Chapter 13

Pediments in Arid Environments

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Introduction

Pediments, gently sloping erosional surfaces of low relief developed on bedrock, occur in a wide variety of lithologic, neotectonic, and climatic settings. Reported on six continents, their distribution spans the range of subpolar latitudes from the Arctic to the Antarctic and the range of climate from hyperarid to humid tropical (Whitaker 1979). On the Cape York Peninsula in tropical north-east Queensland, a tectonically quiescent region of Precambrian granitic and metamorphic rocks overlain by gently dipping Mesozoic and Cenozoic sediments, pediments are a dominant landscape component occurring as fringing piedmonts and strath valleys of the crystalline rocks and as broad erosional plains on the sedimentary rocks (Smart et al. 1980). In the Gran Sabana of south-east Venezuela, a broadly upwarped region of gently dipping early Proterozoic strata, pediments comprise the floors of broad subsequent valleys developed within less resistant parts of the section (Dohrenwend et al. 1995). In the late Tertiary and Quaternary landscape of the Basin and Range geomorphic province of the south-west United States, pediments (developed within a variety of neotectonic settings and on a broad range of igneous, metamorphic, and sedimentary lithologies) occupy piedmont slopes, mountain passes, and broad topographic domes (Hadley 1967, Oberlander 1972, Cooke and Mason 1973, Moss 1977, Dohrenwend 1987a). Clearly, pediments are azonal, worldwide phenomena that tend to form under conditions of relative geomorphic stability where processes of erosion and deposition are locally balanced over the long term such that mass transport and general surface regrading dominate landscape evolution. However, pediments are most conspicuous in arid and semi-arid environments where vegetation densities are low and deep weathering is limited. Hence, they have been studied most intensively in these regions and are generally perceived as an arid land phenomenon.

Pediments have long been the subject of geomorphological scrutiny. Furthermore, 'the origin of these landforms has been controversial since Gilbert (1877) first recognized and described them' (Tator 1952, p. 294). Indeed, 'pediments have attracted more study and controversy, and have sparked the imagination of more geomorphologists, than most other landforms in deserts' (Cooke and Mason 1973, p. 188). In consequence, 'conflicting views concerning the origin of rock pediments in deserts ... probably comprise the largest corpus in the literature on arid landforms' (Oberlander 1997, p.136). This attention and controversy stem from a variety of factors which are inherent in the very nature of these landforms.

(a) Pediments are counter-intuitive landforms. To many, it is intuitive that uplands should be erosional features underlain by resistant bedrock, whereas the adjacent piedmont plains should be depositional features underlain by sedimentary deposits derived from the uplands. To others, it is intuitive that most landscapes should be adjusted such that variations in lithology/structure and form are highly correlated. In reality, most pediments are, at least in part, surfaces of erosion and transportation where the underlying lithology/

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structure may be the same as or very similar to that of the adjacent uplands.

(b) As defined in geomorphology, a pediment is a nearly planar surface of mass transport and/or laterally uniform erosion which functions as a zone of transition between a degrading upland and a stable base level or slowly aggrading lowland. As such, pediments comprise a fairly general class of landforms that occur within many different climatic regimes and geomorphic settings. They are the product of a variety of processes, and the relative significance of these processes varies from one climatic-geomorphic setting to another.

(c) Within any particular climatic-geomorphic setting, the processes which act to form and maintain the pediment operate both non-linearly and discontinuously. It follows, therefore, that both the distribution of form and the operation of process will probably be chaotic (in the true mathematical sense) within the zone of transition represented by the pediment. Such systems cannot be clearly and comprehensively described by linear models of cause and effect.

(d) As transitional landforms, pediments are partly active, partly inactive, partly dissected, and partly buried. Their boundaries are irregular, gradational, and poorly defined. In most cases, they are at least partly obscured by discontinuous veneers of alluvial deposits and/or deeply weathered bedrock; indeed, many actively forming pediments undoubtedly lie unrecognized beneath continuous mantles of such materials. These characteristics commonly frustrate attempts of measurement, description, and analysis.

Unfortunately, the net result of this long history of study is not altogether clear or cogent, and has not produced a clear understanding of the processes responsible for pediment development. In recent years, perhaps in response to the apparent intractability of the development of a clear understanding of pediment-forming processes, attention given to this landform has waned.

Definitions

The confusion and controversy which pervade the study of pediments encompass even the basic definition of the term. Originally used as an architectural term, pediment refers to the vertically oriented triangular termination (or gable) of a gently pitched ridge roof (Dinsmoor 1975, p. 394, Janson 1969, p. 90). When borrowed from architecture and used in the field of geomorphology, however, the term quickly acquired an entirely different meaning (i.e. a gently sloping erosional surface of low relief). Morphologically, the gently sloping erosion surface of geomorphology corresponds much more closely to a gently pitched roof than to the vertical termination of that roof.

Of more fundamental significance, however, is the fact that pediments are morphologically complex landforms that form within a wide variety of geomorphic environments. It is hardly surprising, therefore, that the variability in pediment definition is nearly as great as the variability in pediments themselves (Tator 1953, Cooke and Mason 1973, p. 188, Whitaker 1979, Oberlander 1989).

Only the most general definitions can accommodate the broad range of form and wide variety of geomorphic environments associated with pediments. The definition proposed by Whitaker (1979, p. 432) is a case in point. ‘A pediment is a terrestrial erosional footslope surface inclined at a low angle and lacking significant relief. . . . It usually meets the hillslope at an angular nickline, and may be covered by transported material’. Unfortunately, such definitions are so broad that they limit the utility of the term as a basis for meaningful geomorphic inquiry and comparison. Qualifying modifiers such as those suggested by Tator (1953, p. 53) partly overcome this problem. ‘The term pediment should be retained for the broad (but individually distinct) degradational surface produced by subaerial processes (including running water) in dry regions. Qualifications as to physiographic location may be expressed by the use of mountain, piedmont, or flat-land . . . Additional terminology should include the words suballuvial, for alluviated erosion levels, and subaerial, for non-alluviated levels’. However, the problem of describing pediments in a consistent and scientifically productive manner remains.

Other definitions increase precision but at the expense of general applicability. Consider for example, two definitions that apply specifically to piedmont pediments in the south-western United States:

> The term pediment may be restricted to that portion of the surface of degradation at the foot of a receding slope, which (1) is underlain by rocks of the upland and which is

> either bare or mantled by a layer of alluvium not exceeding in thickness the depth of stream scour during flood, (2) is essentially a surface of transportation, experiencing neither marked vertical downcutting nor excessive deposition, and (3) displays a longitudinal profile normally concave, but which may be convex at its head in later stages of development (Howard 1942, p. 8).
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> . . . pediments are composed of surfaces eroded across bedrock or alluvium, are usually discordant to structure, have longitudinal profiles usually either concave upward or rectilinear, slope at less than 11°, and are thinly and discontinuously veneered with rock debris. The upper limits of pediments are usually mountain/piedmont junctions, although pediments may meet along watersheds; pediments are generally masked downslope by alluvium, and their lower boundaries are the lines at which the alluvial covers become continuous (Cooke 1970, p. 28).

These definitions are useful general descriptions of piedmonts; however, they are not entirely satisfactory when applied to pediment domes and terrace pediments in the same region. Moreover, although generally similar, they are not in complete agreement. Cooke placed the downslope boundary of the pediment where alluvial cover becomes continuous, whereas Howard defined it as the point where the thickness of cover equals the depth of effective stream scour. Tator (1953) concurred that the thickness of alluvial cover commonly approximates the depth of scour. However, Bull (1977) suggested that the downslope limit be defined where the thickness of alluvial cover exceeds one percent of the piedmont length. Such wide definitional disparities obstruct comparative analysis and synthesis of the pediment literature.

For the purposes of the present discussion, a pediment is descriptively defined as a gently sloping erosional surface developed on bedrock or older unconsolidated deposits. This erosion surface may be subaerially exposed or covered by a discontinuous to continuous veneer of alluvial deposits. Its downslope limit may be defined (following the suggestion of Bull, 1977) as that point where the deposit thickness exceeds a small fraction of the pediment length (e.g. 0.5–1.0%) or some arbitrarily defined maximum thickness (e.g. 25–30 m), whichever is less. The bedrock may include essentially any lithologic type with any structural attitude. An erosional surface developed on piedmont or basin fill deposits is no different genetically from an erosional surface developed on plutonic or metamorphic rocks. Indeed, at least some pediment surfaces extend without interruption across high-angle contacts between crystalline bedrock and partly indurated alluvium (Dohrenwend et al. 1986). (Of course, where the underlying sediments are neither indurated nor deformed, discrimination between pediment and alluviated piedmont may be difficult.)

Classification

That pediments are highly variable landforms is further illustrated by the variety of schemes that have been proposed to classify them. Cooke and Mason (1973) offered a classification based on general geomorphic environment: (a) an apron pediment is located between watershed and base level, usually between an upland and a depositional plain; (b) a pediment dome occurs on upland slopes (and crests) that are not surmounted by a mountain mass; and (c) a terrace pediment is developed adjacent to a relatively stable base level such as a through-flowing stream. Twidale (1983) and Bourne and Twidale (1998) presented a morphogenic classification based on relations between surface material and underlying lithology which are used to infer pediment forming processes: (a) a mantled pediment occurs where crystalline bedrock is veneered by a residual weathering mantle and which is inferred to have been formed by subsurface weathering of the crystalline bedrock and wash removal of the resulting debris; (b) a rock pediment forms where crystalline bedrock is exposed at the surface and which is inferred to respresent the exposed weathering front of a formerly mantled pediment; and (c) a covered pediment is characterized by a veneer of coarse debris covering an erosional surface that cuts discordantly across sedimentary strata. Oberlander (1989) advocated a two-part classification based on the relations between the pediment, its adjacent upland, and the underlying lithology and structure: (a) a glacis pediment (*glacis d'erosion en roches tendres*; Dresch 1957, Tricart 1972) is an erosional surface that bevels less resistant materials but does not extend into adjacent uplands underlain by materials of greater resistance, whereas (b) a rock pediment occurs where there is no change in lithology between the erosional surface and the adjacent upland. Applegarth (2004) distinguished between bedrock pediments and alluvial slopes which exhibit a bedrock platform but where this platform is buried 'under alluvial debris at a depth measured in meters' (p. 225). Applegarth's distinction draws into focus the fact that landforms exist on a continuum and that strict definitions are impossible. Pediments form part of the low-angle piedmont, often surrounding mountains. The supply of sediment to the piedmont and that rate at which it is moved across it will determine the depth of alluvial cover and the degree to which the gradient of the bedrock surface accords with the topographic gradient. At one end of the continuum lies a bare bedrock surface; at the other lies a thick accumulation of sediment in which the gradient of the underlying bedrock surface may or may not accord with the topographic gradient. Somewhere along this continuum lies the distinction geomorphologists make between pediments and alluvial fans. The various classifications provide general conceptual frameworks for the study and analysis of pediments; however, they tend to be somewhat parochial in perspective and are not entirely consistent or compatible with one another.

The Pediment Association

Geomorphically, the pediment is only part of an open erosional-depositional system termed the pediment association by Cooke (1970). This system includes the pediment, the upland area tributary to it, and the alluvial plain to which it is tributary (Fig. 13.1). Johnson (1932b, p. 399) conceptualized the system as 'three concentric zones in each of which the dominant action of streams differs from that in the other two: (1) an inner zone, the zone of degradation, corresponding closely to the mountainous highland, in which vertical down-cutting of the streams reaches its maximum importance; (2) An intermediate zone, the zone of lateral corrasion, surrounding the mountain base, in which the lateral cutting by streams attains its maximum relative importance. This is the zone of pediment formation. (3) An outer zone, the zone of aggradation, where upbuilding by deposition of alluvium has its maximum relative importance'.

Morphologically, the pediment is the most stable component of this open system. As surfaces of fluvial transport (in arid and semi-arid environments, at least) which operate over long periods of time at or very close to the threshold for critical power in streams, pediments preserve little evidence of their own evolutionary history. Consequently, relatively little can be learned regarding pediment formation from the pediment surface itself. With the exception of those few places where relict pediment surfaces are preserved beneath lava flows, or other similar caprock materials that can be precisely dated (Oberlander 1972, Dohrenwend et al. 1985, 1986), geologic evidence of pediment formation is best preserved within other components of the pediment association: the geomorphic record preserved in relic upland landforms, the stratigraphic record contained within deposits of the alluvial plain, and the ongoing process/landform transition of the piedmont junction. Thus, analysis of pediment formation is best approached, in most cases, through study of the entire pediment association.

Pediment Morphology

Certainly one of the most remarkable physical attributes of any pediment is the generally planar and

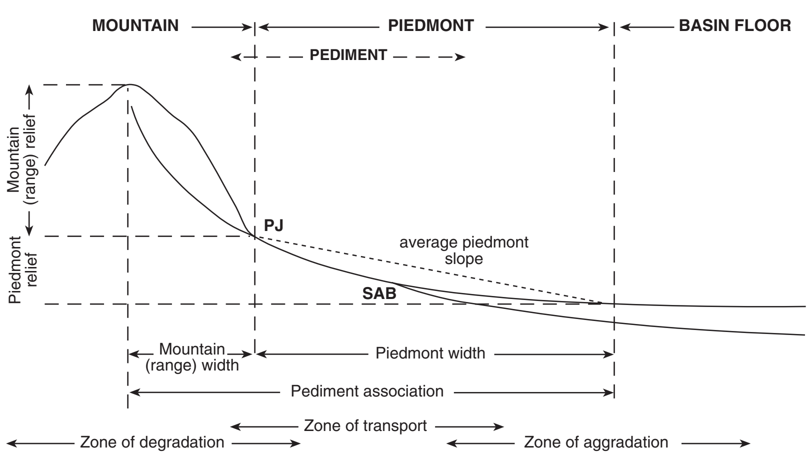

Fig. 13.1 The pediment association 'includes the pediment, the mountain drainage basins tributary to it, and the alluviated plain to which it is tributary' (Cooke 1970, p. 28). PJ = piedmont junction; SAB = subaerial alluvial boundary

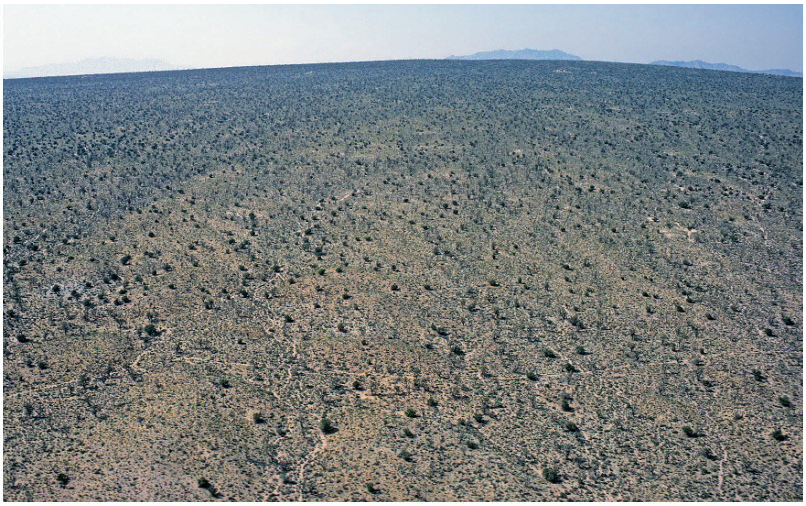

Fig. 13.2 The broad, gently sloping, largely undissected surface of Cima Dome. One of the largest pediments in the eastern Mojave Desert, Cima Dome is approximately 16 km across and 600 m high. Aerial view is south-east

featureless character of most (or at least part of) its surface (Figs. 13.2 and 13.3). Consider, for example, this description of the pediments fringing the Sacaton Mountains of southern Arizona: 'The surface which encircles the mountains is remarkably smooth, being scarred only by faint channels rarely more than a foot deep. Near the mountains it has a slope of 200–250 feet per mile but at a distance of a few miles it is so flat that films of water cover broad areas after heavy rains' (Howard 1942, p. 16). Despite this general simplicity, detailed examination of pediment morphology commonly reveals substantial complexity; '... a pediment which is a clean, smooth bedrock surface is rare indeed' (Cooke and Mason 1973, p. 196). In many instances, the proximal pediment surface is characterized by an irregular mosaic of exposed bedrock and veneers of thin alluvial cover and (or) residual regolith. On active pediments, exposed bedrock areas may be shallowly dissected and gently undulating with several metres of local relief, whereas mantled areas are typically only slightly dissected to undissected with generally less than 1–2 m of local relief. Either type of surface may be interrupted by isolated knobs, hills, and ridges of bedrock that stand above the general level of the surrounding erosional plain.

Significant aspects of pediment morphology include: (a) the character and form of pediment boundaries including the upslope limit (piedmont junction) and downslope limit (upslope limit of alluvial cover);

Fig. 13.3 Partly alluviated pediment surface on metamorphic rocks (including gneiss, granitoid rocks, diabase dykes, schist, marble, and quartzite), west piedmont of the Chemeheuevi Mountains. Aerial view is south across Chemeheuevi Valley, eastern Mojave Desert

(b) characteristic surface profiles, both longitudinal and transverse; (c) patterns of surface drainage and drainage dissection; (d) the form and distribution of major surface irregularities including inselbergs and tors; and (e) the character and distribution of the associated regolith. These morphologic characteristics are determined, at least in part, by a variety of influences including lithology and structure, relative size and shape of the piedmont and its associated upland, tectonic history, and climate. Consequently, they may vary widely even within the limits of an individual pediment.

Form and Character of the Mountain Front

The boundary between an upland area, mountain or mountain range, and its associated piedmont is commonly termed the mountain front. However, 'mountain front is an unfortunate term inasmuch as it suggests an almost linear and continuously outfacing boundary between the mountain mass and the pediment, whereas virtually all mountain fronts are indented to some degree by embayments' (Parsons and Abrahams 1984, p. 256). These embayments are, for the most part, widened valley floors that tend to form along the mountain front whenever the rate of erosional retreat of the valley sides exceeds the rate of incision of the valley floor. Such conditions are common where the mountain mass is no longer undergoing rapid uplift and local base level is either stable or rising. Consequently, the sinuosity (degree of embayment) of a tectonically inactive mountain front tends to increase with time (Bull and McFadden 1977).

The embayed nature of inactive mountain fronts has been recognized by many workers, and its significance relative to the formation and expansion of pediments is well established (e.g. Bryan 1922, Johnson 1932a, Sharp 1940, Howard 1942, Parsons and Abrahams 1984). Indeed, Lawson (1915, p. 42) was among the first to describe the general progression of embayment development. 'The contour of the subaerial [mountain] front for the greater part of the time of its recession is not in reality a straight, or even a gently sinuous line, but is actually indentate. At the indentations are cañons or gullies and from these emerge the greater part of the detritus which forms the embankment [alluvial plain] and which is distributed radially from an apex in, or at the mouth of, every cañon or gully. . . . The identations may be slight in the early stages of front recession, is most pronounced in the middle stages and becomes less intricate in the later stages.' Parsons and Abrahams (1984, p. 258) further define the process as follows, 'For a particular mountain mass during a particular period of time, the relative effectiveness of divide removal [between embayments] and mountain front retreat in extending a piedmont will depend on the relative rates of migration of piedmont junctions in embayments and along mountain fronts and the relative lengths of embayments and mountain fronts'.

Form and Character of the Piedmont Junction

The junction of the mountain slope and piedmont is commonly termed the piedmont junction. In many areas, this boundary is marked by an abrupt and pronounced break in slope. It may be particularly abrupt and well-defined where (a) an active fault bounds the mountain front, (b) marginal streams flow along the base of the upland slope, (c) slopes are capped with resistant caprock, (d) slopes are coincident with bedrock structure (dykes, joints, fault line, steep-dipping to vertical beds, etc.), or (e) a pronounced contrast exists between debris sizes on the hillslope and on the piedmont surface. In other areas, particularly along the dipslope flanks of tilted range blocks, piedmont surfaces may merge smoothly, almost imperceptibly, with the range slope and, in the case of gentle asymmetric tilting, may even extend up to and be truncated by the range crest. In yet other areas, particularly where a proximal pediment is pervasively dissected, the piedmont junction may be represented by a diffuse transition zone as much as several hundred metres wide and highly irregular both in plan and in profile. In such locations, the junction may be so irregular that it cannot be mapped as a discrete boundary. Indeed, even in those areas where the junction appears to be relatively well-defined when viewed from a distance, closer examination commonly reveals considerable complexity in both plan and profile.

The local character of the piedmont junction is determined in part by the nature of the transition from relatively diffuse erosional processes on the hillslope to somewhat more concentrated fluvial processes on the piedmont surface. As Bryan (1922, pp. 54–5) observed '. . . the angle of slope of the mountain is controlled by the resistance of the rock to the dislodgement of joint blocks and the rate at which these blocks disintegrate. . . . Fine rock debris is moved down the mountain slope by rainwash and carried away from the foot of the slope by rivulets and streams that form through the concentration of the rainwash. . . . The angle of slope of the pediment, however, is due to corrasion by the streams, and this corrasion is controlled by the ability of the water to transport debris. . . . The sharpness of the angle between the mountain slope and the pediment is one of the most remarkable results of the division of labor between rainwash on the mountain slopes and streams on the pediment'. Where this transition is profound, the piedmont junction is abrupt; where the transition is more gradual, the junction is typically less well defined. In those areas where the junction separates hillslope from piedmont interfluve the abruptness of the transition appears to be related to hillslope lithology and structure. 'In quartz monzonite areas there is often a marked contrast between debris sizes on mountain fronts and on pediments, which may account for the distinct break of slope between the two landforms in these areas; on other rock types, where the particle-size contrast is less marked, the change in slope is less abrupt' (Cooke and Mason 1973, p. 195).

In the extreme case where no transition in process occurs, there is no piedmont junction. The "sharp break in slope" between the pediment surface and the mountain front . . . exists only in interfluvial areas. . . . The course of any master stream channel from a given drainage basin in the mountains onto the pediment surface and thence to the basin floor below has no sharp break in slope. In the absence of constraints, such as recent structural disturbances, any stream channel will exhibit a relatively smooth, concave upward, longitudinal profile that accords with the local hydraulic geometry. . . . The interfluvial areas, however, generally do exhibit a marked change in slope. . . . The reason for the existence of such a zone is precisely that it is an interfluvial area; the dominant process that operates on the mountain front is not fluvial' (Lustig 1969, p. D67).

The Alluvial Boundary and the Suballuvial Surface

The transition between exposed bedrock and alluvial cover on pediments is not well documented. The very nature of this boundary impedes detailed mapping, measurement or quantitative description. The few general descriptions available show that the subaerial alluvial boundary is gradational, discontinuous, and highly irregular (e.g. Sharp 1957, Tuan 1959, Dohrenwend et al. 1986). Several factors contribute to this complexity. (a) The suballuvial bedrock surface (termed the suballuvial bench by Lawson, 1915) subparallels the surface of the alluvial mantle; its longitudinal profile is nearly rectilinear and its average slope is only very slightly steeper than the average slope of the subaerial surface (Cooke and Mason 1973). (b) On many pediments in the Mojave and Sonoran deserts, alluvial thicknesses of a few metres or less are typical for distances of at least 1 km downslope from exposed bedrock surfaces (Howard 1942, Tuan 1959, Cooke and Mason 1973, Dohrenwend unpublished data). (c) The subaerial alluvial boundary appears to be highly transient, migrating up and downslope across the piedmont in response to long-term changes in stream power and sediment load. Upslope migration is documented by buried palaeochannels of similar dimensions to active surface channels (Cooke and Mason 1973, Dohrenwend unpublished data), whereas downslope migration and proximal exhumation are suggested by the pervasive dissection of exposed bedrock surfaces (Cooke and Mason 1973).

Longitudinal Slope

The longitudinal slope of pediments typically ranges between 0.5° and 11° (Tator 1952, Cooke and Mason 1973, p. 193, Thomas 1974, p. 217). Slopes in excess of 6° are uncommon except in proximal areas. In the south-west United States, average longitudinal slopes typically range between 2° and 4° and rarely exceed 6° (Table 13.1).

There is general agreement that the characteristic form of the longitudinal profile of most pediments is concave-upward (e.g. Bryan 1922, Johnson 1932b,

Table 13.1 Representative pediment slope data (degrees), south-western United States

Area	Range of longitudinal slopes	Mean longitudinal slope	Number of pediments	Reference
Western Mojave Desert (granitic)	0.5–5.37	2.5	37	Cooke (1970)
Western Mojave Desert (non-granitic)	0.75–5.15	2.8	16	Cooke (1970)
Eastern Mojave Desert (granitic)	0.8–5.3	2.0	5	Dohrenwend (unpublished data)
Eastern Mojave Desert (granitic domes)	1.0–6.0	3.4	9	Dohrenwend (unpublished data)
Eastern Mojave Desert (non-granitic)	0.7–4.4	1.9	5	Dohrenwend (unpublished data)
Mojave Desert (granitic)	1.6–8.6*	3.4	17	Mammerickx (1964)
Sonoran Desert (granitic)	1.95–5.0*	3.5	4	Mammerickx (1964)
Mojave and Sonoran deserts (non-granitic)	1.25–4.4*	2.8	5	Mammerickx (1964)
South-eastern Arizona	0.5–2.2	–	–	Bryan (1922)
	0.5–3.3	–	–	Gilluly (1937)
Ruby–East Humboldt Mountains, Nevada (west flank)	0.8–3.8	–	–	Sharp (1940)
Ruby–East Humboldt Mountains, Nevada (east flank)	4.3–5.4	–	–	Sharp (1940)

* Range of mean values.

Lustig 1969, p. D67, Thomas 1974, p. 217). Although Davis (1933), Gilluly (1937) and Howard (1942) described convex-upward profiles for the crestal areas of pediment passes and pediment domes; rectilinear to concave longitudinal profiles typify most slopes on these landforms except in the immediate vicinity of the crest (Sharp 1957, Dohrenwend unpublished data; Fig. 13.2). The pediments of arid and semi-arid regions are generally considered to be graded surfaces of fluvial transport where the surface slope is adjusted to the discharge and load of the upland/piedmont drainage, and the characteristic concave longitudinal profile is the result of this adjustment to fluvial processes. Systematic relations between pediment slope and stream size clearly demonstrate this adjustment. In proximal areas, for example, pediment slope is steeper immediately downslope from intercanyon areas than downstream from valley mouths (Bryan 1922). Also, as Gilluly (1937, p. 332) observed in the Little Ajo Mountains of south-west Arizona, ‘the pediment is, for the most part, concave upward in profile, generally more steeply along the smaller stream courses [3.25°], less steeply along the larger streams [0.75°]’. Local variations in pediment slope that are closely related to lithologic variations also testify to this adjustment. ‘The longitudinal profile (concave) should not be assumed to be a smooth curve in all cases. It is rather, similar to the average stream profile, segmented, the smoother portions being developed across areas of homogeneous rock. Across heterogeneous rocks the profile is segmental, steeper on the more resistant and gentler on the less resistant types’ (Tator 1952, p. 302).

The influences of lithology and climate on pediment slope are not well defined; however, they appear to be relatively insignificant. On the basis of a morphometric analysis, Mammerickx (1964) concluded that bedrock is not a significant factor in determining average pediment slopes in the Sonoran and Mojave Deserts. She supported this conclusion by documenting the apparently random asymmetry of a number of pediment domes in the region. ‘On any dome it is difficult to imagine that the amount of rain or debris supplied to the different slopes is different, yet they are asymmetrical. The asymmetry further is not systematic in any particular direction’ (Mammerickx 1964, p. 423). Comparison of slope means and ranges for granitic versus non-granitic pediments in the south-west United States (Table 13.1) also suggests the validity of this conclusion. The influence of climate on pediment slope also appears to be relatively insignificant. The typical range of longitudinal slopes appears to be essentially the same in wet–dry tropical environments as it is in arid and semi-arid environments (Thomas 1974, p. 223–4).

The influence of tectonism on pediment slope is similarly ill-defined but may be considerably more significant. Cooke (1970) documented a significant positive correlation between the relief:length ratio of the pediment association and pediment slope for 53 pediment associations in the western Mojave Desert. Because the relief:length ratio of the pediment association is very likely at least partly determined within this region by the local history and character of tectonic movement, the possibility of an indirect influence of tectonism on the slope of these pediments is indicated by this correlation. This tentative conclusion is supported by a comparative analysis of the general morphometric characteristics of several diverse neotectonic domains within the south-west Basin and Range province (Dohrenwend 1987b). Average values of piedmont slope for each of these domains show a strong positive correlation with average values of range relief, which in turn show a strong positive correlation with a morphometric index of relative vertical tectonic activity (Table 13.2). This suggests that average piedmont slope is largely determined by the general morphometric character of each domain (e.g. initial range relief, range width and spacing, etc.) which in turn is largely determined by the style, character, and timing of local tectonism.

Transverse Surface Profile

Although the longitudinal pediment profile is, for the most part, slightly concave-upward, the form of the pediment surface transverse to slope may assume any one of several general forms; convex-upward or fan-shaped (Johnson 1932b, Rich 1935, Howard 1942), concave-upward or valley-shaped (Bryan 1936, Gilluly 1937, Howard 1942), or essentially rectilinear (Tator 1952, Dohrenwend unpublished data). These general transverse forms are usually better developed in proximal areas of the pediment (Tator 1952). However, undulating surfaces displaying irregular combinations of convex, concave, and rectilinear slope elements superimposed on one or more of these general forms are also characteristic where proximal areas are pervasively dissected. Undissected distal areas approach a gently undulating to nearly level transverse form (Tator 1952).

Arguments based on the presence or absence of a particular transverse surface form have been used to advocate various conceptual models of pediment development. For example, Johnson (1932b) considered rock fans (a proximal pediment with a general fan shape in plan and a convex-up transverse profile) to be compelling evidence of lateral planation by streams; whereas, Rich (1935) argued that such forms may also be produced by repeated stream capture induced by unequal incision of drainage with upland sources as compared to drainage without upland sources. Howard (1942) suggested that the transverse profiles of pediment embayments may be either convex, concave, or a combination of these forms depending on the relative activity of trunk and tributary streams within the embayment. As these and other models suggest, the tendency for water flow to channelize and then to concentrate via the intersection of channels acts to produce a variety and complexity of transverse surface forms.

Patterns of Drainage and Drainage Dissection

According to Cooke and Mason (1973, p. 197), the general pattern of drainage on pediments is essentially identical to the characteristic drainage patterns of alluvial fans and bajadas. Three interrelated types of drainage commonly develop: (a) distributary networks which radiate from proximal areas to gradually dissipate in medial and/or distal parts of the piedmont; (b) frequently changing, complexly anastomosing networks of shallow washes and rills in medial piedmont areas (Fig. 13.4); and (c) in areas of falling base level, integrated subparallel networks which are generally most deeply incised in distal areas of the alluviated piedmont plain. The transitions between these various drainage types are typically gradational and diffuse.

Detailed examination of pediment drainage, however, reveals a somewhat more complex situation than that portrayed by this generalized model (Fig. 13.5). Although many large pediments are generally smooth and regular with less than a few metres of local relief, a more complex morphology occurs where shallow drainageways locally incise the pediment surface into irregular patchworks of dissected and undissected topography. The undissected areas are mostly flat with shallow, ill-defined, discontinuous to anastomosing drainageways and low indistinct interfluves, whereas dissected areas are typically scored by well-defined

Table 13.2 Morphometric summary of the west-central Basin and Range province average range and piedmont dimensions

Tectonic domain	Average range relief, Rr* (km)	Average piedmont relief, Pr (km)	Average range width, Rw (km)	Average piedmont width, Pw (km)	Average slope of piedmont association (Rr + Pr)/(Rw + Pw)	Averagae piedmont slope† (Pr/Pw)
Central Great Basin	0.73	0.21	6.20	5.90	0.077	0.034
South-east Great Basin	0.63	0.24	3.15	6.00	0.095	0.040
South-west Great Basin	1.38	0.29	6.90	4.90	0.142	0.059
Northern Mojave Desert	0.49	0.30	2.85	8.20	0.072	0.037
Walker Lane Belt						
North-west Goldfield Block	0.64	0.27	3.30	6.10	0.096	0.044
North-east Goldfield Block	0.49	0.24	3.25	7.70	0.066	0.031
Spring Mountains Block	1.19	0.51	9.25	11.9	0.080	0.043

* Average range relief $Rr = 0.15Ti + 0.16$, where the index of relative vertical tectonic activity $Ti = (Rr/Pr) + (Rw/Pw)$; $r = 0.767$, $p = 0.044$.

† Average piedmont slope $(Pr/Pw) = 0.022Rr + 0.024$; $r = 0.829$, $p = 0.021$.

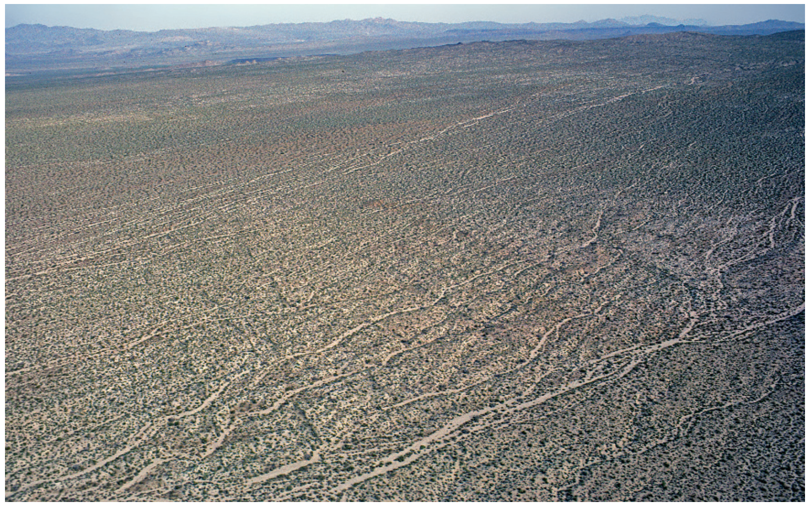

Fig. 13.4 The west flank of the Granite Springs pediment dome (eastern Mojave Desert) has been dissected into a washboard of low rounded interfluves by a fine-textured network of shallow, subparallel to complexly interconnecting washes. Aerial view is north

subparallel drainageways that form shallow regularly spaced valleys separated by rounded interfluves (Dohrenwend et al. 1986, p. 1051). 'If the dissection is not deep (less than 3 feet), the ground surface of the pediment is undulatory and the stream pattern is braided. If the pediment is deeply dissected (more than 3 feet) the stream pattern is essentially parallel, weakly integrated, and not braided' (Mammerickx 1964, p. 423). In the specific case of Cima Dome in the eastern Mojave Desert, 'dissection has been greatest on the east and southeast slopes. . . . Other parts of the dome, particularly the upper slopes and lower south flank, have washes between residual knobs and ridges. These are discontinuous and seemingly due to channelization of runoff between the residuals. . . . Some of the lowermost smooth alluvial slopes have small channels as much as 1 foot deep, 2–3 feet wide, and 100–200 feet long that start abruptly and end in a lobate tongue of loose grus . . .' (Sharp 1957, p. 277).

One of the more noteworthy features of pediment drainage, at least within the desert regions of western North America, is the characteristic tendency for dissection of proximal areas (Fig. 13.6). On some pediments, proximal dissection is largely restricted to small isolated areas along the mountain front; on others, particularly those surrounding small residual mountain masses, it forms broad continuous zones of intricately scored topography that extend as much as 2 or 3 km downslope from the mountain front. This dissection likely results from several diverse influences

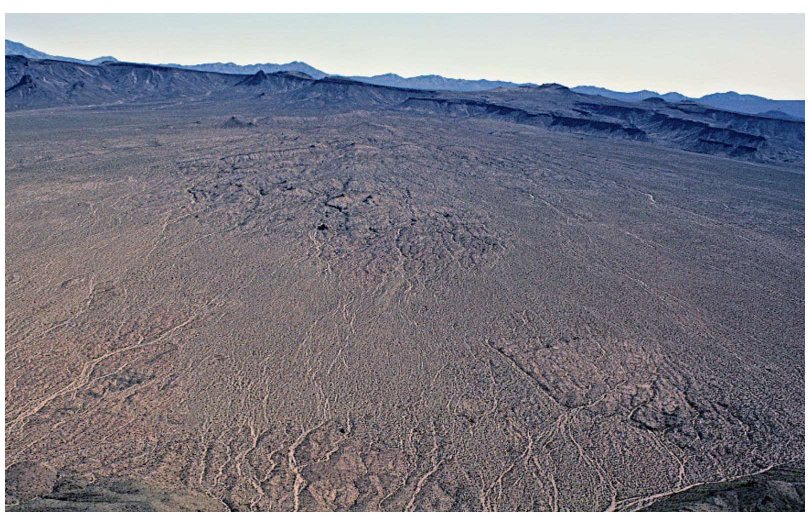

Fig. 13.5 'Chaotic' patchwork of dissected and undissected bedrock surfaces on Cretaceous monzogranite, west flank of the Granite Springs pediment dome, eastern Mojave Desert. Pliocene lava flows of the Cima volcanic field cap the deeply embayed erosional escarpment in the background

Fig. 13.6 Localized dissection of a proximal pediment surface cut on Mesozoic monzogranite. 'Sediment-starvation' has lowered the critical power threshold of this piedmont-sourced drainage system causing intense dissection (of as much as 20 m) of the pediment surface. Adjacent undissected areas are traversed by sediment-satiated streams fed by large drainage basins in the adjacent uplands. Aerial view is south-east across the south piedmont of the Granite Mountains, eastern Mojave Desert

including short-term climate change (Bryan 1922) and the tendency for differential incision of drainage with upland sources as compared with drainage where sources are limited to the piedmont (Rich 1935, Denny 1967). However, regrading of proximal piedmont areas in response to evolutionary reduction of the associated mountain mass may be the dominant influence, particularly where a broad continuous pediment surrounds a small upland mass (Johnson 1932a,b, Howard 1942, Cooke and Mason 1973). 'As the mountains dwindle in size, precipitation and the volume of sediment available for streams decrease. The decrease in load more than compensates for the loss of water, so that the dwindling waters are gradually rejuvenated. They thus lower their gradients, but the lowering is greatest in their headwater regions because farther down the slope the load is increased by pediment debris' (Howard 1942, p. 135). The occurrence of proximal dissection on many pediments indicates that pediment regrading does not necessarily proceed via uniform downwearing of the proximal surface but rather may proceed via shallow incision by a diffuse network of small stream channels.

Inselbergs

Inselbergs are isolated hills that stand above a surrounding erosional plain (Fig. 13.7). They vary widely

Fig. 13.7 Numerous widely scattered tors and small inselbergs (developed on Mesozoic monzogranite) interrupt the otherwise gently sloping pediment plain which forms the east piedmont of the Granite Mountains in the eastern Mojave Desert

in abundance and size from one pediment to other. Inselbergs as tall as 180 m have been described in the Mojave and Sonoran Deserts (e.g. Sharp 1957, Moss 1977); however, heights of 10 to 50 m appear to be much more common in this region (e.g. Tuan 1959). Kesel (1977) reported that inselbergs of the Sonoran Desert have convex-concave slopes where the basal concavity comprises from 50 to 70% and the crestal convexity from 15 to 50% of the profile. One of the more significant characteristics of inselbergs relative to pediment formation is their distribution relative to the pediment surface. They commonly occur as extensions of major range front interfluves, and their size and number tend to decrease with increasing distance from the mountain front (Cooke and Mason 1973, p. 196).

Regolith

The regolith associated with pediment surfaces is, of course, widely variable. Transported regolith, predominantly wash and channelized flow deposits, also includes colluvial deposits within and adjacent to interfluvial piedmont junctions and thin deposits of desert loess on some inactive surface remnants. Residual regolith is mostly *in situ* weathered bedrock. The maximum thickness of transported regolith on pediments is, as discussed previously, a matter of definition. However, average thicknesses in many proximal to medial areas are limited to a few metres (Gilluly 1937, Tator 1952, Mabbutt 1966, Cooke and Mason 1973). As discussed above, these deposits commonly form thin discontinuous to continuous mantles across the bedrock surface, and surface relief is typically very similar in both form and magnitude to that of the underlying bedrock. Channels as much as 1–2 m deep and 10 m wide are common on both surfaces; however, both are mostly planar and relatively featureless. The thickness of residual regolith of most pediments is difficult to measure but appears to be quite variable (Oberlander 1974, Moss 1977). At least several tens of metres of saprolitic bedrock typically underlie the surfaces of many of the granitic piedmonts in the south-west Basin and Range province (Oberlander 1972, 1974, Moss 1977, Dohrenwend et al. 1986).

Influences on Pediment Development

Lithologic and Structural Influences

It has long been recognized that local lithologic and structural variations strongly influence pediment development (Gilbert 1877, p. 127–8, Davis 1933, Bryan and McCann 1936, Tator 1952, Warnke 1969). On the piedmonts flanking the Henry Mountains of southern Utah, for example, Gilbert (1877, p. 128) observed that in 'sandstones flat-bottomed cañons are excavated, but in the great shale beds broad valleys are opened, and the flood plains of adjacent streams coalesce to form continuous plains. The broadest plains are as a rule carved from the thickest beds of shale'. Gilluly (1937, pp. 341–2) reported a similar relation in the Little Ajo Mountains of southwestern Arizona. 'The development of pediments is apparently conditioned by the lithology of the terrane, the softer and more readily disintegrated rocks forming more extensive pediments than do the harder and more resistant rocks. …. They are commonly better developed on [weathered] granitic rocks and soft sediments than on other rocks.'

Among the most conspicuous expressions of local lithologic/structural influence are the inselbergs, tors and other positive surface irregularities that so commonly interrupt the pediment plain (Tuan 1959, Kesel 1977). These erosional remnants are characteristically associated with rocks that, by virtue of their lithology, texture, and/or structure, are more resistant to weathering and erosion than those underlying the adjacent pediment. Such relations are particularly well expressed on Cima Dome in the eastern Mojave Desert where '. . . major knobs and hillocks rising 50–450 feet above the slopes of [the dome] are underlain by rocks other than the pervasive quartz monzonite. . . . At scattered places, dikes and silicified zones in the quartz monzonite form linear ridges or abrupt risers a few to 25 feet high. . . . On upper parts of the dome are numerous knobs and hillocks as much as 35 feet high composed of exceptionally massive quartz monzonite' (Sharp 1957, p. 277).

On a more regional scale, numerous reports from the literature suggest that although pediments occur on many different rock types in arid regions of the south-west United States, the more extensive pediments may be preferentially developed on less resistant lithologies. In the Mojave and Sonoran Deserts, extensive

Fig. 13.8 Dissected *glacis d'erosion* cut across late Tertiary alluvial and lacustrine strata, Stewart Valley, Nevada. Aerial view is south

bedrock pediments bevel deeply weathered, coarse-crystalline granitic rocks in many areas (Davis 1933, Gilluly 1937, Tuan 1959, Mammerickx 1964, Warnke 1969, Cooke 1970, Oberlander 1972, 1974, Moss 1977, Dohrenwend et al. 1986) (Figs. 13.2, 13.4 and 13.7). In the west-central Great Basin (Gilbert and Reynolds 1973, Dohrenwend 1982a,b), pediments occur primarily on middle to upper Cenozoic terrigenous sedimentary rocks (predominantly partly indurated fluvial and lacustrine deposits, Fig. 13.8) and on less resistant volcanic rocks (predominantly lahars and non-welded ash flow tuffs, Fig. 13.9). Pediments also bevel upper Cenozoic terrigenous sediments along the Ruby and East Humboldt ranges of north-east Nevada (Sharp 1940), in the Furnace Creek area of Death Valley (Denny 1967), in the San Pedro, Sonoita, and Canada del Oro basins of south-east Arizona (Melton 1965, Menges and McFadden 1981), along the lower valley of the Colorado River (Wilshire and Reneau 1992), and along the valley of the Rio Grande (Denny 1967). Applegarth (2004) examined mountain slope morphology to discriminate between pediments and bedrock-covered alluvial slopes, and showed that bedrock pediments were characterised by mountain slopes that had fewer joints, larger clasts and steeper gradients than those backing alluvial slopes. Coupled with his finding that catchments upslope of pediments were smaller than those upslope of alluvial slopes, Applegarth's findings point to sediment supply as a key control on the distribution of pediments.

Fig. 13.9 Dissected remnants of a rock pediment and *glacis d'erosion* cut across Mesozoic granitic rocks, Tertiary volcanic rocks, and Tertiary terrigenous sediments on the back-tilted flank of the Wassuk Range, west-central Nevada. View is north-east towards Lucky Boy Pass (elevation c. 8000 ft) and Corey Peak (elevation 10 520 ft)

In many other areas, however, extensive pediments cut across a variety of more resistant sedimentary, volcanic and metamorphic rocks (Tuan 1959, Mammerickx 1964, Cooke and Reeves 1972, Dohrenwend 1987a). Moreover, regional mapping and analysis of the general distribution of pediments within the tectonically active central and western Great Basin (Dohrenwend et al. 1996a) indicates that, within this region at least, pediment development is not clearly related to the distribution of more easily erodible rock types. Rather, the relative extent of rock types exposed on bedrock pediments accords well with the relative extent of those same rock types within the adjacent ranges (Fig. 13.10). This suggests that local tectonic stability may be the dominant control of pediment development in this tectonically active region.

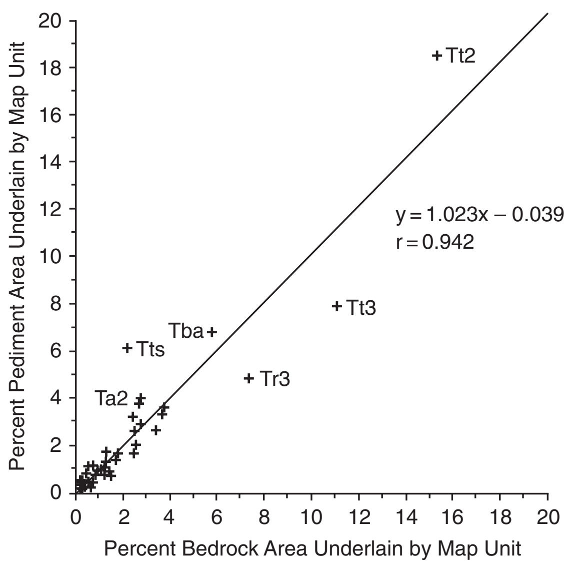

Fig. 13.10 Graph comparing the relative extent of exposed bedrock pediments versus the relative abundance of bedrock in upland areas for each bedrock unit (with total surface exposures greater than 100 km^2) on the geologic map of Nevada (Stewart and Carlson 1977). In essentially all cases, the relative abundance of exposed bedrock pediments underlain by a specific rock type accords well with the relative abundance of that same rock type within the upland areas of the region. Thus, it would appear that pediment development in Nevada has not been strongly influenced by lithologic distribution (from Dohrenwend et al. 1996a). **Ta2** = Tertiary andesite flows and breccias and related rocks of intermediate composition (17–34 Ma); **Tba** = Tertiary basalt and andesite flows (6–17 Ma); **Tr3** = Tertiary rhyolite flows and shallow intrusive rocks (6–17 Ma); **Tts** = Tertiary ash-flow tuffs and tuffaceous sedimentary rocks (6–17 Ma); **Tt2** = Tertiary welded and non-welded silicic ash-flow tuffs (17–34 Ma); **Tt3** = Tertiary welded and non-welded silicic ash-flow tuffs (6–17 Ma)

Tectonic Influences

The bedrock pediments of the south-west United States are located, for the most part, within stable or quasi-stable geomorphic environments where erosional and depositional processes have been approximately balanced for relatively long periods of time. Although pediments have not been systematically mapped or correlated with tectonic environment across the entire region, a clear correspondence between tectonic stability and pediment development is apparent.

General geomorphic analyses of the Basin and Range province document a variety of regional morphometric and geomorphic variations (including the distribution of pediments versus alluvial fans) that are clearly related to variations in the distribution and style of Quaternary faulting (Lustig 1969, p. D68, Bull 1977, Bull and McFadden 1977). For example, Lustig's (1969) morphometric and statistical analysis of Basin and Range morphology concludes that pediments are more abundant in areas characterized by low average values of range relief, height, width, length, and volume, whereas alluvial fans are more abundant where these values are large. This conclusion is generally supported by a regional neotectonic evaluation of the south-west Basin and Range province based on a geomorphic classification of the relative tectonic activity of mountain-fronts (Bull 1977, Bull and McFadden 1977). This study documents several distinct regions of contrasting neotoctonic activity that may be distinguished, in part, by a relative lack or abundance of pediments. Active and moderately active range fronts (with large alluvial fans, few if any pediments, steep hillslopes on all rock types, elongate drainage basins, and low mountain front sinuosities) define a region of pronounced dip-slip faulting between the Sierra Nevada and Death Valley in the south-west Great Basin; and moderately active to slightly active range fronts (with both fans and pediments, relatively equant drainage basins, broader valleys, and moderate mountain front sinuosities) characterize a region of strike-slip faulting in the south-central Mojave Desert. In contrast, inactive range fronts (with broad mountain front pediments, large embayments, and high mountain front sinuosities) distinguish regions of post-middle Miocene tectonic stability in the western and eastern parts of the Mojave Desert (Bull, 1977, Dohrenwend et al. 1991). Thus, the pediments of the Basin and Range province are generally larger and more continuous in areas of greater vertical tectonic stability.

This strong relation between tectonic stability and pediment development is also reflected throughout the south-west Basin and Range province by a general spatial association between extensive pediments and many areas of shallow Miocene detachment faulting (Dohrenwend 1987a). Morphometric comparisons of basin and range morphology (Dohrenwend 1987b), range front morphology (Bull 1977, Bull and McFadden 1977), and the regional distribution of young faults (Dohrenwend et al. 1996b) indicate significantly less post-detachment vertical tectonism in areas with extensive pediments than in other parts of the province. Moreover, a generally low local topographic relief (c. 100–250 m?) has been inferred for areas of shallow deformation along closely spaced listric normal faults in this region (Zoback et al. 1981). This combination of relative tectonic stability and relatively low initial tectonic relief has permitted the formation of numerous pediments that typically encircle their associated uplands and extend downslope as much as 5–10 km from the mountain front.

Regional mapping of pediments throughout the state of Nevada (which includes many of the more tectonically active regions of the Basin and Range province) indicates that pediment development there also has been controlled primarily by spatial and temporal variations in late Cenozoic tectonic activity (Dohrenwend et al. 1996a). Pediment areas with exposed bedrock or thin alluvial cover occupy approximately 15% of the intermontane piedmonts and basins of the region. The more extensive and continuous of these pediments are located in the south-central and western parts of the state in areas of relatively shallow Miocene detachment faulting that apparently have undergone relatively little post-Miocene vertical tectonic movement (see above). Typically, these areas are characterized by a relatively subdued topography consisting of low narrow ranges, broad piedmonts, and small shallow basins. Conversely, pediments are smaller and less abundant within the more tectonically active areas of the state. Included within these tectonically active areas are the longest, most continuous, and most active dip-slip fault zones; the longest and widest ranges; and the largest, deepest, and most continuous Cenozoic basins in the Great Basin.

Within tectonically active areas, pediments are largely confined to local settings of relative geomorphic stability. Exposed bedrock pediments typically occupy proximal piedmont areas immediately adjacent to the range front; however, in some cases pediments may extend from range front to basin axis. Particularly favourable settings include range embayments and both low broad passes within and narrow gaps between ranges. Such settings are especially well suited for pediment development if they are also situated on the backtilted flanks of large asymmetric range blocks or gently upwarped structural highs. Smaller pediments occur on the tilted flanks of small blocks along major strike-slip faults, on upfaulted piedmont segments, and within small embayments adjacent to active fault-bounded range fronts (Dohrenwend 1982c).

Although all of these diverse geomorphic settings are related by a common condition of relative long-term stability, they are quite different in many other respects. Hence, the pediments formed within each of these local geomorphic environments are morphologically distinct. Comparison of the pediments on the east and west flanks of the Ruby–East Humboldt Range in north-eastern Nevada illustrates this relation between pediment morphology and local neotectonic–geomorphic environment (Sharp 1940). The east flank of this asymmetrically west-tilted range is bounded by a major range front fault system. Thus the drainage basins and piedmonts of the east flank are smaller and steeper than those of the west flank, and Pleistocene glaciation was less extensive on the east flank than on the west flank. Consequently, the large pediments of the west flank extend across the backtilted flank of the range, whereas the smaller pediments of the east flank are confined to embayments upslope from the range-bounding fault. Seven surfaces (pediments and partial pediments) occur on the west flank of the range, and these pediments are cut mostly on soft basin deposits. In contrast, the pediments of the east flank are, with one exception, cut entirely on the hard rocks of the mountain block. 'Remnants of pediments on the west flank extend as far as 5 miles from the foot of the mountain slope, and the undissected surfaces were probably even more extensive. The east flank pediments are narrower, at least as exposed, and seldom extend more than 1.5 miles from the mountains. The west flank pediments have gentler gradients and are more nearly smooth than those of the east side. The pediments of the west side are mantled by a comparatively uniform cover of stream gravel; those of the east side have only local patches of stream gravel. . .' (Sharp 1940, pp. 362–3).

Climatic Influences

The influence of climate on pediment development has not been systematically studied and, therefore, is not very well documented or understood. Pediments occur within the broad range of climates from subpolar to humid tropical. They are particularly abundant and well developed, however, in arid/semi-arid and tropical wet–dry regimes. Indeed pediments are so characteristic of these two rather disparate regimes that they represent somewhat of an embarrassment to advocates of climatic geomorphology (Chorley et al. 1984, p. 486). Descriptive reports in the literature indicate that the pediments of most climatic realms are generally similar in form and landscape position. For example, 'broad, gently sloping [erosional] surfaces which extend from the base of hillslopes, and which in some cases pass beneath alluvial accumulations and in others terminate at a break in slope leading down to the river channel or floodplain, undeniably exist in the tropics' (Thomas 1974, p. 218). These tropical pediments are characterized by generally smooth, concave-up longitudinal profiles with declivities ranging between 1° and 9°. Also they are most commonly interpreted to function as surfaces of transport between hillslope and riverine plain. These similarities of form, position, and function notwithstanding, it is not at all clear that the processes of mass transport which maintain these surfaces are at all similar from one climatic regime to another. However, it is apparent that these surfaces are for the most part typically associated with either local or regional settings of long-term geomorphic stability.

Because long-term geomorphic stability appears to be a general requirement for pediment formation, it is very likely that the development of many pediments, particularly larger ones, may transgress periods of unidirectional climate change and that such changes may profoundly affect pediment development. For example, several lines of evidence indicate that the granitic pediments of the Mojave and Sonoran Deserts were fully developed by late Miocene time and since that time have been modified primarily by partial stripping of a thick saprolitic regolith (Oberlander 1972, 1974, Moss 1977). According to Oberlander, the late Miocene Mojave–Sonoran region was probably an open woodland interspersed with grassy plains that extended across extensive cut and fill surfaces of low relief surmounted by steep-sided hills. Pediment formation within this landscape appears to have involved an approximate balance between regolith erosion (probably by slope wash, rill wash and channelized flow processes) and regolith renewal (by chemical breakdown of granitic rock along a subsurface weathering front). This balance was apparently upset by increasing aridity resulting in a loss of vegetative cover which triggered regolith stripping and concomitant formation of the relatively smooth pediment surfaces that characterize the modern landscape. Evidence supporting this general scenario includes saprolite remnants preserved beneath late Tertiary lava flows, the elevated positions of these lava-capped palaeosurfaces above present erosion surfaces, continuity between the relic saprolite and boulder mantles on hillslopes, and general grusification as much as 40 m below present pediment surfaces. Interestingly enough, it is generally agreed that the pediments of many tropical and subtropical areas have also developed 'across pre-weathered materials or have been significantly modified, at least, by erosional stripping of deeply weathered materials (Mabbutt 1966, Twidale 1967, Thomas 1974, p. 218–20).

Pediment Development and Time

It is generally perceived that extensive pediments are a characteristic feature of old landscapes (Lawson 1915, Bryan 1922, Davis 1933, Howard 1942) and that their formation requires a condition of general landscapes stability wherein erosional and depositional processes are approximately balanced over long periods of time (cyclic time of Schumm and Lichty 1965). There is little doubt that the more extensive pediments of the Basin and Range province are, at least in part, relics of considerable age. Throughout this region, local burial of piedmonts and pediments by late Cenozoic lava flows provides convincing evidence of their long-term morphologic stability, even in regions of neotectonic activity (Figs. 13.11 and 13.12, Table 13.3). Moreover,

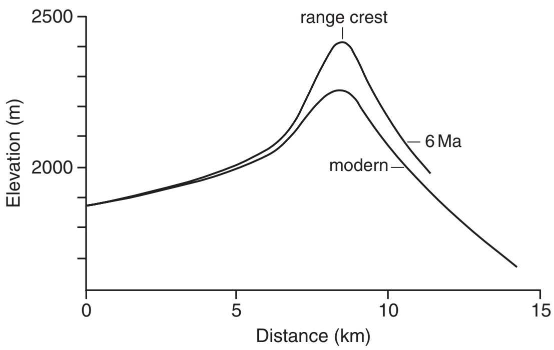

Fig. 13.11 Longitudinal profiles comparing modern and late Miocene erosional surfaces on the west flank of the Reveille Range. Since latest Miocene time, rates of incision and downwearing have ranged between 20 and 35 m per million years in crestal and upper flank areas and between 5 and 20 m per million years in lower flank and proximal piedmont areas. Medial and distal piedmont areas have undergone less than 5 m of net downwearing

ubiquitous deep weathering beneath extensive pediments in the Mojave and Sonoran Deserts indicates at least pre-Quaternary ages for the original surfaces of these pediments (Oberlander 1972, 1974, Moss 1977).

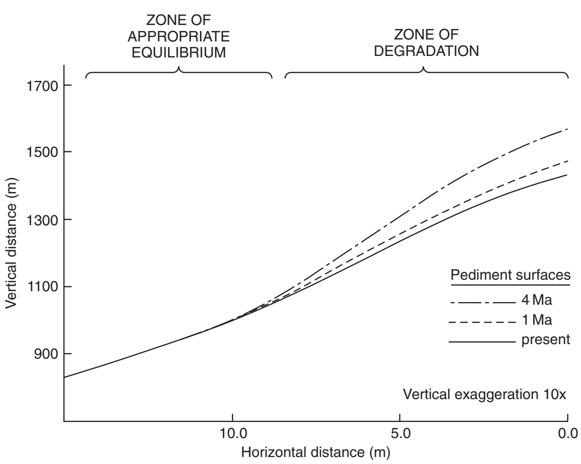

Fig. 13.12 Empirical model of pediment dome evolution in the eastern Mojave Desert based on a compilation of elevation differences between active pediment surfaces and relict surfaces capped by dated lava flows. The 1 Ma and 4 Ma surfaces were reconstructed using a smoothed plot of average downwearing rates versus distance from dome summits. Upper flank areas have been eroded, midflank areas have remained in a state of approximate equilibrium, and lower flank areas (not shown) have probably aggraded. Horizontal distance is measured from the dome summit (from Dohrenwend et al. 1986)

Average Rates of Downwearing and Backwearing

One reliable approach for estimating average erosion rates in the Basin and Range province involves the documentation and analysis of erosion surfaces capped by Tertiary and Quaternary basaltic lava flows. Average downwearing rates can be estimated by measuring the average vertical distance between an erosional palaeosurface capped by a lava flow and the adjacent modern erosion surface, then dividing this height difference by the K/Ar age of the lava flow. Using this approach, average downwearing rates have been estimated for several widely separated upland and piedmont areas within the Great Basin and Mojave Desert. For upland areas, these estimates range between 8 and 47 m per million years for periods of 0.85–10.8 m.y. (Table 13.3a); whereas for proximal piedmont areas, the estimates range from less than 2–13 m per million years for periods of 1.08–8.9 m.y. (Table 13.3b). During similar time periods, medial piedmont areas have remained largely unchanged (Table 13.3c). These data document a general evolutionary scenario of upland downwearing and proximal piedmont regrading which has been regulated in part by the long-term stability of the medial piedmont which serves as a local base level for the upper part of the pediment association (Figs. 13.11 and 13.12).

More recently, the use of ^{10}Be and ^{26}Al cosmogenic nuclides has provided another method for estimating rates of downwearing over periods of 10^3–10^5 years. Nichols et al. (2002, 2005), working in the Mojave Desert, have obtained rates of 38 (^{10}Be) and 40 (^{26}Al) m per million years for the Chemehuevi Mountain, and 31 and 33 m per million years, respectively, for Granite Mountain. Proximal pediment downwearing of the Chemehuevi Mountain pediment was estimated at 10 and 21 m per million years. These rates are in accord with those obtained from the dating of lava flows.

Average rates of backwearing or slope retreat can be estimated in a similar manner by measuring the average horizontal distance between the maximum possible original extent of a lava flow which caps an erosional palaeosurface and the present eroded margin of that flow, then dividing this horizontal distance by the K/Ar age of the lava flow. Estimated rates of slope retreat determined by this method range between 37 and 365 m per million years (Tables 13.3a,b). Of course, these estimates of average backwearing should be used

Table 13.3 (a) Erosion rates in upland areas of the south-west Basin and Range province inferred from comparisons between active and basalt-capped relic erosion surfaces

Location	Minimum age of relic surface (Ma)	Maximum average downwearing rate (m per m.y.)	Maximum average rate of slope retreat (m per m.y.)	Reference
Fry Mountains, southern Mojave Desert	8.90 ± 0.9	8	–	Oberlander (1972)
Cima volcanic field, eastern Mojave Desert	4.48 ± 0.15	11	290	Dohrenwend et al. (1986) Turrin et al. (1985)
Cima volcanic field, eastern Mojave Desert	3.85 ± 0.12	29	365	Dohrenwend et al. (1986) Turrin et al. (1985)
Cima volcanic field, eastern Mojave Desert	0.85 ± 0.05	30	320	Dohrenwend et al. (1986) Turrin et al. (1985)
White Mountains, south-west Great Basin	10.8	24	–	Marchand (1971)
Buckboard Mesa, southern Great Basin	2.82	47	–	Carr (1984)
Columbus Salt Marsh, central Great Basin	3.04 ± 0.2	21	37	Dohrenwend (unpublished data)
Reveille Range, central Great Basin	5.70 ± 0.20	34	230	Ekren et al. (1973)
Reveille Range, central Great Basin	3.79 ± 0.34	20	200	Dohrenwend et al. (1985, unpublished data)

Table 13.3 (b) Erosion rates in proximal piedmont areas of the south-west Basin and Range province inferred from comparisons between active and basalt-capped relic erosion surfaces

Location	Minimum age of relict surface (Ma)	Maximum average downwearing rate (m per m.y.)	Maximum average rate of slope retreat (m per m.y.)	Reference
Fry Mountains, southern Mojave Desert	8.90 ± 0.90	<2	–	Oberlander (1972)
Cima volcanic field, eastern Mojave Desert	3.64 ± 0.16	11	85	Dohrenwend et al. (1986)
Reveille Range, central Great Basin	5.76 ± 0.32	8	230	Dohrenwend et al. (1985, unpublished data)
Reveille Range, central Great Basin	5.58 ± 0.30	2.5	190	Dohrenwend et al. (1985, unpublished data)
Lunar Crater volcanic field, central Great Basin	2.93 ± 0.30	9	–	Turrin and Dohrenwend (1984, unpublished data)
Lunar Crater volcanic field, central Great Basin	1.08 ± 0.14	13	–	Turrin and Dohrenwend (1984, unpublished data)
Quinn Canyon Range, central Great Basin	8.5 ± 0.7	15	–	Dohrenwend (unpublished data)
Columbus Salt Marsh, central Great Basin	3.04 ± 0.2	13	50	Dohrenwend (unpublished data)

Table 13.3 (c) Late Cenozoic basaltic volcanic fields on piedmonts in the south-west Basin and Range province where net vertical erosion/deposition has been low (generally less than 5 m in most medial and/or distal piedmont areas) since lava flow emplacement

Area	Latitude	Longitude	Age range (Ma)	Reference
Buffalo Valley, northern Great Basin	40.35°N	117.3°W	0.9–3.0	Dohrenwend (1990a)
Reveille Range, central Great Basin	38.1°N	116.2°W	3.8–5.8	Dohrenwend et al. (1985) Dohrenwend (1990b)
Lunar Crater volcanic field, central Great Basin	38.25°N	116.05°W	c. 0.1–2.9	Turrin and Dohrenwend (1984) Dohrenwend (1990c)
Quinn Canyon Range, central Great Basin	37.9°N	115.95°W	7.8–9.2	Dohrenwend (unpublished data)
Clayton Valley, western Great Basin	37.8°N	117.65°W	0.2–0.5	Dohrenwend (1990d)
Columbus Salt Marsh, central Great Basin	38.1°N	118.05°W	2.8–3.2	Dohrenwend (unpublished data)
Crater Flat, southern Great Basin	37.1°N	116.5°W	c. 0.1–3.7	Carr (1984), Turrin et al. (1991)
Big Pine, south-west Great Basin	37.05°N	118.25°W	<0.1–1.2	Gillespie (1990)
Saline Valley, south-west Great Basin	36.85°N	117.7°W	0.5–0.7	Dohrenwend (unpublished data)
Cima volcanic field, eastern Mojave Desert	35.25°N	115.75°W	c. 0.1–1.1	Dohrenwend et al. (1986) Dohrenwend (1990e)

only as general approximations. The armouring effect of blocky basalt talus inhibits the backwearing of basalt capped hillslopes; moreover, substantial uncertainties are inherent in determining the original extent of most lava flows. However, these average backwearing rates are comparable with the 0.1 to 1-km-permillion-year rates of range front retreat and pediment formation that have been previously estimated or assumed for diverse areas of the Basin and Range province (Wallace 1978, Mayer et al. 1981, Menges and McFadden 1981, Saunders and Young 1983).

If one assumes that the estimated rates of downwearing and backwearing compiled in Table 13.3 are, in fact, representative of erosion rates in arid regions of the south-western United States, then a typical upland in this region has probably undergone from as little as 50 m to as much as 250 m of downwearing and from 0.2 to 2 km of backwearing since late-Miocene time. During this period, the medial areas of the piedmont fringing this typical upland would have experienced little morphologic change. Hence, it is very likely that many of the larger pediments of the Mojave and Sonoran Deserts (which typically are at least several kilometres wide) have been forming since at least late Miocene time. This pattern of landscape evolution is particularly well documented on the slopes and piedmonts of the Reveille Range in south-central Nevada (Fig. 13.11). Since latest Miocene time, denudation rates along the crest and upper flanks of the Reveille Range have ranged between 20 and 35 m per million years. Downwearing of lower flank and upper-piedmont areas has ranged between 5 and 20 m per million years; and most middle and lower piedmont areas have remained in a state of approximate erosional equilibrium (Dohrenwend et al. 1985, 1986).

Direct Evidence of Pediment Age

The local burial of Tertiary pediments remnants by late Miocene and Pliocene basalt flows provides convincing evidence for a long history of pediment evolution (Oberlander 1972, 1974, Dohrenwend et al. 1985, 1986). North of the San Bernardino Mountains in the south-western Mojave Desert, late Miocene lava flows cover pediment remnants cut across saprolitic weathering profiles. These remnants now stand as much as 60 m above adjacent pediment surfaces (Oberlander 1972). In the vicinity of Cima dome in the north-eastern Mojave Desert, Pliocene lava flows bury remnants of ancient pediments cut across less intensely weathered materials (Fig. 13.13); and flows younger than 1.0 Ma bury pediment remnants locally capped by soils similar to those developed

Fig. 13.13 Early Pliocene basaltic lava flows of the Cima volcanic field preserve remnants of a large Miocene pediment dome cut across deeply weathered monzogranite and Tertiary terrigenous sedimentary rocks. Aerial view north-east towards the 80 to 120-m-high erosional scarp which bounds the western edge of the Pliocene flows. A continually evolving pediment surface extends downslope from the base of this deeply embayed escarpment. This modern pediment is pervasively dissected by a finely textured, subparallel network of shallow (<3 m deep) washes

Fig. 13.14 Quaternary basaltic lava flows (c. 0.1–1.0 Ma) of the Cima volcanic field veneer remnants of a continually evolving pediment surface formed on Cretaceous monzogranite and middle Tertiary terrigenous sediments. Deep incision of a c. 1.0 Ma flow complex (*left* foreground) reflects the delicate erosional–depositional balance of this active pediment surface. Aerial view is north-west

on adjacent Quaternary surfaces (Fig. 13.14). In this area, differences in height between buried surface remnants and adjacent active pediment surfaces are systematically related to the ages of the overlying lava flows (i.e. at equal distances downslope from the crest of the pediment dome, the older the buried remnant the greater its height above the adjacent active surface). Moreover, the ubiquitous presence of deep weathering profiles underlying other extensive granitic pediments in the Mojave and Sonoran Deserts suggests at least pre-Quaternary ages for the original pediment surfaces (Oberlander 1972, 1974, Moss 1977). These relations indicate that, since late Miocene time, pediment surfaces cut across deeply weathered granitic rocks have evolved more or less continuously by progressive stripping of a thick late Tertiary weathering mantle (Oberlander 1972, 1974, Moss 1977, Dohrenwend et al. 1986). Hence, these pediments are, at least in part, ancient forms of late Miocene age.

Pediment Processes

As Cooke and Mason (1973, p. 203) aptly pointed out, attempts to deduce the formation of an existing landform from observation of the processes presently operating on that landform are often unsuccessful because this approach commonly confuses cause and effect. Mere spatial association does not establish a cause and effect relation between form and process. Indeed, such relations are particularly unlikely in the case of long-lived landforms (such as pediments) that have evolved through complex histories of tectonic and climate change. The various processes operating on such landforms and the relative effectiveness of these processes undoubtedly has changed dramatically through time. Even when a cause and effect relation between form and process does exist, such a relation may well be unidirectional (i.e. the form determines the process, but the process did not produce the form). For example, sheetflood erosion cannot produce a planar pediment surface because a planar surface is necessary for sheetflooding to occur (Paige 1912, Cooke and Warren 1973, p. 203). Similarly, weathering and subsequent removal of the weathered debris are unlikely to produce a pediment surface but will act to maintain a pre-existing surface if these processes are uniformly applied (Lustig 1969, p. D67). Thus, the assumption that the processes presently operating to maintain or modify pediment form are the same as those responsible for pediment formation is not justified.

Processes on Pediment Surfaces

In the desert regions of western North America, pediments serve as integral parts of the piedmont, and the processes operating on pediment surfaces are much the same as those operating on other parts of the piedmont plain (Dohrenwend 1987a). Processes which act to maintain or modify pediment surfaces include surface and subsurface weathering, unchannelized flow (sheet flow, sheetflooding, rill wash), and channelized flow (gully wash, debris flow, and stream flow). In most cases, the relatively diffuse processes of weathering and unchannelized flow will tend to be more uniformly distributed in space than the more concentrated processes of channelized flow. Where the combined effect of these processes acts uniformly over the entire pediment surface, the surface will be maintained as an active pediment. Where the combined effect is not uniformly applied, however, incision will occur most likely in response to decreases in stream sediment load (Fig. 13.6) and/or increases in stream discharge. Surfaces abandoned by fluvial action as a result of incision form isolated islands of relative stability where the processes of subsurface and surface weathering (particularly salt weathering), soil formation, desert loess accumulation, and stone pavement development are locally dominant.

Under conditions of tectonic and base level stability, the piedmont surface is divisible into three general process zones: an upper zone of erosion, an intermediate zone of transportation, and a lower zone of aggradation (Johnson 1932a,b, Cooke and Warren 1973, p. 197, Dohrenwend et al. 1986). These zones shift upslope and downslope across the piedmont in response to changes in the overall upland/piedmont drainage system, and the boundaries between zones are both gradational and complex. Their precise position at any given time is determined by the relative rates of debris supply and removal across the piedmont plain. If the rate of debris supply to the piedmont increases or the rate of removal decreases, these zones tend to migrate upslope; conversely, if the rate of supply decreases or the rate of removal increases, they tend to migrate downslope (Cooke and Mason 1973). On most piedmonts in arid and semi-arid environments, the rates of debris supply and removal are largely determined by channelized flow processes; thus the relative extent and position of each general process zone reflects the average location, over time, of the threshold of critical power for each component of the piedmont drainage system. On average, stream power exceeds critical power in the zone of erosion, approximately equals critical power in the zone of transportation, and falls short of critical power in the zone of aggradation (Bull 1979). This general scenario is supported by (a) regional patterns of proximal piedmont dissection in the Mojave and Sonoran Deserts; (b) long-term trends of piedmont and pediment erosion documented by spatial relations between volcanic landforms and active piedmont surfaces (Dohrenwend et al. 1985, 1986, Dohrenwend 1990a,c); and (c) morpho-stratigraphic relations in the Apple Valley area of the western Mojave Desert (Cooke and Mason 1973).

These conceptual divisions of pediment surfaces are supported by rates of denudation obtained from ^{10}Be and ^{26}Al measurements. On the Chemehuevi Mountain pediment in the Mojave Desert Nichols et al. (2005) obtained denudation rates of 22 m per million years at a distance of 1 km from the mountain front compared to 8 m per million years 3 km from the mountain front. Nuclide data indicate that the lower part of the pediment is an area of sediment transport and deposition, with the interfluve areas effectively stable.

In the zone of erosion, channelized flow processes typically dominate and commonly result in shallow to moderate dissection of the proximal piedmont. However, surface regrading may also proceed without significant incision via the operation of unchannelized flow and lateral shifting of channelized flow. In the zone of transportation, stream power and critical power are essentially balanced over the long term such that net surface erosion is nearly undetectable. Surface processes in this zone are probably dominated by lateral shifting of channelized flow in shallow anastomosing gullies and washes (Rahn 1967, Cooke and Mason 1973, Dohrenwend et al. 1986). Rates of sediment transport obtained from ^{10}Be and ^{26}Al data are found to be highest in confined bedrock channels (10s of metres per year), intermedaite (metres per year) in incised alluvium, but only decimetres per year in highly permeable active alluvial surfaces (Nichols et al., 2005).

Subsurface Weathering

Subsurface weathering processes have strongly influenced pediment development in many regions, particularly in humid tropical, tropical wet–dry, and semi-arid environments (Ruxton 1958, Mabbutt 1966, Thomas 1974, p. 218–26, Oberlander 1989). Such processes are critical to the formation or modification of many pediments, particularly those developed on crystalline bedrock (Oberlander 1974, 1989, Moss 1977). Clearly, weathered and disaggregated bedrock can be much more readily loosened, entrained, and eroded by wash and stream action than unweathered bedrock.

Indeed, the apparent predilection of pediments for areas underlain by granitic rocks is due, in large measure, to the susceptibility of these rocks to subsurface weathering and to the particular mechanical characteristics of the residual grus. This well-sorted, sand-sized, non-cohesive material forms non-resistant channel banks that are highly susceptible to erosion by laterally shifting channelized flow. Hence, granitic pediments possess a highly effective mechanism of self-regulation which tends to suppress fluvial incision. Wherever the threshold of critical power is exceeded by stream flow, erosion of the channel banks rapidly increases sediment load and raises the critical power threshold to the level of available stream power. The profound contrast in surface morphology between granitic and metamorphic piedmonts in the Mojave Desert serves as a graphic illustration of this phenomenon. Granitic piedmonts are, in most cases, essentially undissected and relatively featureless plains of active transport, whereas metamorphic piedmonts are more commonly dissected into intricately nested surfaces of diverse age.

Controlled largely by the availability of moisture, the rates and distribution of subsurface weathering processes are closely related to local geomorphic setting and to shallow subsurface form and structure. Conditions favouring moisture accumulation and retention will typically intensify weathering processes. On arid and semi-arid piedmonts, moisture accumulation and retention commonly occurs (a) beneath footslopes within the piedmont junction, (b) along the alluvium–colluvium and bedrock interface, (c) along buried bedrock channels, and (d) in areas of intersecting bedrock fractures. Hence, the surface described by the subsurface weathering front is typically highly irregular, and stripping of the residual regolith in areas of proximal dissection commonly reveals complex and intricate etching of the bedrock surface.

Piedmont Modification by Subsurface Weathering and Regrading

Subsurface weathering and the subsequent erosional modification of piedmonts or other surfaces of low relief has played a central role in the expansion and/or regrading of many pediments (Mabbutt 1966, Oberlander 1972, 1974, 1989, Moss 1977, Dohrenwend et al. 1986). In the Flinders Ranges of South Australia, for example, 'the abrupt change in slope in the piedmont zone (junction) is caused principally . . . by differential weathering at the foot of the scarp. . . . The bedrock 1.5 km from the scarp is weathered only to a moderate degree. . . . Close under the scarp, however, similar strata are much more intensely altered. Kaolin is abundant, bedding only vaguely discernable, and the rock has lost much of its strength. . . . Runoff from the often bare hillslopes percolates into the rock strata where it reaches the plain, causing [this] intense weathering' (Twidale 1967, p. 113). Also in central Australia, '. . . grading on . . . granitic and schist pediments has largely acted through the mantles, whereby the smoothness of the depositional profiles has been transmitted to suballuvial and part-subaerial pediments. . . . On granitic pediments . . . suballuvial notching and levelling proceed through weathering in the moist subsurface of the mantle. On schist pediments control of levelling by the mantle is less direct in that its upper surface is the plane of activity of ground-level sapping and of erosion by rainwash and sheetflow. . . . [These processes] are most active near the hill foot, where mantling and stripping alternate more frequently; the lower parts of both schist and granitic pediments . . . appear to be largely and more permanently suballuvial . . .' (Mabbutt 1966, pp. 90–1). Although subsurface weathering processes have strongly influenced pediment development in many areas and profoundly modified pediment surfaces in many others, it would appear unlikely that these processes actually 'control' pediment development, at least in arid and semi-arid environments. The preservation of tens of metres of intensely weathered bedrock beneath active granitic pediments demonstrates that these landforms are transport-limited (not weathering- or detachment-limited). The present form of the active surface is the product of fluvial erosion and transport; it is not closely related to either the position or the form of the subsurface weathering front. Moreover, even though stripping of the proximal areas of many pediments has exposed the intricately etched (weathering-limited) surfaces of formerly buried weathering fronts, deeply weathered regolith remains beneath adjacent mantled surfaces of fluvial transport. Thus even this pediment regrading is primarily controlled by fluvial processes; only the detailed form of the proximal pediment has been significantly influenced by the position of the weathering front.

Models of Pediment Formation

Within the general context of the pediment association (i.e. degrading upland, aggrading alluvial plain, and intervening zone of transition and transport), pediment development can be usefully perceived as the necessary result of upland degradation (Lustig 1969, p. D67). As the pediment association evolves and mass is transferred from the upland to the alluvial plain, the diminishing upland is replaced by an expanding surface of transportation, the pediment. Hence, most models of pediment development implicitly link upland degradation and pediment development; and the major differences between these models centre primarily on the issues of mountain front (or piedmont junction) retreat and upland degradation. Accordingly, the various models that have been presented in the literature can be grouped according to the dominant style of upland degradation proposed. These styles are

(a) range front retreat where channelized fluvial processes (mainly lateral planation) predominate (Gilbert 1877, p. 127–9, Paige 1912, Blackwelder 1931, Johnson 1931, 1932a,b, Field 1935, Howard 1942);
(b) range front retreat where diffuse hillslope and piedmont processes predominate (Lawson 1915, Rich 1935, Kesel 1977);
(c) range front retreat assisted by valley development where the relative dominance of diffuse hillslope and piedmont processes versus channelized flow varies according to general geomorphic environment (Bryan 1922, 1936, Gilluly 1937, Sharp 1940);
(d) range degradation dominated by drainage basin development (valley deepening and widening, embayment formation and enlargement) (Lustig 1969, p. D67, Wallace 1978, Bull 1979, 1984, Parsons and Abrahams 1984).

This trend in dominant styles of upland degradation clearly shows a progressive maturation of Gilbert's original (1877) model. Thus, as Parsons and Abrahams (1984) point out, these models are not mutually exclusive; indeed, they are more complementary than contradictory.

Range Front Retreat by Lateral Planation

Models of lateral planation emphasize the importance of channelized fluvial processes. Erosion of the mountain mass and retreat of the mountain front is considered to be effected primarily by the lateral shifting of debris-laden streams issuing from within the mountains. The diffuse action of weathering, slopewash, and rill wash on hillslopes is acknowledged but is not considered essential to pediment formation. The resulting pediment surface is continually regraded by the lateral shifting of these streams across the piedmont.

> Whenever the load [of a stream] reduces the downward corrasion to little or nothing, lateral corrasion becomes relatively and actually of importance. . . . The process of carving away the rock so as to produce an even surface, and at the same time covering it with an alluvial deposit, is the process of planation. . . . The streams . . . accomplish their work by a continual shifting of their channels; and where the plains are best developed they employ another method of shifting. . . . The supply of detritus derived from erosion . . . is not entirely constant. . . . It results from this irregularity that the channels are sometimes choked by debris and . . . turned aside to seek new courses upon the general plain. . . . Where a series of streams emerge from the adjacent mountain gorges upon a common plain, their shiftings bring about frequent unions and separations, and produce a variety of combinations (Gilbert 1877, pp. 127–9).
>
> . . . the essence of the theory lies in the broader conception that rock planes of arid regions are the product of stream erosion rather than in any particular belief as to the relative proportions of lateral and vertical corrasion. . . . Every stream is, in all its parts, engaged in the three processes of (a) vertical downcutting or degrading, (b) upbuilding or aggrading, and (c) lateral cutting or planation. . . . The gathering ground of streams in the mountains, where greater precipitation occurs, will normally be the region where vertical cutting is at its maximum. . . . Far out from the mountain mass conditions are reversed. Each stream must distribute its water and its load over an everwidening sector of country. The water disappears, whether by evaporation or by sinking into the accumulating alluvium. Aggradation is at its maximum. . . . Between . . . [these regions] . . . there must be a belt or zone where the streams are essentially at grade. . . . Thus from the center outward are (1) the zone of degradation, (2) the zone of lateral corrasion, and (3) the zone of aggradation. Heavily laden streams issuing from the mountainous zone of degradation are from time to time deflected against the mountain front. This action combined with the removal of peripheral portions of the interstream divides by lateral corrasion just within the valley mouths, insures a gradual recession of the face or faces of the range. Such recession will be aided by weathering as well as by rain and rill

> wash; but it must occur even where these processes are of negligible importance (Johnson 1932a, pp. 657–8).
>
> . . . lateral planation may be carried on by streams of all sizes, by distributaries as well as by tributaries, by individual channels of a braided stream, and by sheetfloods (Howard 1942, p. 107).

Range Front Retreat by Diffuse Hillslope and Piedmont Processes

A second group of models identifies diffuse hillslope processes (mainly weathering, slopewash, and rill wash) coupled with removal of the resultant debris from the piedmont junction by sheetwash and rill wash as the dominant mechanism of mountain front retreat and pediment extension. Lateral planation by mountain streams is acknowledged as locally significant but is not considered to be an essential process. Sheetflooding is considered to be the dominant process of pediment surface regrading.

> If we examine a typical region in an old age stage of the arid cycle, where scattered residuals of former mountain masses stand on broad rock pediments. . . . Obviously the mountains are wasting away by weathering, and just as obviously the weathered products must be carried away over a gradient steep enough to permit them to be moved. The rock, though weathered, cannot be removed below the line of this gradient. Consequently as the mountains waste away, a sloping rock plain, representing the lower limit of wasting, must encroach upon them from all sides. . . . The gradient of this rock plain must be that necessary for the removal of the waste material – no more and no less – therefore, the rock will be covered by only a thin and discontinuous veneer of waste in transit (Rich 1935, p. 1020).
>
> The prevailing conditions of stream load force the desert streams to corrade laterally, as Johnson has pointed out, and in many places such corrasion contributes actively to the formation of pediments, especially along the sides of canyons debouching from the desert mountains. Nevertheless, lateral corrasion need be only a contributing, and not an essential factor in the formation of pediments, and a very minor factor in the retreat of mountain fronts (Rich 1935, p. 1021).

Range Front Retreat Assisted by Valley Development

A third group of models builds on the complementary nature of the previous models and accommodates their differences in emphasis by recognizing that the relative importance of weathering, unchannelized flow, or channelized flow processes varies according to specific geomorphic setting.

> The conclusions reached as to the formation of pediments under the various geological, topographic, and climatic conditions of the Ruby–East Humboldt region are as follows:
>
> (1) Pediments are formed by lateral planation, weathering, rill wash, and rain wash. The relative efficacy of these various processes is different under geologic, topographic, and climatic conditions.
> (2) Lateral planation is most effective along large permanent streams and in areas of soft rocks.
> (3) Weathering, rill wash, and rain wash are most effective in areas of ephemeral streams, hard rocks, and a low mountain mass.
> (4) All variations from pediments cut entirely by lateral planation to those formed entirely by the other processes are theoretically possible, although in this area the end members of the series were not observed and perhaps do not actually exist in nature (Sharp 1940, p. 368).

Range Degradation by Drainage Basin Development

The first three groups of models focus on the evolutionary retreat of the mountain front. In contrast, the models of this group address the broader issue of the overall degradation of the mountain mass. Their emphasis is on the evolutionary development of upland valleys through non-uniform erosion of the mountain mass by the concentrated action of channelized streamflow. It is significant to note that these models do not require parallel retreat of the mountain front as the primary mode of pediment expansion. These more comprehensive models emphasize the dominant role of fluvial erosion in the degradational evolution of the entire pediment association, and they illustrate the increasing sophistication of geomorphic theory as applied to the analysis of landscape evolution.

> The many discussions of pediment surfaces have focused on the wrong landform. . . . There is no question that the processes of subaerial and suballuvial weathering occur on pediment today, nor that fluvial erosion also occurs. A pediment must exist prior to the onset of these processes, however, and in this sense the origin of the pediments resides in the adjacent mountain mass and its reduction through time. . . . Given stability for a sufficient period of time, the consequences of mountain re-

duction must inevitably include the production of a pediment, whether in arid or nonarid regions. The nature of the surface produced may vary, and it may be mantled by, or free of, alluvium. However, it simply represents an area that was formerly occupied by a mountain. . . . The only real "pediment problem" is how the reduction or elimination of the mountain mass occurs (Lustig 1969, p. D67).

. . . it is obvious that the pediment grows most rapidly along the major streams. In every indentation in the mountain front and in places where streams emerge from the canyons onto the plains the rate of formation of the pediment is rapid, and consequently extensions of the pediment into the mountains are common. . . . The erosion of the mountains at the headwaters of many streams is much faster than in the lower portions of the same streams. . . . Consequently the headwater slopes may recede more rapidly than the side walls of the valleys (Bryan 1922, pp. 57–8).

The rates of mountain front retreat are basically unknown, but by any reasonable assessment they are slow in relation to rates of processes that are operative in drainage basins. This is clearly true because the headwater region of any given drainage basin also consists of steep walls that are virtually identical to those of the mountain front in interfluvial areas. In these headwater regions the same processes of weathering and removal of debris occur. Hence, the rates of retreat of the bounding walls in the headwaters of drainage basins must be at least as great as the rate of retreat of the mountain front in interfluvial areas. Also, however, the drainage basins represent the only parts of any mountain range that are subjected to concentration of flow and to its erosional effects, and these basins must therefore be the principal loci of mountain mass reduction (Lustig 1969, p. D67).

During the valley downcutting that occurs after mountain-front uplift, the width of the valley floor will approximate stream width at high discharges. Valley-floor width decreases upstream from the front because of the decrease in the size of the contributing watershed. . . . With the passage of geologic time, the stream will widen its valley as it approaches the threshold of critical power. As lateral cutting becomes progressively more important, the stream will not spread over the entire valley floor during high discharges. The approximation of a threshold condition migrates gradually upstream as the upstream reaches downcut. . . . The configurations of the . . . pediment embayments . . . are functions of the rates of lateral cutting and/or hillslope retreat along the stream and the time elapsed since lateral erosion became predominant at various points along the valley. More than a million years may be needed to form pediment embayments (Bull 1979, pp. 459–60). . . . it seems likely that, where a significant proportion of the piedmont junction occurs within embayments, the actual rate of piedmont formation will be much greater than that achieved by mountain front retreat alone. In embayments the rate of piedmont formation will depend upon the length of the embayments as well as upon the rate of piedmont junction migration (Parsons and Abrahams 1984, p. 258).

General Model of Pediment Formation

Fundamental Concepts

A number of fundamental geomorphological concepts provide insights useful for developing a comprehensive understanding of pediments and their formation.

(a) Analysis of a geomorphic system is in large part determined by the temporal and spatial limits that are used to define the system (Schumm and Lichty 1965). For example, the concept of dynamic equilibrium provides useful insights concerning the tendency for adjustment of slope declivity to lithology and/or structure within an upland area: however, the concept is less useful when applied to the problem of upland degradation over 'cyclic time' (unless mass continues to be added to the system through uplift). Depending on the time and space perspective of the analysis, processes may appear to be steady or fluctuating, continuous or discontinuous, and forms may appear to be stable, unstable, or quasi-stable. When considering the evolution of a landscape or the interrelations among its components, it is essential to define the system within those scales of space and time that are appropriate to the problem at hand. Regarding the specific problem of pediment formation, the appropriate spatial scale is very likely the pediment association (Cooke 1970) and the appropriate temporal scale is cyclic time (Schumm and Lichty 1965). Viewed from this perspective, emphasis is placed on interdependent changes between landscape components and processes in a changing open system (Bull 1975).

(b) Most geomorphic systems are multivariate and operate both non-linearly and discontinuously; therefore, these systems may respond complexly to mass and energy inputs (whether these inputs are non-linear or linear, continuous or discontinuous). 'When the influence of external variables such as isostatic uplift is combined with the effects of complex response and geomorphic thresholds, it is clear that denudation, at least during the early stage of the geomorphic cycle, cannot be a progressive process. Rather, it should be comprised of episodes of erosion separated by periods of relative stability, a complicated sequence of events' (Schumm 1975, p. 76).

(c) Process transitions and thresholds further complicate the operation of most geomorphic systems; and the behaviour of such systems may be particularly complex at or near these transitions. In the specific case of pediments, the locus of pediment formation is generally considered to be the piedmont junction, a zone of abrupt transition for both morphology and process. Moreover, the pediment itself serves as the transition between range and alluvial plain, and over the long term, pediment drainage operates at or very close to the threshold of critical power in streams. 'In the western Mojave Desert, evidence for the movement of the upper limits of alluvium across piedmont plains is provided by such features as channels buried beneath alluvium downslope of the limits, and upslope of the limits by the truncation of soil and weathering profiles and the presence of alluvial outliers. Movement of the boundary need not be accompanied by dissection of the plain, but this is frequently the case' (Cooke 1970, p. 37). Like many boundary zones in non-linear systems, this continuously shifting transition is both complex and chaotic. Hence, the timing and duration of periods of erosion, transport, or aggradation at any point in the system may very well be indeterminant, particularly where fluvial components of the system are operating at or are approaching the threshold of critical power.

(d) Form and process are closely interrelated within the pediment association. At any point in time, form is an initial constraint on process. As the distribution of form changes through time in response to process, the distribution of process changes accordingly. For example, '. . . the relative importance of divide removal [between embayments] and mountain front retreat . . . is strongly dependent on the size and relief of the associated mountain mass, and hence . . . it is likely to change in a systematic way as the mountain mass diminishes in size through time. Specifically . . . as a mountain mass diminished in size there is a tendency for . . . divide removal to become progressively less important relative to mountain front retreat' (Parsons and Abrahams 1984, p. 265). Similarly, the relative importance of lateral planation very likely also diminishes with time. 'In an arid or desert region, flow over a long time and detrital load are most likely to exist where an ephemeral stream emerges from a canyon having a considerable length and a fairly large drainage basin within a mountain area. Such conditions prevail in the cycle of the erosion from youth to postmaturity. In old age, however, the detrital load is small, the dissection of the original mountain mass is far advanced, and the streams are easily diverted out of their channels. All of these factors tend to minimize lateral planation' (Bryan 1936, p. 775). Thus within the pediment association, different processes operate at different rates and times in different parts of the system and the distribution of these processes changes in space through time.

(e) As an open geomorphic system, the pediment association is 'a complex multivariate phenomenon which responds at different rates to different external variables such that it need not be in equilibrium with all variables at once' (Palmquist 1975, p. 155). Thus, the various components of the pediment association are not necessarily adjusted to one another or to the system as a whole. Rather, as interdependent components of an evolving open system, the progressively degrading uplands and aggrading alluvial plains of the pediment association are examples of landforms that are not necessarily attracted toward a steady state. 'For many landforms, height, volume, or other dimensions change progressively with time instead of tending toward a time-independent size or configuration' (Bull 1975, p. 112).

(f) Pediments may be formed by many different combinations of processes operating within and constrained by a variety of tectonic, climatic, and lithologic/structural environments. Moreover, as slowly developing long-lived landforms, pediments may be subjected to substantial changes in these environmental factors as they develop. Thus many pediments may be, in part at least, forms inherited from former conditions. This possibility severely limits the profitable application of morphometric analysis to the question of pediment development. Unless a high degree of similarity of initial size and shape and of subsequent morphogenetic history can be demonstrated, morphological differences between pediments or their associated uplands cannot be used to develop general inferences about range degradation or pediment formation.

Factors Affecting Pediment Formation and Expansion

A model of pediment development which applies equally well to all possible geomorphic situations is likely to be either oversimplified with respect to any specific situation or unmanageably complex if all possible situations are analysed in detail; consequently, the following summary focuses on piedmont pediments in arid and semi-arid environments.

(a) As piedmont surfaces, pediments serve as zones of transition (and transport) between uplands and alluvial plains. In arid and semi-arid regions, the redistribution of mass between the degrading upland (where stream power generally exceeds critical power) and the aggrading alluvial plain (where stream power is generally less than critical power) proceeds mainly by physical processes. Erosion within the upland is driven primarily by the action of concentrated water flow, development of integrated drainage networks from rills to rivers and the concomitant formation and expansion of valleys and embayments (Lustig 1969, p. D67, Wallace 1978, Bull 1979). The more diffuse and uniformly distributed processes of hillslope weathering and erosion operate within this general geomorphic framework of fluvially carved valleys and embayments. Moreover, the great bulk of material removed from the mountain mass is transported via streamflow, and debris transport across the pediment surface as well as deposition on the alluvial plain are likewise dominated by fluvial action.

(b) Long-term geomorphic stability appears to be a fundamental requirement for the formation and maintenance of extensive bedrock pediments. Within the south-western United States, pediments are larger and more abundant in areas of low vertical tectonic activity where base level is either stable or slowly rising. Such conditions facilitate the approximate long-term balance between erosional and depositional processes on piedmonts that is essential for pediment development. 'The pediment, instead of being abnormal and restricted to certain localities, is widespread throughout . . . [the desert regions of western North America] and is the type of plain normally developed during quiescent periods. Alluvial fans on the other hand, probably cannot be made under static conditions. They are built where normal gradients have been changed by faulting, warping, lateral erosion, or other special causes' (Blackwelder 1929, p. 168).

(c) Pediment formation is unidirectional through time. 'When tectonic stability exists, a geomorphic system is partially closed to materials. . . . Thus the [general] behavior of the system during denudation becomes predictable; a continued decrease in relief must occur. Therefore, a dynamic equilibrium within a drainage basin or a hillslope as a whole cannot exist for material flux except when the rate of uplift equals the rate of denudation. . . . The imbalance in material flux means that changes in the size, elevation or form of the system must occur' (Palmquist 1975, pp. 155–6). Although individual components of the pediment association (e.g. hillslopes, valley bottoms, stream channels, piedmont junctions, etc.) may be generally adjusted to lithology and process over the short term; each component and its relations to the other components change systematically and asymmetrically over the long term. As a fluvially dominated open system, the pediment association is attracted towards a stationary state; it is 'indeed directed through time' (Montgomery 1989, p. 51).

(d) Pediment formation (expansion, maintenance, and modification) varies in time and space, both non-linearly and discontinuously. At any point in time, processes of pediment formation vary in space (Sharp 1940). For example, as a pediment develops and expands along a mountain front, it is also maintained as a surface of transportation in more distal areas. Expansion involves fluvial processes (in channels, valleys, and embayments) and hillslope processes (on valley and embayment sideslopes and on mountain front interfluves). Maintenance involves weathering, sheet flow, and streamflow processes on the piedmont. Moreover, the relative significance of these processes also changes over time (Bryan 1936). Initially when the mountains are large and the piedmonts narrow, upland degradation predominates; eventually as the mountains become smaller and the piedmonts broader, pediment maintenance and regrading dominate. The rates at which these processes operate also vary (both in time and space). Degradation in response to uplift is initially rapid and gradually slows (Morisawa 1975). Other factors being equal, larger streams generally erode more rapidly and, therefore, occupy larger deeper valleys with lower gradients than do smaller streams (Bull 1979). Consequently, the effects of fluvial action are seldom distributed uniformly along the mountain front. The mountain front is distinctly non-linear in plan (Lawson 1915) and, in most instances, does not retreat parallel to itself (Parsons and Abrahams 1984).

(e) The retreat of mountain front spurs appears to be less important with regard to the evolution of mountain fronts (and pediment development) than fluvial downwearing and embayment development (Lustig 1969, p. D67, Bull 1984). In the absence of significant fluvial erosion, tectonically inactive mountain fronts may remain relatively unchanged over periods of as much as several million years (Dohrenwend et al. 1985, Mayer 1986, Harrington and Whitney 1991). Strongly asymmetric ranges clearly illustrate the dominance of fluvial erosion on mountain front degradation. The small subparallel drainages along the scarp flank erode relatively slowly inducing little change; whereas the large integrated drainages of the backtilted flank erode rapidly inducing drainage basin development, valley widening, embayment enlargement, valley sideslope convergence and spur decline. The presence of deeply dissected, partly stripped remnants of older, higher, and more steeply sloping relic surfaces along many mountain fronts also argues against general mountain front retreat as a primary mode of mountain mass reduction.

(f) Medial piedmonts are very stable and often persist with little dissection or regrading over periods of several million years (Dohrenwend et al. 1985, 1986). Although the threshold of critical power in streams may be crossed abruptly, a generally long-term balance between stream power and critical power appears to be maintained across a considerable breath of many piedmont surfaces. Where a drainage system is able to adjust channel form and roughness in response to limited variations in discharge and sediment load, it may accommodate such changes while at the same time continuing to operate at the critical power threshold. Also where the piedmont is broad, changes in discharge or load can be accommodated, at least in part, by upslope or downslope migration of the erosion-transport and transport–deposition limits of the critical power threshold (Cooke and Mason 1970). The stability of the medial piedmont implies that it is the local base level for mountain degradation, proximal piedmont dissection, and pediment formation.

(g) The critical requirement for pediment formation (and maintenance) is that all piedmont drainage must operate, on average, at or very close to the threshold of critical power across a broad zone of transition between upland and lowland. Within this zone, episodes of dissection and aggradation must be limited in both time and space. Because pediment formation is a slow process that is largely concentrated along the junction between upland and piedmont, pediment formation and maintenance must proceed simultaneously in different parts of the system. Upland slopes may retreat and/or decline by a variety of processes, but conditions must be maintained such that an expanding pediment replaces the shrinking mountain mass. Embayment formation and expansion, inselberg development, pediment formation and maintenance all result from the erosional–depositional balance of the medial piedmont.

(h) Although long-term change is relatively uniformly distributed across the pediment; short-term changes may be non-uniformly and ‘chaotically’ distributed. The precise character of these short-term changes is determined in part by the system’s ability to ‘self regulate’ responses to short-term variations in discharge or sediment load. This self regulation is significantly influenced by the availability of readily erodible and transportable materials. If such materials are abundant and uniformly distributed across the piedmont, the tendency for lateral erosion usually exceeds the tendency for vertical incision and the system will regrade uniformly in space without significant dissection. However, if these materials are not abundant and/or are not uniformly distributed, the tendency for vertical incision may equal or exceed the tendency for lateral erosion and the system will very likely regrade locally and discontinuously.

(i) Proximal dissection is a common characteristic of both pediments and alluviated piedmonts. This regrading occurs in response to a sustained increase in stream power (duration and/or intensity of flow) or a sustained decrease in critical power (amount and/or calibre of available load). Possible causes include short-term climate changes (Bryan 1922), long-term undirectional climate change, tectonic tilting of the piedmont, complex response of the upland–proximal-piedmont drainage system (Rich 1935, Denny 1967), and isostatic adjustment of the piedmont association (Howard 1942, Cooke and Mason 1973). Over the long term, however, the most likely cause would appear to be the redistribution of mass within the pediment association. Regrading is constrained by the stability of the medial piedmont; if the medial piedmont is undissected and morphologically stable then proximal dissection must be the result of change within the adjacent upland. As uplands diminish in size, both stream discharge and sediment load decrease but the effect of the decrease in load appears to predominate.

It is a general misconception that most alluviated piedmonts are primarily if not entirely depositional landforms. In fact after initial formation, even bajadas may be as much zones of regrading and transport as they are loci of deposition. Proximal dissection is common on both pediments and bajadas, and the tendency for dissection by piedmont drainage underscores the delicate erosional–depositional balance of many proximal and medial piedmont areas. Field measurement of badlands dissection (Schumm 1956, 1962) and comparison of the relative positions of adjacent modern and relict piedmont surfaces capped by late Tertiary lava flows (Dohrenwend et al. 1985, 1986) also document the erosional predilection of proximal piedmont areas.

(j) Subsurface weathering undoubtedly facilitates pediment formation via conversion of resistant bedrock into non-resistant regolith that can be disaggregated into readily loosened and easily transported detritus. However, it would appear unlikely that the operation of sursurface weathering processes actually controls pediment development in those situations where the overall mode of landscape evolution proceeds via upland retreat and replacement by expanding piedmonts (e.g. Bryan 1922, Johnson 1932b, Lustig 1969, p. D67, Parsons and Abrahams 1984). As the preservation of tens of metres of intensely weathered bedrock beneath granitic pediments in the south-western United States clearly demonstrates, many pediment surfaces are not closely related to either the position or the form of the subsurface weathering front. Rather, the existence of an alluviated piedmont, pediment, or similar surface of low relief would seem to be a required precondition for development of a quasi-planar weathering front and for subsequent planation of its deeply weathered regolith mantle. Hence, exhumation of the weathering front, mantle-controlled planation, and other similar processes are probably most effective where the overall mode of landscape change is one of stepwise weathering and stripping in a region of low relief (e.g. Mabbutt 1966).

Conclusion

There is a large literature on pediments. This literature has provided a wealth of information on the form of pediments and the settings in which they are found. We know much about the character and variability of pediments. In contrast, information about pediment processes is notably less. Even setting aside the difficult problem of the relationship of contemporary processes to pediment development, knowledge of these processes, *per se*, and their significance for contemporary sediment budgets in deserts is weak, and much needs to be done. In large measure, the process paradigm of geomorphology in the second half of the twentieth century appears to have passed pediments by. The origin of pediments and their long-term evolution has been the subject of much speculation, but this speculation has remained rooted in the approach that characterised geomorphology of the early part of the twentieth century. The problem is an immense one. Such evidence as we have indicates that pediment evolution is slow. Consequently, these landforms must have developed through times of significantly varying climate. In the study of badlands, Howard (1997) uses a model of drainage-basin evolution (Howard, 1994) to simulate landscape evolution. There is concordance between the output of the model and the real landscape, but he notes that this concordance should be regarded as preliminary. 'The simulation model process rate laws and model parameters have not been directly validated and calibrated by field observation. Long term process observations would be valuable. In particular, rates of erosion in bedrock rills and channels could be measured and related to drainage area, channel gradient and rainfall history to calibrate an erosion law' (p. 224). Although the problem is a simpler one, inasmuch as badlands evolve rapidly and the vicissitudes of climate can reasonably be ignored, the approach would seem to provide a way out of the current *impasse* in advancing understanding of pediment evolution (e.g. Strudley et al., 2006). Suitably constrained by data on rates of sediment movement and downwearing (e.g. Nichols et al., 2005) numerical modelling of this type could provide a way of testing the various models that have been proposed for pediment development.

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