently active processes and that, if there have been changes at all, they may have been only in the relative rates of processes. The continuity of large-scale cliff retreat processes has also been stressed by Davidson (1967) in the Circle Cliffs area, southeastern Utah and by Shawe et al. (1968) in the Slick Rock district, Colorado.

Any conclusion concerning activity or stability of geomorphic processes must try to avoid some potential hazards of misinterpretation. Stability or activity is highly dependent on the altitudinal location and on

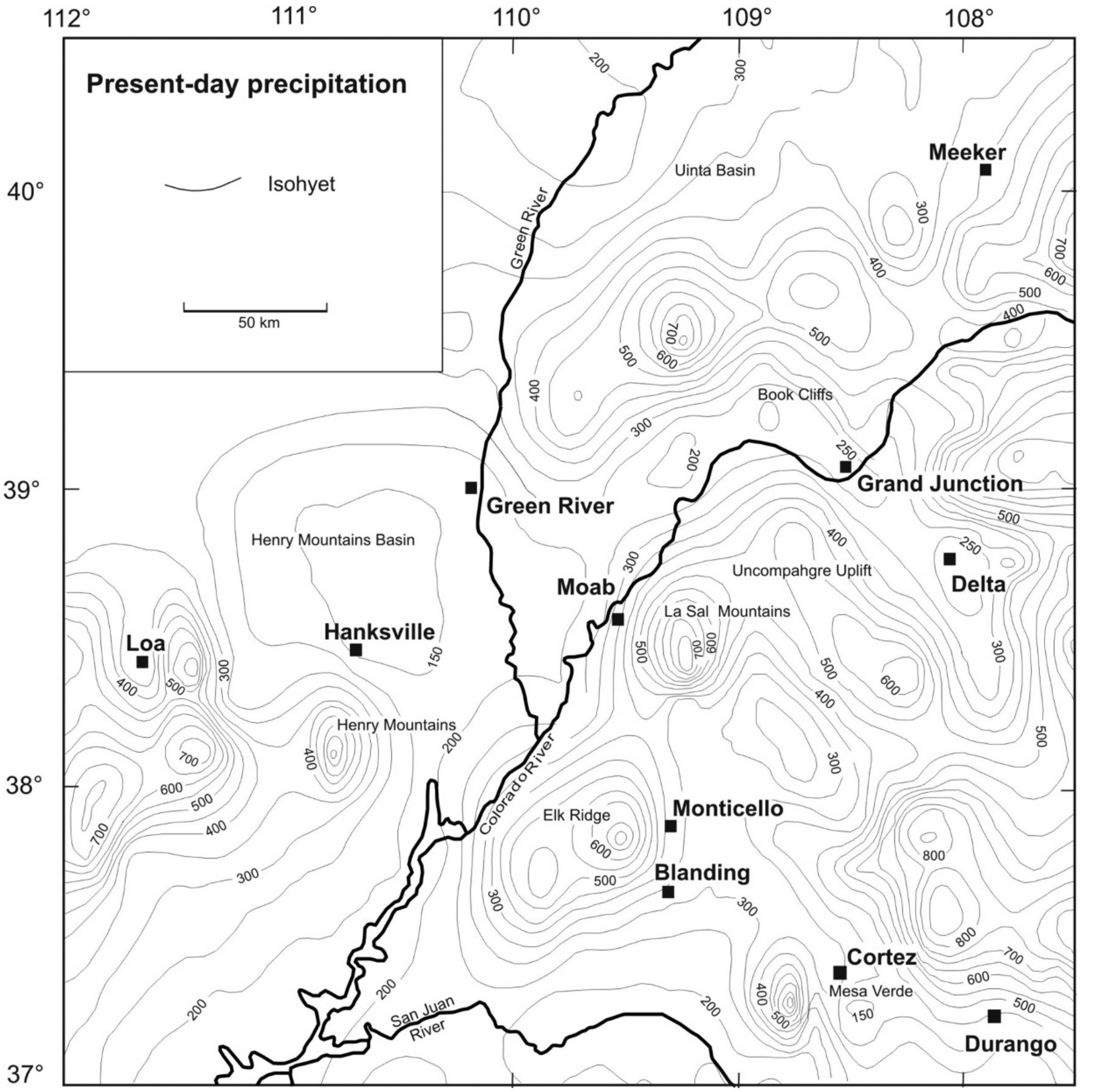

Fig. 22.2 Distribution of mean annual precipitation on the Colorado Plateau in southwestern Colorado and southeastern Utah. Some of the localities mentioned in the text are shown

the lithological compostion of the slope (Schmidt and Meitz 2000a,b). It is also extremely hazardous to judge from a one- or two-year field investigation or even from a 100-year photographic record (Baars 1971, Shoemaker and Stephens 1975) whether a specific type of landform in a desert environment is stable or not. A period of 100 years covers not more than 1% of the Holocene. Geomorphological events on a hillslope are not only highly sporadic in time but also unevenly distributed in space in contrast to fluvial action. Gravitational events may be ordered according to the magnitude of the masses of material involved. They range from the fall of small grains through the fall of individual stones, slabs, and blocks, to rockfalls, rockslides, slumps, and landslides. Masses ranging from 10^{-3} g through 10^{8} g to 10^{12} g are transported by the different processes. Obviously the authors who deduce present-day morphological stability from their observations refer to giant landslides with lengths of several hundred metres or large slump blocks with lengths of 50 metres or more (e.g. Reiche 1937, Strahler 1940, Mullens 1960). Gravitational processes of smaller magnitude such as falls of individual rock fragments, minor rockfalls of a few cubic metres, and slab failures along joints or exfoliation structures have also been observed by authors, who otherwise deny the present-day activity of cuesta scarps (Strahler 1940). 'After severe summer rains and during early spring the Vermilion and Echo Cliffs sometimes resound with noise of falling rocks. The blocks that fall are sometimes 15 or 20 feet in diameter' (Phoenix 1963, p. 40). The activity of processes has to be evaluated in the light of magnitude and frequency concepts. Moreover, on the scarp slopes the gravitational processes are spatially and temporally discontinuous. The variability in space and time increases with increasing process magnitudes. It is difficult to decide whether a particular type of high-magnitude process is really extinct or whether we just meet a cuesta slope in the interval between two such rare events depending on the specific recurrence intervals (Viles and Goudie 2003).

When there are no signs of present-day high-magnitude events (rockfalls, slump blocks, or landslides) on a particular cuesta scarp section, this does not mean that there is a general state of morphodynamic stability. So the cuesta scarp section under consideration may well experience events of smaller magnitude, while at other scarp sections events of greater magnitude may happen at the same time (Fig. 22.3a,b). Only the chronological integration of processes of different magnitude and frequency and the spatial integration of complete scarp sections will lead to a reliable evaluation of process activity in desert regions.

A space-integrating argument in support of the view of present-day morphological activity is the current high rate of suspended sediment transport of the Colorado River and its tributaries (Smith et al. 1960, Iorns et al. 1965, Schmidt 1985, Graf 1985). The overall denudation rate lies between 70 and 350 $m^3\,km^{-2}\,y^{-1}$ (or mm per 1000 years) (Schmidt 1985), but it is difficult to determine how much material is eroded from the slopes and how much originates from the alluvial valley fills which have been dissected since the end of the 19th century (Graf 1983). Erosion measurements on lower scarp slope rocks yield additional evidence for high process intensity. Colbert (1966) determined a denudation rate of 5.7 $mm\,y^{-1}$ for steeper parts of the lower scarp slope of the Echo Cliffs, northern Arizona, in the bentonitic shales of the Petrified Forest member of the Chinle Formation. Surface lowering in the easily erodible Mancos Shale at the foot of the Book Cliffs, southeastern Colorado, varies between 0.2 and 2.2 $mm\,y^{-1}$ and is a function of inclination and land use (Lusby 1979). In the same area a rate of denudation of 2.7 mm was measured in a single event (Hadley and Lusby 1967). Intensive lowering on Mancos Shale slopes (0.15–1.5 $mm\,y^{-1}$) was also determined by stake exposure measurements (Schumm 1964). The high erosion rates in the lower slope rocks are also indicative of the general erosional activity of the Mancos Shale slopes (cf. Fig. 22.8b).

In Northern Africa, too, different views have been expressed concerning activity or stability of gravitational processes on cuesta scarps. The cuesta scarps in Nubian Sandstone on the margins of the Mourzouk basin have been investigated by Barth and Blume (1975) and Grunert (1983). Grunert stated that the landslides on the western flank of the basin are 'fossil' landforms dating from more humid periods in the middle Quaternary with annual precipitation amounts in excess of 400 mm (presently about 10 mm)! The landslides are presently destroyed by fluvial erosion. Barth and Blume (1975) take a broader view, pointing out that there are cuesta sections with extremely intensive denudational processes and others without any recent activity. The variations are explained by non-climatic controlling factors such as differences in resistance between caprock and lower slope rocks, differences in the thickness ratio, and differences in the elevation above base level. Apparently these controlling factors influence the resistance to change of the cuesta scarps.

In a study of limestone scarps in the northwestern Sahara, Smith (1978) concluded that karstic and sapping processes and, consequently, cliff retreat are now stagnant. This view is again expressed in another case study on Cenomanian–Turonian limestone scarps in south-eastern Morocco (Smith 1987). He deduced that the landforms are undergoing superficial modifications by weathering, but that the landscape as a whole has been essentially stable since mid to late Quaternary times (Tensiftien: about 1.3×10^5 years BP). He contended that undermining of the caprock is largely the product of spring sapping. The present author has also investigated the limestone cuesta scarps in Southern Morocco (Schmidt 1987b, 1989a). Certainly karst processes and sapping by karst water were much more efficient during humid periods than today. But rockfalls are not only caused by spring sapping and related

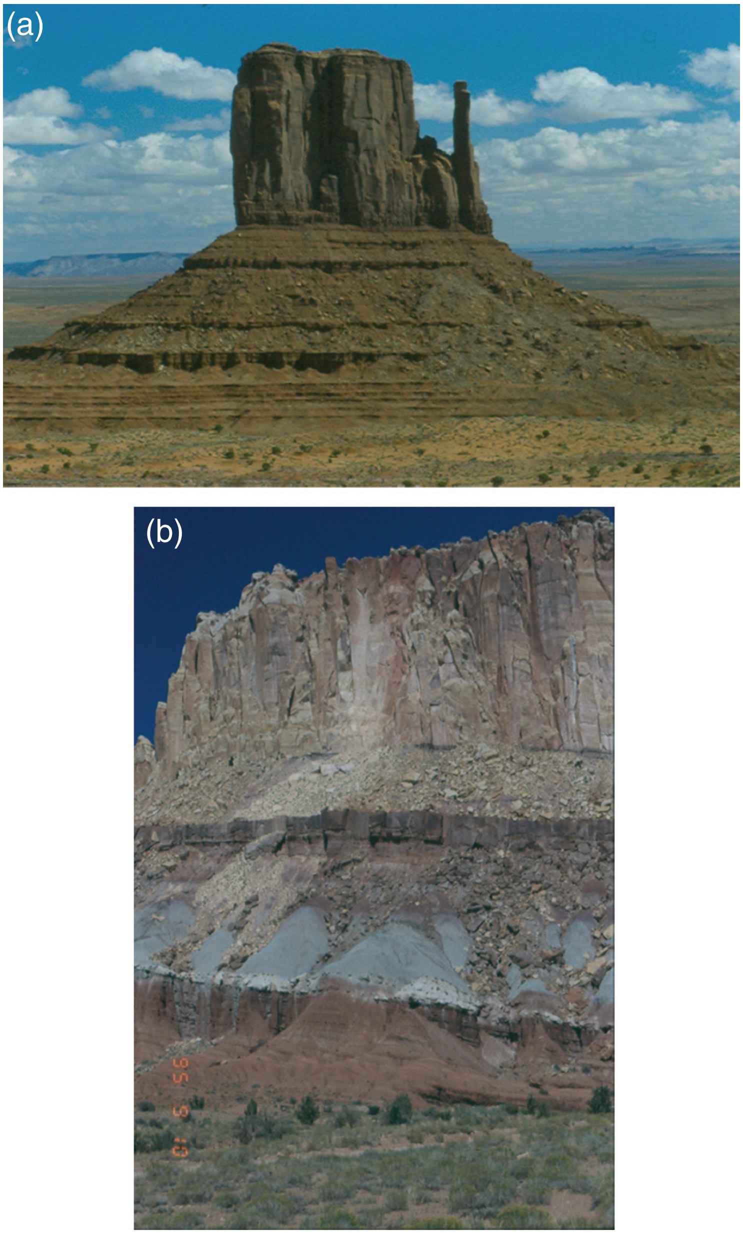

Fig. 22.3 (**a**) West Mitten Butte in Monument Valley, Arizona. Vertical cliff is formed in de Chelly sandstone with resistant Triassic layers on top, the lower slope is developed in the Organ Rock siltstones of the Cutler Formation. Note the intercalation of more resistant strata, which form ledges on the lower scarp slope. The slope belongs to slope type 3 (see Fig. 22.5). The lower slope is almost talus-free, which might lead the observer to assume general geomorphic stability. (**b**) Complex scarp in Capitol Reef National Park in southeastern Utah. The upper part consists of a vertical cliff in the Wingate Sandstone underlain by Chinle redbeds forming a type 3 slope. A fresh rockfall with large blocks covers the lower slope extending donwards to the convex slope (type 2) in the Petrified Forest Member of the Chinle Formation. Apparently, rockfalls can be triggered without pluvial sapping. This document might lead the observer to assume general geomorphic activity. However, a reliable evaluation of process activity or stability must use a time and space integrating survey

alcove formation. Slope dissection and steepening are much more important for the destabilization of the caprock, and these mechanisms are also active in the present dry period. Limestone talus lies on and below the pediment level, which was formed in the last pluvial of the Pleistocene (Soltanien) (Schmidt 1989a) (Fig. 22.4).

In the upper Colorado River basin, where there is a great variety of different climatic regimes with rainfall amounts generally in the range between 150 and 600 mm, the tributaries with the lowest precipitation and specific yield have the highest rates of mechanical denudation (Schmidt 1985), which shows that solid load production is most active in dry environments. Barth and Blume (1973) investigated cuesta scarps in several localities in the dry regions of the United States with different degrees of moisture availability from 165 mm annual precipitation with ten arid months on the central Colorado Plateau to 425 mm of annual precipitation with seven arid months in the Black Hills

Fig. 22.4 Cuesta scarp (Hamada de Meski, southern Morocco) capped by Cenomanian/Turonian limestones, which are underlain by soft continental redbeds. Climato-cyclic talus flatirons have developed on the scarp slope and pediment flatirons in the foreland. Different generations of flatirons can be discerned. The backslopes of the flatirons are firmly consolidated with carbonate cement. The talus flatiron on the right is assumed to be of Soltanien (Würm) age, the pediment flatiron on the left is assumed to be of Tensiftien (Riss) age. Note the coarse debris in the slope rill below the younger level

in Wyoming. The latter conditions might be representative of the situation in the last pluvial on the Colorado Plateau (Barth and Blume 1973, Schmidt and Meitz 2000b). They demonstrated that the cuesta scarps with the strongest morphological activity are found in the drier areas with precipitation amounts lower than 300 mm. In these areas the landform elements are interpreted as being in equilibrium with present-day processes and the authors rejected Ahnert's (1960) view of greater process activity in pluvial periods.

The preceding discussion shows that the question of alternating climates and of stable or unstable periods is difficult to resolve (i) when the climatic changes only cause minor alterations (changes in degree) in the process regime, (ii) when different climatically induced triggering processes – spring sapping under more humid conditions or lower slope dissection under more arid conditions – result in the same morphological response (rockfalls), and (iii) when different lithologies are involved, which show dissimilar reactions to increasing/decreasing humidity. The stability of erosion-controlled slopes is generally strengthened by increasing humidity (denser vegetation cover), whereas lower slopes controlled by gravitational processes become generally more susceptible to denudational activity when humidity increases.

Slope Types and Their Reactions to Climatic Change

Cuesta scarp slopes have developed in various lithological contexts and in different altitudinal belts on the Colorado Plateau in the southwestern United States. Cuesta scarps are found in elevations ranging from 1300 to 3000 m with corresponding mean annual precipitation amounts between less than 150 and more than 600 mm (Fig. 22.2). There is also a wide lithological variety of caprocks and soft lower slope substrates composing the cuesta scarps (Barth and Blume 1973, Schumm and Chorley 1966, Schmidt 1988). In the semi-arid parts (<250 mm) in the lower elevations (1300–1800 m) four different slope types were identified (concave, convex, straight and vertical) (Schmidt 1987a, 1988, Schmidt and Meitz 1996). Figure 22.5 shows the principal slope types with a brief description of their main characteristics. The classification in this altitudinal belt refers essentially to the form and process features of the lower scarp slope, which are mainly dependent on the lithological and structural attributes of the soft lower slope substrates. The form variations of the upper scarp slopes are not as evident because in the semi-arid belt, with its selective weathering and erosion, all the resistant caprocks tend to form nearly vertical cliffs. The form and process response to changing climatic conditions was systematically investigated in the more humid higher elevations of the Colorado Plateau (southern flank of Uinta Basin, Uncompahgre Uplift, Elk Ridge, Mesa Verde, for locations see Fig. 22.2). Specific attributes of the scarp slopes such as vegetation and soil cover, dissection density, activity of mass movements, talus production and general morphometry were documented (Schmidt and Meitz 1996, 2000a,b). The data collection and interpretation showed how slope attributes change with increasing altitude and precipitation under the present climatic regime (Fig. 22.6). This information can be used as an indicator of morphoclimatic changes during the cold (humid) phases of the late Quaternary in a space-time substitution. Altitudinal shifts of the hygric environment during the last glacial period

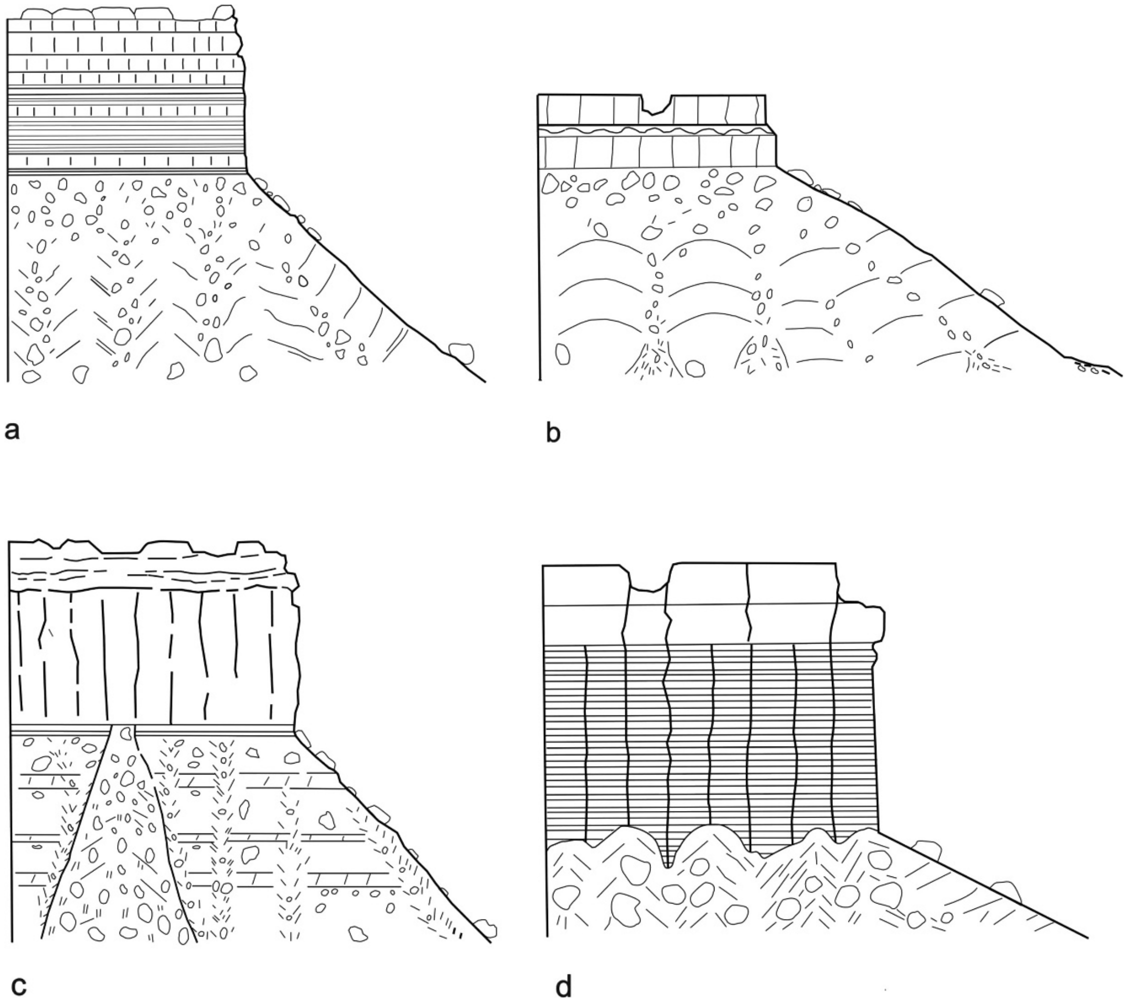

Fig. 22.5 Litologically controlled different types of scarp slopes in the arid to semi-arid altitudinal belt (Schmidt 1988, Schmidt and Meitz 2000b). (**a**) Slope type 1: concave lower slope, average inclination 35°, deeply dissected by slope rills, debris concentrated in the rills, pediment at scarp foot, developed in impermeable homogeneous shales (type 1 is shown in Fig. 22.8b, developed in the impermeable Mancos Shale), (**b**) Slope type 2: convex slopes with lower inclinations (30°), in swelling bentonitic clays, influenced by piping, only moderately dissected, debris concentrated in the upper parts of the lower slope and at the scarp foot (type 2 is shown in the central section of the complex slope in Fig. 22.3b, developed in the Petrified Forest Member of the Chinle Formation), (**c**) Slope type 3: slopes composed of segments of different inclinations in heterogeneous lower slope rocks with intercalations of more resistant beds, the resistant beds impede deep dissection, parts of the slope covered by cones of landslide material, in the overall profile the slopes are straight with inclinations close to 40° (type 3 is shown in Figs. 22.1 and 22.3a,b). (**d**) Slope type 4: vertical cliffs witrh talus slopes at their base in evenly bedded gypsiferous fine grained sandstones, siltstones and shales with vertical joints (type 4 is shown in the lower part of Fig. 22.3b, developed in the Moenkopi Formation)

resulted in associated shifts of vegetation and morphoclimatic belts, and lower elevations were affected by present-day high-altitude process combinations.

The specific responses of the principal slope types: (1) concave; (2) convex; (3) straight are demonstrated in Fig. 22.6 for north-facing slopes. Slope type 4 (vertical) showed no systematic altitudinal variations. Stratigraphic examples for the individual slope types are indicated in the captions of Fig. 22.6. In the arid to semi-arid altitudinal belt slope type 1 is deeply dissected with talus concentrated in the slope rills (cf. Fig. 22.5a). It has a concave profile with an average inclination of about 35°. The establishment of woodland vegetation in the more humid belts leads to slope

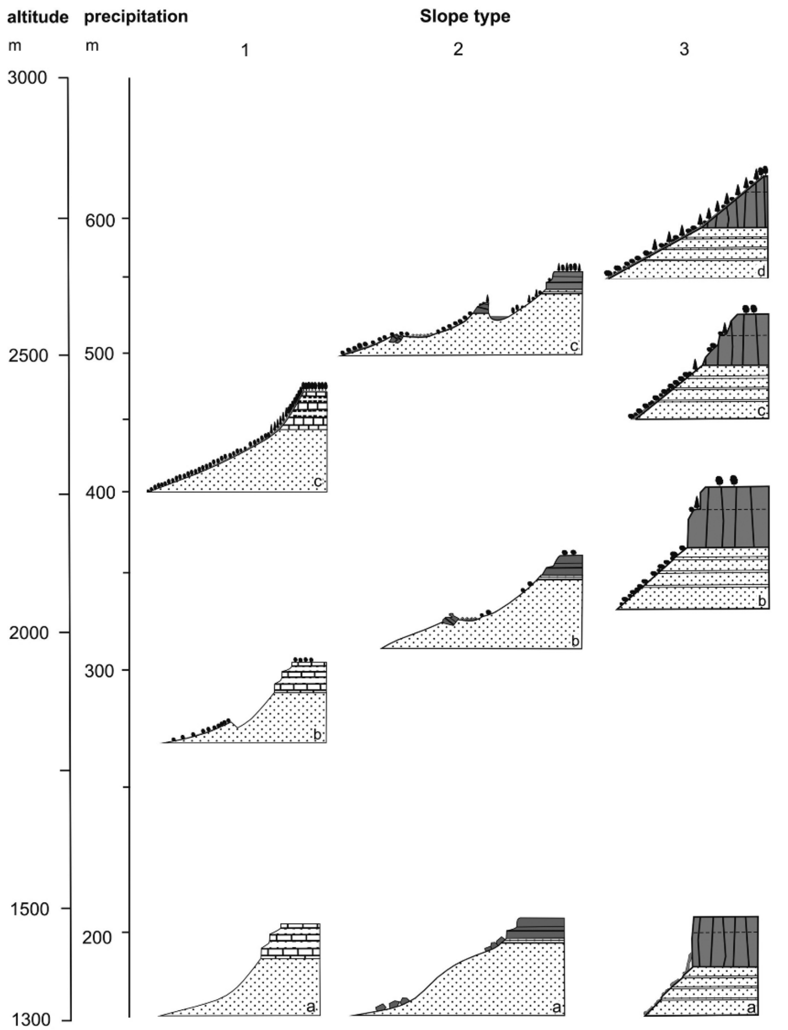

Fig. 22.6 Altitudinal and hygric zonation of the geomorphological attributes of slope types 1-3 on north-facing slopes (on south-facing slopes the altitudinal belts are shifted upwards).

1) concave slope (e.g. Mesaverde Sandstone/Mancos Shale cuesta scarp)

(**a**) arid steep, densely and deeply dissected slope (see Fig. 22.5a)
(**b**) arid to semi-arid dissected slope with talus flatirons (<350 mm)
(**c**) smooth, only moderately dissected slope in altitudes with more than 400 mm rainfall

2) convex slope (e.g. Dakota Sandstone/Brushy Basin Shale cuesta scarp)

(**a**) arid zone convex slope (see Fig. 22.5b)
(**b**) semi-arid zone with inactive mass movement accumulations in the lower parts of the slope
(**c**) active mass movements, irregular slope profile with depressions (>500 mm)

3) Straight slope type (e.g. Glen Canyon Group/Chinle Formation cuesta scarp)

(**a**) arid zone straight lower slope with vertical cliff in the upper slope (see. Fig. 22.5c, see also Fig. 22.3a)
(**b**) beginning stabilization of lower slope and beginning segmantation of the cliff (precipitation about 400 mm)
(**c**) development of small platforms on the cliff, denser vegetation cover, reduction of dissection and lower slope inclination (precipitation between 400 and 550 mm)
(**d**) moderately inclined straight slope in caprock and lower slope rocks, densely vegetated (rainfall >550 mm)

stabilization, reduced dissection and inclination and the development of a smooth profile with a continuous debris cover (profile c for slope type 1 in Fig. 22.6), Talus and pediment flatirons have originated on the Mancos-Shale slope as a consequence of alterations between cold(humid) and warm(dry) climatic phases in the intermediate altitudinal belt (profile b for slope type 1 in Fig. 22.6, see also Figs. 22.8b and 22.9a). In the semiarid altitudinal belt slope type 2 is moderately dissected and has a convex profile with an average inclination of about 30° (Fig. 22.5b). With increasing humidity the convex slope profile disappears and mass movement activity and associated irregular slope forms become more and more characteristic for the scarp slope (profile b and c for slope type 2 in Fig. 22.6). The lower limit of inactive landslides is indicative of how far humid phase mass movement processes extended downslope. On slope type 3 scarp form variations are not only found on the lower, but they are also visible on the upper scarp slope. On the lower slope in the arid to semiarid altitudinal belt dissection is less intense than on slope type 1 and talus is concentrated on the divides between the slope rills. The slope has a straight profile with inclinations close to 40° (see Fig. 22.3a for an examples from the Monument Valley rock sequence). Inclination, dissection and the amount and size of talus decrease in more humid zones. In the drier elevations the sandstones form vertical cliffs, which with increasing altitude and humidity change to horizontally segmented and compartmental cliffs and eventually to a straight slope (profiles b, c and d for slope type 3 in Fig. 22.6). Due to their relatively high resistance to change, the duration of the last cold (humid) phase was not long enough to shift the sandstone slopes of the individual altitudinal belts from their given states of form-process equilibrium to completely different ones.

Only hillslope form elements that give evidence of significantly different processes of weathering and erosion will help to elucidate the effects of climatic change and the succession of different climatic regimes. The slopes must be able to preserve forms created by past processes, i.e. the form elements must be firm and stable enough to survive climatically induced process change for a longer period of time. There are cuesta scarp hillslopes, which display multiphase assemblages of climato-genetic form elements. On slope type 2 inactive mass movement accumulations are found in the presently stable dry altitudinal belt. On slope type 1 talus and pediment flatirons give clear evidence of climatic change (Fig. 22.6).

Talus and Pediment Flatirons

Talus and pediment flatirons on the slopes of dryland cuesta scarps have a triangular to trapezoidal shape with their tops directed towards the scarp (Fig. 22.4). Talus flatirons are located on the scarp slope and pediment flatirons in the transition zone between scarp and foreland. In their distal parts, pediment flatirons may merge into river or lake basin terraces. Both the talus and pediment flatirons were parts of formerly active continuous slope systems, from which they received their debris cover. They have been detached from the presently active scarp by small rills running perpendicular to the general slope inclination, by slope steepening, and by cliff retreat. The inner slope of the flatirons is inclined towards the cuesta scarp and consists of the soft bedrock material of the lower portion of the scarp. The outer slope, which is inclined towards the scarp foreland with slope angles from 2 to 30°, is covered by a debris mantle consisting of resistant caprock talus protecting the soft lower slope rock from being eroded (Figs. 22.4 and 22.8b). The talus is generally not more than 5 m thick and has an average thickness of 2 m. In the proximal parts, however, the talus may contain blocks with *a*-axis diameters of more than 2 m which are remnants of rockfall debris.

Models of Flatiron Formation

Talus and pediment flatirons have attracted growing attention in recent years. They have been described from a variety of arid and semi-arid environments including the southwestern United States (Koons 1955, Blume and Barth 1972, Barth and Blume 1973, Schmidt 1988, 1996, Schmidt and Meitz 1996, 2000a,b), Spain (Gutiérrez et al. 1998, 2006, Sancho et al. 1988), the Saharan countries (Ergenzinger 1972, Grunert 1983), southern Morocco (Joly 1962, Dongus 1980, Schmidt 1987b, 1989a), Saudi Arabia (Barth 1976), Syria (Sakaguchi 1986), Cyprus (Everard 1963), and

Israel (Gerson 1982, Gerson and Grossman 1987). In these publications different interpretations have been offered for the processes and climatic controls involved in the formation of these landforms. The debate has centred on the question of whether the flatirons are the product of climatically controlled distinct processes or whether they develop independently from climatic change. The respective conceptions of flatiron formation are called the climato-cyclic model and the non-cyclic model (Schmidt 1989b, 1996). Only in the former case can the flatirons be used as 'keys to major climatic fluctuations' (Gerson 1982, p. 123).

The non-cyclic model, which was first proposed by Koons (1955), is characterized by the sequence of events as depicted in Fig. 22.7a. No climatic change is needed for this sequence to occur. The successive stages of non-cyclic flatiron development from talus-free slopes to fresh talus cones and finally to complete separation of cones can be mapped in the field in close spatial proximity (Fig. 22.7b). On the Colorado Plateau non-cyclic flatirons are most frequently found on slope type 3 in the arid to semi-arid altitudinal belt (Schmidt 1988, 1989b).

The climato-cyclic model explains talus and pediment flatirons as consequences of climatically dependent alternations of significantly different processes on scarp slopes (Gerson 1982, Gerson and Grossman 1987, Gutiérrez et al. 1998, 2006, Schmidt 1989a,b, 1996, Schmidt and Meitz 2000b). In the more humid periods (pluvials) scarp slopes are controlled by aquatic debris transport and slopewash, which results in the formation of smooth, undissected concave slope profiles with some gravitational activity in the proximal parts. When these conditions are replaced by more arid climatic regimes (interpluvials), dissection of the smooth debris slopes begins, bedrock is exposed in the rills, and gullying and rock collapse are the dominant processes. Parts of the formerly continuous slope are separated from the active scarp, and talus and pediment flatirons begin to form, with the backslopes of the flatirons representing the remnants of the smooth debris slopes (Figs. 22.4, 22.8a,b and 22.9a). In the current dry period, dissection of the smooth debris slopes and talus removal are ubiquitously observed processes on scarps in Northern Africa and the Middle East (Gerson 1982, Schmidt 1987b) as well as in the southwestern US on slope type 1 in the lower elevations (Figs. 22.8b and 22.9a) (Schmidt 1996).

Lithological and Structural Prerequisites for the Formation of Talus and Pediment Flatirons

Apart from climatic influence a number of lithological and structural controls are essential to the formation of the two separate types of flatirons (Schmidt 1989b). Scarps with **non-cyclic talus flatirons** need to be fed with rockfall material which covers the complete lower slope with a talus cone. This continuous talus cone can only be supplied by thick caprocks and on scarp slopes where the thickness ratio between soft rock and caprock is not much greater than unity. The angle of the lower slope must be high enough to destabilize large parts of the caprock, and dissection of the lower scarp slope must not be so deep that the rockfall talus

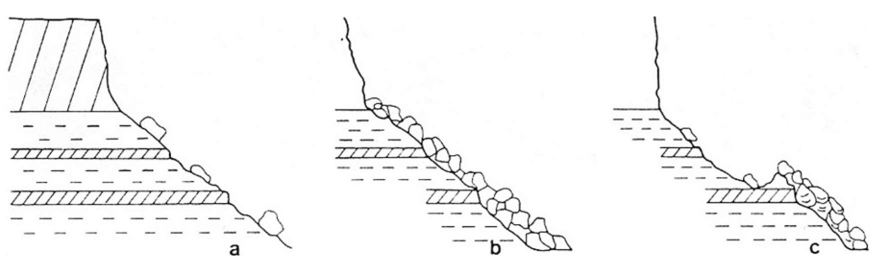

Fig. 22.7a The non-cyclic model of talus flatiron development (from Schmidt 1989b). (**a**) The caprock is undermined and destabilized by dissection and steepening of the lower slope. (**b**) A rockfall is triggered, which covers the complete lower slope with a talus cone, at the same time reducing the inclination of the lower slope. (**c**) the flanks of the talus cone are undermined by slope rills, the apex of the cone is detached from the scarp by tributaries to the major slope rills, and the talus cone becomes isolated from the scarp and a talus flatiron originates. After this succession of events the lower slope is again dissected and steepened, until a critical state of caprock destabilization is reached and a new rockfall is triggered

Fig. 22.7b Cuesta scarp with Glen Canyon sandstones overlying redbeds and bentonitic claystones (base of the scarp) of the Chinle Formation, Comb Ridge, Southeastern Utah (slope type 3). Different stages of non-cyclic talus flatiron development can be observed along this reach of the Comb Ridge. From right to left: talus flatiron separated from the scarp slope, flanks undercut by slope rills, Chinle redbeds are exposed on the flanks, apex detached from the caprock; on the left a talus cone in parts still connected with the vertical cliff; between these talus cones and on the extreme left scarp sections relatively free of talus. Note the coarse-grained talus (>3 m) on the flatiron backslopes

is trapped in the large slope gullies. All of these conditions are met on slope type 3 on the Colorado Plateau (cf. Fig. 22.7b).

As the **climato-cyclic talus flatirons** attest to past climates and climatic change, the conditions necessary for their development are of particular interest.

(a) In the more humid phase, a smooth concave debris-flow-controlled slope profile must be formed as the initial stage of flatiron formation. This is only possible in thick lower slope rocks of relatively uniform resistance. In North Africa the continental redbeds below the Cretaceous limestones belong to this category of rock type (Fig. 22.4), as does the Mancos Shale

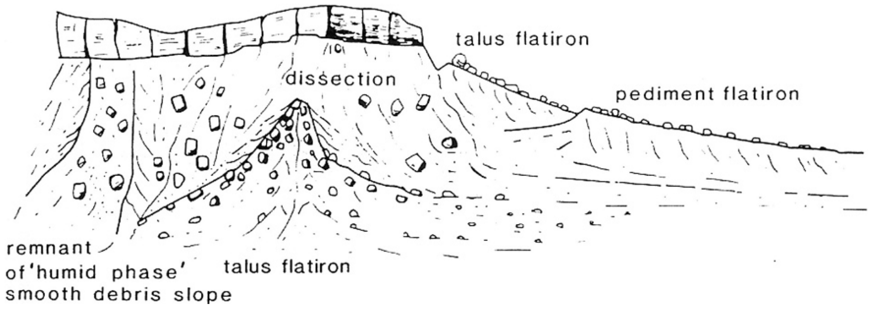

Fig. 22.8a The climato-cyclic model of talus and pediment flatiron development (from Schmidt 1989b, 1996). A smooth concave debris-flow-controlled slope is formed in the more humid phase (pluvial). The slope is then dissected and steepened in the more arid phase (interpluvial). Parts of the debris slope are detached from the active scarp by dissection and scarp retreat, and talus and pediment flatirons originate. The flatiron backslopes are remnants of the humid phase debris slope

Fig. 22.8b Scarp slope (slope type 1) on the eastern Book Cliffs, Colorado. Mancos Shale crops out on the lower scarp slope. Caprock is the Sego Sandstone of the Mesaverde Group. Remnants of the smooth, concave cold(humid) phase debris slope have been preserved. Parts of the debris covered slope are stll connected to the upper slope (left side of the photo). But most parts have been detached from the active slope, and talus flatirons have developed. The flatiron backslopes are covered with a continuous talus mantle and support the growth of dwarf junipers. The presently active warm(dry) phase steep Mancos Shale slope is deeply dissected and almost free of vegetation. Talus is concentrated in the slope rills

Fig. 22.9a Multi-level flatiron sequence on the slope of the eastern Book Cliffs, Colorado and Utah. The higher pediment flatiron in the right centre stands 50 m above the lower level. This lower level flatiron can be seen in the left centre close to the foot of the Book Cliffs. Its backslope is covered by dwarf forest. A schematic diagram of the sequence is shown in Fig. 22.9b

in the southwestern United States. Where the Mancos Shale crops out in the lower slope (slope type 1), cyclic flatirons are found on the Colorado Plateau (Schmidt 1996), such as on the slopes of the Mesa Verde near Cortez, Colorado and on the Book Cliffs, near Grand Junction, Colorado (Figs. 22.8b and 22.9a) (for locations see Fig. 22.2).

On lower scarp slopes with intercalated resistant layers (slope type 3, Figs. 22.3a and 22.5c), the resistance to change is too great for smooth debris slopes to form, and, consequently, for cyclic flatirons to develop (Schmidt and Meitz 1996). Intensity and duration of the past process were not able to gain ascendancy over lithological form resistance (Equation 22.1). On the other hand, free faces, as in the Moenkopi or Summerville Formations (Fig. 22.5d) with their short relaxation times, are also not characterized by flatiron forms (Schmidt and Meitz 2000b).

(b) The outer slopes of the talus and pediment flatirons need a highly protective talus cover to survive periods of cyclic duration. Only very resistant caprocks such as limestones, quartzites, or well-cemented sandstones and conglomerates can supply the lower slope with a sufficiently protective talus cover. The protective effect is enhanced when the talus becomes cemented by carbonates. Limestone and cemented sandstones and conglomerates can supply the slope debris with a lateral input of water containing carbonates in solution. In southern Morocco, for instance, the caprocks either guarantee the supply of carbonate cement (limestones, cemented conglomerates) or are very resistant (limestones, quartzites) (Schmidt 1987b, 1989b). Also the caprocks described from Israel (Gerson 1982) are very resistant (flints, limestones). On the Colorado Plateau, however, many of the caprocks are poorly cemented sandstones such as the Navajo, Entrada or de Chelly sandstones, but some have a stronger carbonate cement, such as parts of the Mesaverde Sandstones (Figs. 22.8b and 22.9a). If the flatirons carry a resistant backslope cover, their resistance to change and their tendency to be preserved will increase. The intensity and duration of the subsequent processes are not competent to obliterate the legacies of the past processes (Equation 22.2).

(c) The thickness ratio between soft rock and caprock is another important controlling factor. When the caprock is relatively thin and the thickness ratio in excess of 5, the mass of caprock material is not sufficient to cover and protect the complete lower slope with a talus apron, especially when the caprock is not very resistant. On the other hand, when the thickness ratio is less than unity, talus production exceeds talus transport and removal, the lower slope 'drowns' in debris and no smooth concave debris slope – the initial stage of flatiron formation – can develop. Additionally, great differences in resistance between caprock and soft rock promote cyclic flatiron formation.

(d) On complex scarps, where more than one caprock–soft-rock sequence is involved in scarp composition, the intercalated caprocks on the slopes impede the formation of a continuous concave debris slope and, hence, the development of talus or pediment flatirons (cf. Fig. 22.3b).

(e) Undercutting rivers or washes at the foot of the scarp will steepen the basal part of the lower slope and preclude basal concavities from being formed (Fig. 22.1). Also deeply incised canyons in the immediate scarp foreland with the resulting high vertical distance to base level make the evolution of cyclic flatirons impossible by decreasing the slope's resistance to change.

Climatic Thresholds

In those areas where the smooth debris slopes are being destroyed, ubiquitous dissection is taking place, and a new generation of cyclic flatirons is being formed. The annual precipitation lies between 75 and 175 mm in southern Morocco (Schmidt 1987b) and between 150 and well below 100 mm in Israel (Gerson and Grossman 1987). In both areas high precipitation intensities with intensive gullying have been recorded. The question arises: what is the threshold precipitation at which slope dissection gives way to smooth debris slope formation, and vice versa? Gerson and Grossman (1987) specifiy that a change to a moderately arid (150–250 mm) or semi-arid (250–400 mm) climate will cause the onset of 'debris-flow-controlled' processes. In a study of a cuesta scarp in the northwestern Tibesti area, Ergenzinger (1972) came to the conclusion that annual precipitation amounts in excess of 200 mm are needed for the formation of a smooth slopes controlled by sheetwash.

Investigations of Cenomanian–Turonian limestone scarps in southern Morocco showed that, although flatiron development is extremely active on the southern margin of the High Atlas, no slope dissection with flatiron formation occurs in its northern foreland (Schmidt 1989b). In this area mean annual precipitation exceeds 200 mm. In southern Morocco the 200-mm isohyet seems to represent a critical threshold value for the formation of talus relics on scarps with limestone caprocks underlain by soft continental redbeds. Apparently the smooth undissected concave lower scarp slopes, which are more densely vegetated than the slopes south of the High Atlas, are in equilibrium with the processes operating under the current climatic regime. Also no flatirons are formed in the Middle Atlas where precipitation exceeds 400 mm. Caution is needed in spatially extrapolating this threshold precipitation value of 200 mm, which is defined only for a specific set of lithological, structural, environmental, and biotic conditions.

On the Colorado Plateau smooth concave slopes are presently being formed on slope type 1 in areas with precipitation amounts exceeding 400 mm (profile c for slope type 1 in Fig. 22.6). They are dissected on cuesta scarp slopes where mean annual precipitation is 350 mm or less (profiles a. b for slope type 1 in Fig. 22.6). A detailed evaluation of the critical conditions and thresholds initiating variations in slope morphometry and process attributes for diverse slope types with different lithological and structural characteristics has been elaborated by Schmidt and Meitz (1996, 2000a,b) for the Colorado Plateau (see Fig. 22.6).

Timespans Required for the Formation of Climato-Cyclic Flatiron Forms and the Interpretation of Flatiron Sequences

The resistance to change and the relaxation times of hillslopes (see above) strongly control the timespans needed for the climato-cyclic talus relics to develop. Short-term climatic changes of a few hundred years or less and individual events are generally not recorded on hillslopes (Gerson 1982, Gerson and Grossman 1987, Littmann and Schmidt 1989). Gerson (1982) gave a tentative minimum estimate of 25 000–50 000 years for a smooth debris slope to develop after dry-phase dissection. And several thousand years are required to convert a talus slope into a slope controlled by gullying and dissectional activity (Gerson and Grossman 1987). Dissection of smooth debris slopes during the late Holocene dry period is an example (Littmann and Schmidt 1989). On the other hand, Gutiérrez et al. (2006) derive intra-Holocene ages for their youngest ^{14}C dated flatiron sequences in northeastern Spain.

There is a general paucity of chronostratigraphic information for coarse-grained hillslope deposits, since they are mostly free of fossils, organic carbon remains, or archaeological evidence. With the present shortage of absolute age determinations in most cases only some inferences about relative chronologies are possible. According to the model of climato-cyclic flatiron formation, the latest continuous smooth debris slope formed during the last wet period of at least several thousands years' duration. In Israel a talus relic has been dated by Mousterian artefacts to have originated during the early to middle Würm, which was a relatively wet period in the eastern Mediterranean region (Gerson and Grossman 1987). In southern Morocco many of the backslopes of the youngest flatiron generation merge into Soltanien (Würmian) terrace surfaces (Schmidt 1987b, Littmann and Schmidt 1989). Judging from the late Quaternary climatic record for areas where climato-cyclic flatirons have developed (northern fringe of the Sahara, Middle East, eastern Mediterranean area), periods of greater moisture availability coincide with certain stages of the last glaciation (Gerson 1982, Schmidt 1987b). Littmann (1989) showed that the Atlas region remained wetter than today until the late glacial. More arid climates with slope dissection seem to have prevailed in interglacial times (such as the Holocene).

On the slopes and in the foreland of many cuesta scarps, sequences of talus and pediment flatirons are found, each flatiron backslope preserving the latest stage of a pluvial period (Figs. 22.4, 22.8b and 22.9a). The inter-flatiron erosional gaps are the product of scarp recession during the dissection-controlled interpluvial and the subsequent debris-flow-controlled pluvial periods. They represent the denudational activity of a complete interpluvial–pluvial cycle. Multicycle flatiron generations are best preserved on scarp spurs and on the flanks of erosional outliers, where the intensity of slope denudational processes is reduced. In southern Morocco up to four successive generations have been found on the limestone scarps (Schmidt 1987b). Six levels were mapped in the Makhtesh Ramon area in Israel (Gerson and Grossman 1987). The number of flatiron generations is an indication of the number of late and middle Quaternary pluvial periods, and the distance between two successive flatirons (Fig. 22.9b) can be used for estimating scarp retreat during an interpluvial–pluvial cycle (Gerson 1982, Gutiérrez et al. 1998, Sancho et al. 1988, Schmidt 1987b, 1989a,b). To estimate rates of retreat, the lengths of the cycles must be known, which, in many cases, is still a matter of speculation. In Morocco the cycle Tensiftien (Riss) to Soltanien (Würm) lasted about 10^5 years, and the cycle Amirien (Mindel) to Tensiftien about 2×10^5 years. With these rough figures rates of retreat of about 0.5 m per 1000 years were calculated for the limestone scarps in southern Morocco (Schmidt 1987b, 1989a). In the southwestern United States rates of retreat of 3.5 m per 1000 years were estimated for the Mesaverde Sandstone/Mancos Shale cuesta scarp of the Book Cliffs for the Illinoian-Wisconsin interval (Schmidt 1996). Since the last pluvial, the scarp has retreated a mean distance of about 10 m (Fig. 22.9a,b). Progress in dating techniques will certainly help to fit the cyclic flatiron sequences into the climatic pattern of the Quaternary and make it possible to use them for palaeoclimatic interpretations in a more detailed way than presently possible (Gutiérrez et al. 2006). One of the most promising methods is surface exposure dating (Dorn and Philips 1991). Yet in the special case of the Book Cliffs the sandstones of the Mesaverde group on the backslope of the flatirons were too easily weathered to allow reliable datings (Schmidt and Meitz 1997).

Conclusion

Relief elements on desert hillslopes may serve as valuable indicators of the low-frequency climatic variations during the Quaternary because, owing to their relatively high resistance to change and their usually long relaxation times, they record changes of long duration and high process effectiveness. Particularly bipartite slopes (cuesta scarps) may bear clear testimony of climatic change. Yet many of the interpretations of climatic change derived from hillslope form and process evidence are based on inference. Surveys of sediment yield (Schmidt 1985) and comparative investigations of cuesta scarps in areas of varying moisture availability (Barth and Blume 1973, Schmidt and

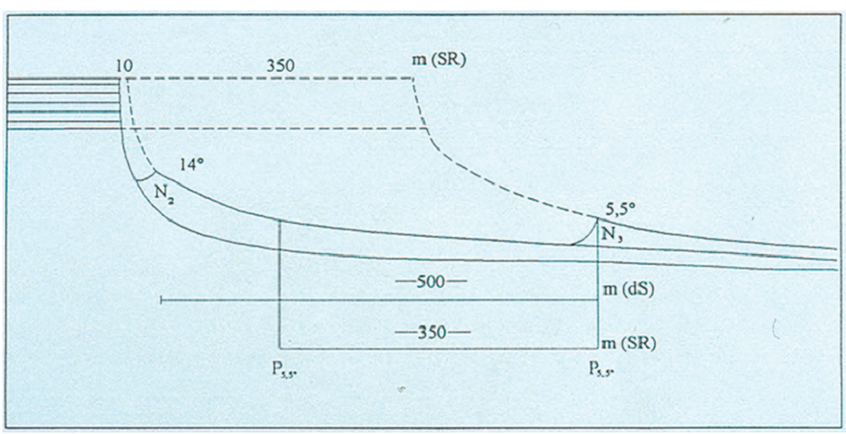

Fig. 22.9b Schematic diagram of the flatiron sequence shown in Fig. 22.9a. Sequences of flatirons can be used to estimate amounts of scarp retreat. Amounts of scarp retreat can be obtained by measuring the distance from the top of the older (*higher*) flatiron to the point of equal inclination (P 5,5°) on the next younger (*lower*) flatiron backslope. The amount of retreat (SR) in the example is 350 m. Measuring the distance between the tops of successive flatirons or the distance to scarp slope (dS) will overestimate the amount of scarp recession, because the inner slopes of the flatirons have also retreated after their formation. Assuming that the cycle from end-Illinoian to end-Wisconsin lasted about 10^5 years, a rate of retreat of about 3.5 m per 1000 years is obtained. Since the last cold(humid) phase (Wisconsin) the scarp has retreated about 10 m (Schmidt 1996)

Meitz 1996, 2000a,b) strongly indicate that in the more arid regions of the southwestern United States denudational processes are in a state of high morphodynamic activity. On the other hand, landslide-controlled slopes are more active during humid periods. On a worldwide scale, the understanding of different process combinations under changing climatic conditions remains fragmentary. With regard to gravitational processes on cuesta scarps, a promising field of research seems to be the association of alluvial chronologies with rockfalls and slope deposits to decide whether events have been more frequent during specific periods in the past.

Process alternations, which show not only changes of process rates but of process type, are of major interest for palaeoclimatic interpretations. Landform assemblages which give evidence of geomorphic processes of different character are talus and pediment flatirons. Present possibilities of dating these landform successions are still unsatisfactory, though progress has been made (Gutiérrez et al. 2006). But surface exposure dating may permit the dating of landforms in deserts previously thought to be undatable (Dorn and Philips 1991, Beck 1994) and may help attain greater precision in the palaeoclimatic interpretation of hillslope form sequences. Climate, however, is only one of several factors controlling hillslope processes, and results obtained in one region cannot readily be transferred to other regions in different structural and lithological settings.

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