Channel-scale erosional bedforms in bedrock and in loose granular material: character, processes and implications

PAUL A. CARLING, JÜRGEN HERGET, JULIA K. LANZ, KEITH RICHARDSON and ANDREA PACIFICI

Summary

High-energy fluid flows such as occur in large water floods can produce large-scale erosional landforms on Earth and potentially on Mars. These forms are distinguished from depositional forms in that structural and stratigraphical aspects of the sediments or bedrock may have a significant influence on the morphology of the landforms. Erosional features are remnant, in contrast to the depositional (constructional) landforms that consist of accreted waterborne sediments. A diversity of erosional forms exists in fluvial channels on Earth at a range of scales that includes the millimetre and the kilometre scales. For comparison with Mars and given the present-day resolution of satellite imagery, erosional landforms at the larger scales can be identified. Some examples include: periodic transverse undulating bedforms, longitudinal scour hollows, horsehoe scour holes around obstacles, waterfalls, plunge pools, potholes, residual streamlined hills, and complexes of channels. On Earth, many of these landforms are associated with present day or former (Quaternary) proglacial landscapes that were host to jökulhlaups (e.g. Iceland, Washington State Scablands, Altai Mountains of southern Siberia), while on Mars they are associated with landscapes that were likely host to megafloods produced by enormous eruptions of groundwater. The formative conditions of some erosional landforms are not well understood, yet such information is vital to interpreting the genesis and palaeohydraulic conditions of past megaflood landscapes. Correct identification of some landforms allows estimation of their genesis, including palaeohydraulic conditions. Kasei Valles, Mars, perhaps the largest known bedrock channel landscape, provides spectacular examples of some of these relationships.

2.1 Introduction

Large floods on Earth are known to erode loose granular material and bedrock to form channels. For example, jökulhlaups can erode bedrock (Baker, 1988; Tómasson, 1996; Carrivick et al., 2004). However, many moderate river flows, as well as large floods, are capable of eroding bedrock over prolonged periods of time by abrasion induced by bedload (Sklar and Dietrich, 1998, 2001) and suspended load (Richardson and Carling, 2005). A major challenge will be to determine the relative importance of long-term slow erosion induced by frequent flows and major 'cataclysmic' floods that can induce very noticeable changes in channel forms. By analogy, water floods of a variety of scales should also be capable of eroding granular materials and bedrock on Mars and evidence of such fluvial erosion would strengthen the argument for water-induced geomorphological activity on the Earth's neighbouring planet.

Channel-scale erosional bedforms are worthy of study for at least two reasons. Firstly, correct identification of key features may elucidate the nature of the flow hydraulics and, in some cases, the magnitude of the formative discharge may be determined. Erosional antedunes, for example, (detailed later in this review) can be related to transcritical and supercritical flow regimes. Secondly, in the case of Mars, identification of erosional 'fluvial' bedforms implies the presence of water rather than some other liquid or deformable flowing solid such as ice. However, qualitative and quantitative understanding of the origin of erosional bedforms and their geomorphological significance is immature. Erosional bedforms in loose granular materials have received the most attention especially in relation to flow around non-deformable obstacles (e.g. horseshoe scour around bridge piers) but erosion in bedrock has been little studied. Consequently, this review begins with a brief historical perspective that largely considers the bedrock erosion of the Channeled Scablands of Washington State, and this is followed by a proposed typology of large-scale erosional bedforms that provides a framework for description and discussion. Examples of eroded bedforms on Earth and Mars are considered.
2.2 Historical perspective

Many scientists in the nineteenth century were interested in fluvial catastrophism as an agent in transforming the landscape. Whilst most invoked the Noachian biblical flood, a few grasped the importance of large-scale fluvial action caused by natural events rather than a flood induced by a deity. Parker (1838), for example, recognised that large-scale dry valleys, locally termed ‘coulees’ in Washington state, were fluvially eroded: a theme furthered by others (Russell, 1893; Dawson, 1898; Salisbury, 1901; Calkins, 1905). Baulig (1913), in particular, recognised that the coulees could be water-eroded and identified dry waterfalls, water-scoured rock basins and plunge pools, but attributed these to the normal action of a prior course of the Columbia River. However, he commented specifically on the scale of the features, presumably having some difficulty reconciling the scale with normal fluvial action.

Bretz’s (1923a) geomorphological study of the Columbia Plateau in southeastern Washington state may be taken as the benchmark introduction to the study of catastrophic erosional landscapes. A map accompanies the paper of Bretz and it depicts patterns of abandoned anastomosed canyon-like waterways – the coulees. Many of these channels contain seemingly abandoned waterfalls, cataracts, plunge pools, hanging tributary and distributary valleys and large-scale potholes, all eroded into the basalt bedrock. The entire composite became known as the ‘Channelled Scabland’ (Baker, 1978a) and most of this historical perspective considers the scabland (see also Baker, 2008). Bretz (1924) supported his arguments in favour of vigorous fluvial action by describing large-scale erosional features in the modern Columbia River near The Dalles. Bretz (1923b) also identified streamlined loess hills as flood-eroded remnants. Bretz (1930b) synthesised his ideas inasmuch as he realised that the total suite of erosional features could only be genetically related by invoking a short-duration, large-volume and high-velocity flood. As recorded by Baker (1973b), the scabland erosional features include more than one hundred overfit anastomosing channels, hanging valley junctions, gigantic potholes, large longitudinal grooves up to 5 m deep and 50 m wide, bedrock pool–riffle sequences at 10 to 15 km spacings, streamlined loess hills and flood-sculpted lozenge-shaped residual outcrops of basalt. Malde (1968) described similar features associated with the Pleistocene breakout flood from Lake Bonneville through the Snake River in Idaho.

Although the plucking of bedrock on a large scale by catastrophic floodwaters became understandable with the hydrodynamic study of Baker (1973a) there are antecedents. There is evidence from contemporary correspondence that Pardee had considered as early as 1910 that the Scablands might be related to catastrophic drainage from a large Pleistocene lake. Bretz identified the source of the floodwaters as glacial Lake Missoula in 1928 and although he made his views known, he did not publish at this time. Harding, without consultation or acknowledgement, published aspects of the ideas of Bretz in 1929. However, the intellectual attribution was properly made when the full details of an abrupt ice-dam failure were published (Bretz, 1930a, 1932). Pardee published his ideas linking glacial lake Missoula with the floods in 1942.

But it was not until later in the twentieth century that the hydraulic formation of large-scale channels in the scablands was considered in detail (Baker, 1973a, 1973b). Better conceptualisation of the progress of floodwaters across the landscape included the recognition that floodwaters might be hydraulically ponded at constrictions and bendways in the flood course, a concept that can be tested using modern hydraulic models of flood progression and field evidence. Such constrictions, causing local elevation of the water surface, are sometimes marked by supposed scour lines in the landscape, such as scars eroded into loess that are presumed high-water marks. This scenario was described by Wilt (1972, 1977a, b) for late Pleistocene flooding down the Columbia River. In the case of Siberian Altai Quaternary floods, on a scale similar to the Missoula floods, ponding might also have occurred in the Katan River valley at Big Ilgumen (Carling et al., 2002c). The Altai floods have left little evidence of erosional landforms; rather, it is the erosion of the Yakima basalt in the scabland by the Missoula flood that has left the most visually impressive erosional topography. The distinctive rock-jointing habit of the basalt means that colonnade and entablature structures are well developed, often as columns of prismatic jointed blocks. Baker (1973a) demonstrated that large flood-transported blocks and cobbles emanated from these columns. Bretz (1924, 1932; Bretz et al., 1956) recognised that macroturbulence and the distinctive jointing patterns in the basalt when combined could lead to ready bedrock erosion and Embleton and King (1968) suggested that cavitation might have caused the scour of deep Scabland rock basins. Often plucking was concentrated at tension-jointed antincillar crusts, usually well represented at divide crossings that are common in the Scablands. Water-scoured crossings consist of narrow, generally parallel-sided, short, flat-floored and steep-walled bedrock channels (Baker, 1978c) where water-filled valley flows spilled over into adjacent channel systems, often truncating rock slip-off slopes leading to truncated bluffs that Bretz (1928) called ‘trenched spur buttes’. Divide crossings also are important because they are often a component in the anastomosed channel pattern.

2.3 Processes of bedrock erosion

There are several accounts of the mechanisms of fluvial erosion in bedrock channels, and it is not necessary to
describe the processes in detail here. The interested reader is referred to the useful works of Allen (1971a), Selby (1985), Hancock et al. (1998), Wohl (1998), Wende (1999) and Whipple et al. (2000). However, brief definitions of the various processes are provided here for reference (see also Richardson and Carling, 2005).

1. Plucking, also known as quarrying and jacking, involves the removal of whole blocks of rock from the boundary by lift and drag forces acting on the block. The blocks are delineated by joints and other structural features of the rock and undergo a period of preparation prior to their entrainment by the flow. Block preparation involves the widening and propagation of cracks and the loosening of blocks by a combination of hydraulic forces, bedload impact, wedging of sediment within cracks and physical and chemical weathering (Whipple et al., 2000). Related to plucking is the process of flaking in foliated rocks such as shales and slates, which occurs on a smaller scale. Flaking involves the removal of thin rock fragments by entrainment in the flow, bedload impact, wind or ice.

2. Abrasion, also known as corrosion, involves the wearing away of a surface by numerous impacts from sediment particles transported in the fluid. These particles may be carried in suspension or as bedload, or both. Each impact, if sufficiently energetic, breaks off a small piece of the boundary material, which may be at the scale of a small part of a grain in the case of suspended load, or as much as a substantial flake or chip of rock in the case of coarse bedload.

3. Fluid stressing, also known as eversion, refers to the removal of particles by the turbulent stresses exerted directly on the boundary by a clear fluid. At the scale of individual grains and grain aggregates, this process is likely to be important only in clays and poorly consolidated rocks. However, at larger scales, this mechanism is important in flaking and it is the process by which prepared blocks are plucked.

4. Cavitation occurs in very rapid flows when the instantaneous dynamic fluid pressure locally drops below the vapour pressure and bubbles of water vapour appear. The water effectively boils, though energy is supplied by the high-velocity water flow instead of by external heating. The bubbles very rapidly collapse, however, and the pressure shock wave thus generated erodes any nearby boundary. Cavitation is known to occur on engineering structures (Arndt, 1981), but whether it occurs in natural channels is still a matter of debate (Baker, 1973a, b; Sato, et al. 1987; Wohl, 1992b; Baker and Kale, 1998; Hancock et al., 1998; Tinkler and Wohl, 1998a; Gupta et al., 1999; Whipple et al., 2000).

Corrosion occurs on soluble rocks, where it is also known as dissolution, and on chemically reactive rocks, where it is referred to as chemical weathering. Soluble rocks include limestone, marble and evaporites. Other rocks, such as quartzites, extrusive igneous rocks, granitic rocks and rocks with a high calcareous content such as marl, are slightly soluble or have soluble components and may also show evidence of dissolution. Chemically reactive rocks have unstable components, such as feldspars, olivine and sulphide minerals, which break down into various decay products and are subsequently easily removed. Corrosion may occur subsurface or subaerially. It acts both directly in removing the rock or components of it, and indirectly in physically weakening the rock and assisting its removal by the other erosion mechanisms. It is possible that the growth of algae, moss and lichen, etc. on the rock enhances corrosion by inhibiting evaporation and maintaining a film of water in contact with the rock when surfaces elsewhere are dry.

Physical weathering acts indirectly in fluvial erosion by physically weakening the rock (Whipple et al., 2000) and increasing its surface area as a result of crack propagation, thereby enhancing erosion by the mechanisms listed above. Perhaps the most important physical weathering processes are wetting/drying cycles, which may be particularly important in the processes of flaking and weathering due to the crystallisation of salts (Sparks, 1972), and freeze/thaw cycles, which may play a role in crack propagation, block loosening and rock disaggregation. Blocks of rock can also be removed directly by ice lift, when a layer of ice forms on the rock surface under low flow in winter, and rises to the surface during a subsequent flood, carrying with it small attached blocks of rock (see Carling and Tinkler, 1998) for a discussion on the effect of ice.

2.4 A typology of channel-scale erosional bedforms in bedrock

Bedrock channel morphologies have traditionally been considered to be determined by the physical and structural properties of the substrate (Ashley et al., 1988), and to have boundaries that are essentially imposed on the flow and cannot be adjusted by hydraulic processes because of their high resistance (Baker and Pickup, 1987) or are adjusted only rarely by high-magnitude, low-frequency
events in which the fluid forces exceed the high resistance threshold. This is in contrast to alluvial rivers, whose boundaries have long been known to be deformable by relatively low-magnitude, high-frequency events, and which are known to adjust their geometries according to hydraulic and sediment conditions (Wolman and Miller, 1960). However, in recent decades there has been an increasing realization that bedrock channel boundaries should also be considered deformable (presumably over long time scales in response to low-frequency, high-magnitude events) and that channel adjustments are conditioned both by hydraulic controls and by any imposed bedrock structural controls. This is shown by the development of bedforms in bedrock channels apparently analogous to those found in alluvial channels (e.g. Keller and Melhorn, 1978; Baker and Pickup, 1987; Ashley et al., 1988; Wohl, 1992a; Wohl and Grodek, 1994; Koyama and Ikeda, 1998; Wohl and Ikeda, 1998), and by the recognition that definable channel width-area and slope-area relationships exist for bedrock rivers (Montgomery and Gran, 2001; Kobor and Roering, 2004). Furthermore, the conclusion that rivers in bedrock can adjust their channel boundaries is a logical consequence of the observation that many bedrock channels are extensively covered with water-eroded bedforms (Richardson and Carling, 2005).

Eroional bedforms in bedrock fluvial channels on Earth are extremely diverse in morphology, orientation and scale. Herein, the only features considered in detail are channel-scale features that might be identified in plan from aerial photographs and satellite images. Eroional bedforms are defined with respect to both positive and negative bed elevations including extensive fields of undulating forms. Eroional bedforms are therefore residual surfaces developed within a stratustrum. The morphology of negative features may be simple, representing a single locus of enhanced erosion, or compound, consisting of multiple centres of erosion. Planform geometries range from near-circular, large-scale potholes to extended linear furrows. Longitudinal, transverse, obliquely orientated and unorientated bedforms occur. In terms of spacing, erosional bedforms may exist as isolate individuals, as conjugate individuals with shared boundaries, or as coalesced individuals (Richardson and Carling, 2005).

An important consideration in any discussion of erosional bedforms in bedrock relates to the range of features to be considered and the processes by which they are generated. The dominant process will depend principally on the structure of the substrate; rock that has closely spaced fractures or planes of weakness will be dominated by the plucking of blocks and fragments, while other more homogeneous rocks will generally be dominated by abrasion (Bretz, 1924; Whipple et al., 2000). Cavitation is often reported as being of possible importance in extreme events. This was an idea promulgated in early flood studies (e.g. Dahl, 1965; Embleton and King, 1968) and one that has persisted (e.g. Baker and Kale, 1998; Gupta et al., 1999; Whipple et al., 2000). Without gainsaying this supposition, there is little overt evidence for the action of cavitation in fluvial systems (Barnes, 1956) and this matter is not considered further in this chapter. Corrosion is generally the most important agent of erosion in carbonate rocks, and may also be significant in rocks prone to chemical weathering, such as granite. The shapes of plucked fragments and the spaces they leave behind are defined by the geometry of the structural elements of the substrate, while features scoured by abrasion in relatively homogeneous surfaces reflect local hydrodynamic patterns. However, at the channel scale it should be remembered that plucking of blocks may result in channel-scale features that are morphologically similar to channel-scale abraded features. A hypothetical example would be series of step-pool pairs that might be clearly definable in a plucked bedrock system although therein the mesoscale roughness of the bed would reflect the structural control (e.g. a joint pattern). A tangible example of such a feature is the c. 40 m diameter pothole eroded through plucking by the catastrophic Pleistocene Missoula floods described by Baker (1973b).

This section illustrates the range of erosional bedforms in bedrock described on Earth. Published descriptions come mainly from observations of active forms in modern rivers rather than relict superfluvial bedforms. However, it is assumed that the fundamental processes responsible for bedforms operative in modern rivers would also be operative in extreme events and that most sculpted forms essentially exist as a self-similar suite of features over a wide range of scales, and in flows over a wide range of discharges (Kor et al., 1991; Kor and Cowell, 1998). Importantly, some bedforms have dimensions directly related to flow velocity, for example (small-scale) scallops (Curl, 1966; Blumberg and Curl, 1974). Similarly, longitudinal furrows described by Ikeda (1978), amongst others, seem to scale with flow depth. Bedrock pool–riffle sequences (Keller and Melhorn, 1978) appear to scale with the width of the flow-field. Deriving such relationships for channel-scale bedforms would provide potentially powerful tools to reconstruct aspects of palaeofloods (Springer and Wohl, 2002).

2.4.1 Concave features

Potholes Near-circular potholes are a well-known member of a suite of concave sculpted forms (terminology from Richardson and Carling, 2005) that ranges from equant plan-view geometries to extended, highly linear furrows. Potholes may be of similar size to the channel width (Richardson and Carling, 2005). Examples, as reported from the Missoula flood tract, can be upwards of 40 m in diameter but of variable depth. Within the eponymous
Potholes Coulee of the Channeled Scabland, small lakes up to 250 m in diameter, which are often quasi-circular, are probably potholes, being associated with smaller but nevertheless readily identifiable large potholes, which are often an integral part of the scoured bed of dry channels. The Potholes cataaract complex (Plate 1) also exhibits large pothole-shaped plunge pools. Many modifications on the simple pothole morphology can be found. Large potholes may also be breached as they enlarge into neighbouring sculpted forms; extremely convoluted surfaces can result from the coalescence of a group of closely spaced potholes, for instance. Consequently, for studies of relict large-scale features on Earth and Mars, the absence of a classic rounded plan view for a suspect pothole need not preclude the identification of the feature as a pothole.

**Flutes** Flutes are generally shallow scour hollows with a rim that is typically parabolic in plan view, convex in the upstream direction, and open in the downstream direction. They occur immediately downstream of some defect in the surface, which sets up vortex scour in its wake (Allen, 1971b). First described by Maxson and Campbell (1935), they are well known in sedimentary environments (Dzulynski and Walton, 1965) and have been studied experimentally (Allen, 1970a, b). Flutes are also common in bedrock, although they are morphologically less diverse than their sedimentary counterparts and typically lack a median ridge, a common feature of sedimentary flutes. Nevertheless, flutes in bedrock occur with a wide range of width:length and width:depth ratios. Small-scale flutes with overhanging upstream ends are common in bedrock (e.g. Maxson and Campbell, 1935; Hancock et al., 1998).

**Furrows** Closed linear concave features (with a length more than twice their width) are termed furrows herein (see Richardson and Carling, 2005) for a more detailed typography). Generally, these are orientated longitudinally and again a diverse spectrum of forms exists. Those with a relatively high width:length ratio have elliptical outlines in plan view and are named short furrows. Parallel-sided furrows are more elongated forms whose rims on either side are parallel for some distance. The rims may be cuspat or rounded. Straight, curved, sinuous and bifurcating varieties occur. Longitudinal furrows can be initiated at defects and sometimes occur in the shear zone between the wake of a large obstacle and the free stream (Tinkler, 1997) but generally are unrelated to any such feature. They may often be found in the steeply dipping region leading into a knickpoint. Longitudinal furrows can often be found in groups of parallel individuals, sometimes with regular spacing (Blank, 1958; Ikeda, 1978; Wohl, 1993). Compound longitudinal furrows, in which a very long feature is subdivided into several shorter furrows by ridges, are common. Wohl (1993) described long furrows that contained regularly spaced short depressions. These depressions grew in a downstream direction as the furrow deepened, and became small offset lateral potholes on alternate sides of the furrow. With further distance downstream, these potholes coalesced into an inner channel with undulating walls (see also Wohl et al., 1999).

Not all concave features are orientated with the assumed flow direction. Furrows may slant obliquely down a sloping channel margin in the downstream direction. Other furrows are orientated transverse to the flow direction (e.g. Shrock, 1948; Jennings, 1985; Kor et al., 1991). Transverse furrows are often found in the lee of some defect or line of defects that runs in a cross-stream direction, but they also occur in the absence of any obvious irregularity. Transverse furrows can have simple or compound morphologies; in the latter case, they appear to be constituted by smaller, initially isolate bedforms that have coalesced in the transverse direction. A peculiar form of the transverse furrow is that in which a single well-defined, relatively narrow furrow runs almost the entire width of the channel (Tinkler and Wohl, 1998b).

Small-scale to medium-scale linear or sinuous furrows often occur in open bedrock channels (Richardson and Carling, 2005). However, as a cautionary note, large-scale linear furrows have also been associated with subglacial meltwater (e.g. Bradwell, 2005) and with the action of ice-streams (e.g. Canals et al., 2000; Dowdeswell et al., 2004). Various scale-independent furrows known collectively as s-forms are also associated with subglacial meltwater (Kor et al., 1991). Larger-scale lineation believed to be of fluvial origin has been reported from the Channeled Scabland (see above) and which presumably parallels the palaeoflow. Often these lineations are picked out by the presence of linear lakes that fill the depressions: examples include Lena Lake and Table Lake within the Grand Coulee and as parallel and anastomosed networks of channels immediately upstream of the famous Dry Falls (Plate 2). Patton and Baker (1978) describe scabland grooves up to 300 m wide, 25 m deep and up to 2.5 km in length. Lineation might be caused by longitudinal vortices in the flow (e.g. Coleman, 1969) and this is the preferred model for long linear parallel depositional ridges that develop on Earth in shallow coastal seas in subtidal locations (Blondeaux, 2001). Within fluvial systems (Ikeda, 1978) and (less exactly) within coastal waters, the spacings of these ridges tend to scale with flow depth, with spacing variation induced possibly by other subordinate parameters. It is not clear, however, why they should develop so persistently across wide swaths of the flow path, as is illustrated below for erosional examples. Erosional grooves of similar and lesser scale to depositional ridges also develop on Earth in intertidal and subtidal muddy environments, but also occasionally in soft bedrock
Figure 2.1. A portion of Mars Orbiter Camera (MOC) image M07-00614. The location is in Aalabasca Valles, near 9.66 N 204.19 E. The overlying elevational data points are from the Mars Orbiter Laser Altimeter (MOLA); the values are negative because this region is below the mean datum for Mars. The black line on the right graphically represents the elevational data. The image is 3.02 km across. The scene is illuminated from the upper left.

(Allen, 1969; 1987). These ridges and grooves may bifurcate, usually with the grooves bifurcating in the direction of the dominant tidal flow, but are persistent over tens of metres to hundreds of metres as distinctive features. More recently, extensive and visually striking parallel ridges and grooves have been identified on the continental rise in the Gulf of Mexico (Plate 3), where they persist for tens of kilometres and parallel measured benthic currents (Bryant, 2000; Bean, 2003), and in the Gulf of Lions, Mediterranean Sea (Canals et al., 2006). These features are very similar in appearance and scale to bedforms recorded on Mars, where strongly parallel linear systems (Figure 2.1) have been identified together with weakly and strongly anastomosed networks of eroded channels (Figures 2.2 and 2.3) that have not been well studied. Often the furrows on Mars are contained within, and are parallel with, distinct canyon walls and so may be formed by flow-induced vorticity, which adjusts to the channel expansions and contractions (compare Plate 3 and Figure 2.1), and diverge 'smoothly'.

Figure 2.2. Longitudinal grooves and braided channel systems cut into bedrock within Ares Valles, Mars. The view is developed from an HRSC image. (Images courtesy of HRSC. THEMIS Public Data Releases, Mars Space Flight Facility, Arizona State University, 2006, http://themis-data.asu.edu.)

Figure 2.3. 500 m high putative former waterfall in Ares Valles, Mars. Flow bottom right to top left. Note the linear grooves that are well developed close to the lip of the falls and the secondary channels to the left of the main transverse fall. The view is developed from an HRSC image. (Images courtesy of HRSC. THEMIS Public Data Releases, Mars Space Flight Facility, Arizona State University, 2006, http://themis-data.asu.edu.)

(Figure 2.1) around obstacles similarly to ocean-floor grooves on Earth (Bean, 2003). Elsewhere, parallel-grooved terrain conceivably could form within large lake bodies, such as might have existed on Mars (Parker et al.,...
1993; Parker, 1994), as wind-driven Langmuir circulation cells can cause large-scale longitudinal vortices to be established within a fluid (Leibovich, 2001), although it is not known if this causes erosion of the bed of the water body.

**Large-scale channels, inner channels and knickpoints**
Distinctive linear bedrock channels, reminiscent of river courses, occur on both Earth and Mars. On Earth, these channels may be dry or they may contain small streams that cannot be responsible for cutting the larger channels. Channels may be relatively straight, sinuous and highly meandering or they may form as a network of interconnecting channels termed anastomosed systems (Figures 2.2, 2.3 and Plate 4). Structural control may be evident locally but otherwise the channels are cut within bedrock with the morphology conditioned by hydraulic action. Lateral benches or terraces may be evident together with longitudinal variation in channel width (beaded) and longitudinal variation in bed elevation, including rifflake pools sequences and steep cascades. Suites of smaller-scale bedforms may be evident within these channels and some of these were described above.

A common channel cross-section morphology is that of the incised inner channel with bedrock benches on either side that may extend for hundreds of metres or indeed many kilometres. Descriptions of inner channels include those of Bretz (1924), Baker (1973b), Nemec et al. (1982), Baker and Pickup (1987), Wohl (1992a, b, 1993), Wohl et al. (1994), Baker and Kale (1998) and Wohl and Achyuthan (2002). In the Mekong River through Laos, for example, inner channels occur frequently (Gupta, 2004). Very deep and narrow channels are termed 'slot canyons'. A closely related phenomenon to that of the extensive inner channels is the more locally developed, deeply incised knickpoints that occur in vertical fallattles or inclined cascades or other short and very steep reaches. These knickpoints represent locally enhanced channel incision. In the experience of the authors, inner channels almost always possess a knickpoint at their head, implying that even extensive inner-channel formation occurs through knickpoint retreat (Baker, 1973b; Wohl, 1993; Wohl et al., 1994), although an exception is noted below. However, the formation of inner channels has been observed experimentally through the simple incision of a plane bed (Shepherd and Schumm, 1974; Kodama and Ikeda, 1984; Wohl and Ikeda, 1997; Johnson and Whipple, 2007). A common mechanism of knickpoint retreat in homogeneous rocks is that of pothole growth and coalescence in the reach immediately above the knickpoint (Wohl, 1998; Wohl et al., 1999). This results in inner channels and slot canyons with convoluted wall topography (Eldon, 1917, 1918; Jennings, 1985; Wohl, 1999; Kunert and Coniglio, 2002) in which undulations on opposite sides are generally out of phase (Wohl, 1993; Wohl et al., 1999; Wohl and Achyuthan, 2002). However, an inner channel has also been observed with in-phase wall undulations (Richardson and Carling, 2005). A stair-like succession of knickpoints may be termed a cataract (Wohl, 1998). Knickpoint migration is also an important mechanism of erosional escarpment retreat (Weissel and Seidl, 1998).

Deeply incised inner channels are common within the scabland tracts and often emanate from former locations of cataracts. Potholes Coulee and Frenchman Springs are two such channels that seem to have developed by headward retreat of a knickpoint beneath floodwaters. Examples of inner channels in modern rivers are usually associated with knickpoint retreat (Richardson and Carling, 2005). Elsewhere floodwaters within the scablands excavated the pre-flood stream valleys to leave pre-flood tributaries joining the main stems as hanging valleys. However, as a cautionary note, it should be recognised that deeply incised inner channels can also result not from floodflow and knickpoint retreat over short periods of time but through tectonic uplift and incision over many thousands of years.

**Erosional constrictions and hydraulic jumps**
Flow through channel constrictions can cause intense erosion of bedrock and modelling studies have indicated that Missoula flood flows through several scabland constrictions (notably Soap Lake, Wallula Gap and Staircase Rapids) would each have been characterised by a hydraulic jump with critical or supercritical flow within the constrictions. Thus, narrow but short lengths of channel that are essentially short canyons (gorges) that lie downstream and upstream of broader sections of channel might be erosional features attributable to critical flow in large floods. The presence of a flood-formed landform in this instance is not readily determined from field study, but through hydraulic modelling. An example is the constriction on the Mae Chaem River in Thailand, where river flow (maximum recorded flow > 1000 m³ s⁻¹) is constrained to flow through a vertical slot a few metres wide (Kidson et al., 2005). Identification of such sites is important as they can be used to constrain palaeo-discharge estimates.

**2.4.2 Convex features, undulating features and composite features**
Unlike concave features, there is little agreement on terminology and the identification of channel-scale convex undulating and composite features. There are relatively few published descriptions of large-scale forms in relict flood landscapes. Thus the simple subdivision in this section reflects the limited knowledge base rather than an exhaustive consideration. It is not known how this class of bedform relates to the erosional antidunes discussed in a subsequent section.
Convex features are not as diverse as their concave counterparts, presumably reflecting the relative uniformity of flow patterns that exist around positive features as compared with flow patterns within negative features. The most important feature in terms of abundance is the upstream-facing convex surface. This bedform is all but ubiquitous in bedrock channels with significant roughness relief, and consists simply of the rounding and streamlining of the upstream surfaces of positive features. Other convex features include bladed forms, in which erosion within two or more contiguous scour hollows produces an intervening keeled ridge, faceted obstacles (e.g. Wohl, 1992b; Maxson, 1940; Baker and Pickup, 1987) and streamlined hills, some of which may be analogous to a group of similar features in aeolian environments, often termed yardangs, which are well described on both Earth and Mars (e.g. El Baz et al., 1979; Greeley and Iversen, 1985; Goudie, 1999).

The major group of undulating features is the hummocky forms, of which two types occur. The gently rounded type are found as longitudinal, transverse and non-directional forms with wavelengths for those so far described at the decimetre scale (Richardson and Carling, 2005). Although very common in massive lithologies, they are often overlooked as a bedrock bedform, perhaps because of their subtle appearance. Previous descriptions of hummocky forms include the 'undulating surfaces' of Kor et al. (1991) and Tinkler (1997), and the 'water-smoothed undulatory surfaces' of King (1927). Equally common are the second type, the ripple-like and dune-like sharp-crested hummocky forms (SCHPs). These are frequently described, albeit under a variety of names, for example the 'eversion marks' of Angeby (1951), the 'horns' of Jennings (1985), the 'ripple-like bedforms' of Sevon (1988), Hancock et al. (1998) and Whipple et al. (2000) and the 'hummocky surfaces' of Wohl (1992b). Crest planforms range from two-dimensional to highly three-dimensional and the troughs may take on complex morphologies.

**Cataracts** Abandoned near-vertical or inclined waterfalls within dry channels can be impressive evidence of formative flood waters. The most famous of these is Dry Falls shown on the Coulée City Quadrangle map of the Grand Coulee, which is 120 m high and 5.5 km wide and developed in jointed basalt (Plate 2). Below the falls, the former plunge pool is occupied by Dry Falls Lake. Equally impressive are the steep horseshoe-shaped cataracts to the east that plunge into Red Alkali Lake and Castle Lake. Bretz (1932) and Bretz et al. (1936) argued that these various falls were developed sub-fluvially rather than by sub-aerial erosion and knickpoint retreat. This supposition was later supported by evidence of high-water marks (Baker, 1973a). Figure 2.3 shows a 500 m high cataract on Mars. Of note is the well-developed furrow system upstream of the lip of the falls. Many of these furrows become well incised at the lip and continue down the face of the falls as chute and chimney furrows (terminology after Richardson and Carling, 2005).

**Step-pool and chute-pool sequences** Step-pool sequences consist of a series of cataracts that often are spaced at roughly regular intervals and are found in channels that are steeper than those that support the pool–riffle sequences described below (Wohl, 2000; Wohl and Legleiter, 2003). Steps in such sequences can be very steep, near-vertical drops in the channel floor, but whereas water is in freefall over a step, water remains largely in contact with the rock surface as supercritical flow over an inclined chute. The intervening pools may be deeply scoured into the bedrock. In many cases, such regular variations in gradient need not be structurally controlled (Carling et al., 2005) and where structural control is important (Wohl, 2000) it is often evident in air photographs or satellite images. Kodama and Ikeda (1984), Wohl and Ikeda (1997) and Koyama and Ikeda (1998) have produced repeating steps experimentally in homogeneous cohesive substrates, whilst Wohl and Grodek (1994), Koyama and Ikeda (1998) and Wohl and Ikeda (1998) describe field examples of step–pools in homogeneous substrate. Step–pool sequences may become modified by the process of pool breaching, of which there are two varieties. Breaching may occur when a headcut progresses from the immediate downstream pool through the intervening step (Carling et al., 2005) and such breaches should be visible in aerial photographs and satellite images. Drainage of the upstream pool during low flow is enhanced such that its low-flow volume is reduced, but during high flow ponding occurs such that the pool volume is large. Thus the wetted perimeter of such a series of breached steps and intervening pools, when viewed from the air, appears as a string of 'narrow beads' (pools) and intervening 'threads' (breached steps) during low flow and as a string of 'broad beads' during high flows. More rarely, the base of a pool may be breached, although the step on the downstream side remains intact. This occurs when there is a strong counter-current acting against the headwall of the downstream pool, which in time causes the recesion of the base of the headwall until a breakthrough is made into the base of the upstream pool. In this case, the upstream pool is drained completely. The breached step may then constitute a channel scale ‘arch’, which will only be evident in plan-view images if the lighting is oblique and produces a definitive shadow depicting the nature of the arch. However, in wet systems, the ‘disappearance’ of water that is clearly visible upstream and downstream of an arch is a good indicator of the possible presence of an arch or other subterranean conduit.
Pool–riffle sequences As noted for step–pools, a common feature of bedrock channels is quasi-periodic variations in gradient. In low-gradient channels this is usually termed a pool–riffle sequence (e.g. Dury, 1970; Keller and Melhorn, 1978; Baker and Pickup, 1987; Wohl, 1992a; Thompson, 2001; Wohl and Legleiter, 2003) whereas in high-gradient channels, step–pool (and chute–pool) sequences result. Riffles may constitute either bedrock highs upstream and downstream of bedrock-floor pools (i.e., riffles are high points in an erosional setting) or isolated accumulations of coarse gravel (depositional setting) separated by rock-floor pools. Based largely upon visual appearance and morphology, most authors reporting these features deem them analogous to pool–riffle sequences in alluvial rivers. However, whereas there has been considerable progress in recent years in understanding the formation and maintenance of alluvial pool–riffle sequences (see Wilkinson et al., 2004), the controlling hydraulic processes in bedrock channels are not well understood. In contrast to alluvial rivers wherein the amplitude of both pools and riffles may adjust at the same time scale, in a bedrock river over relatively short time scales there is a greater opportunity to reconstruct an alluvial riffle but excavation of a pool requires a longer time period. The genetic distinction and any relationship of pool–riffle sequences in bedrock channels to step–pool systems has not been explored. Patton and Baker (1978) reported a crude palaeo-pool–riffle sequence incised into bedrock in the Cheney-Palouse Scabland and this seems to be the only example of a ‘fossil’ sequence that has been published. Very deep pools (<70 m deep) have been reported from low-gradient bedrock reaches of the River Mekong and bathymetric maps show intervening rocky riffles, but it is not known if these are features of regular spacing. Thus pool–riffle sequences in low-gradient bedrock channels are little reported and controlling mechanisms not elucidated. Keller and Melhorn (1978), however, found that average riffle spacing in bedrock pool–riffle sequences was identical to that in alluvial channels at between five and seven channel widths. Undulating channel beds have been noted within some small run-off Martian channel systems in the southern highlands (Kerészstúri, 2005; Lanz et al., 2000) but not in the large outflow channels of the northern highlands.

Erosional antidunes and critical flow Large-scale ‘dune-like’ transverse ridges have been described from a variety of locations on Earth using satellite, aerial photography and ground and bathymetric surveys (Baker, 1973a; Alt, 2001; Carling et al., 2002c; Fildani et al., 2006), but the majority of these are largely depositional forms. MOC images of the Martian surface show transverse ridges that may be analogous to dune-like bedforms on Earth (Burr et al., 2004). Although sometimes called ‘ripples’ in the older literature, given the scale of the bedforms and the high turbulence levels of megafloods on Earth it is reasonable to conclude that these features are not ripples, which cannot form in fully turbulent flows (Carling, 1999). Rather they are most likely dunes or antidunes. Both dunes and antidunes develop by erosion of the underlying sediments within the troughs between bedforms and deposition of sediments to form the bedforms. However, these features tend to migrate within the flow such that preserved bedforms consist wholly of deposited sediments (see Carling et al., this volume Chapter 3).

Descriptions of erosional wavy surfaces with the morphological characteristics of antidunes are rare (Fildani et al., 2006). Bedforms developed in mobile sediments tend to be limited in size by negative feedback mechanisms associated with the flow and bedform migration and by poorly understood system instabilities that tend to split bedforms more often than merging them. In contrast, the shape and spacing of antidunes eroded into bedrock will be subject to flow controls alone as there is no bedform migration and no bedform splitting and amalgamation, rather only erosion of the crests, troughs and flanks. Over-deepening eventually will be precluded by reduced shear stresses at the bed. However, there are no systematic studies of erosional antidunes that could provide scaling data on bedform geometry and formative flows. Nevertheless, antidunes forming in water flows observed on Earth cover at least two orders of magnitude in scale (0.1 m to 10 m) and are known to increase in dimensions with flow depth (Allen, 1984). Thus it is not improbable to observe larger ‘fossil’ antidunes (100 m to 1000 m) in the landscape of Earth or indeed Mars that would have scaled with the magnitude of the water floods. Antidunes are especially useful for palaeoflood reconstruction because the wavelength of the spacing of the transverse ridges corresponds to the spacing of the standing water waves that formed them, and this latter spacing can be related to the flow speed (U) of the fluid (Kennedy, 1963; Allen, 1984; Tinkler, 1997). Froude number, Fr, is defined as the flow speed, U, ratioed to the gravity wave speed, \sqrt{gh}, where g is the gravitational acceleration and h is the flow depth. Thus, given that antidunes form in transcritical flows, Fr > 0.82 to c. 1.2, it is possible to derive a water depth (h). This depth and flow speed, if coupled with information on the limits of the flow width (W), can indicate the discharge (Q) magnitude through the continuity equation: \[Q = UhW.\]

Possible megaflood erosional antidunes occur near the village of Chagan Uzun in the Chuja Basin, south-central Siberia (Carling et al., 2002c; Herget, 2005). These landforms consist of large-scale (tens of metres high; wavelengths of up to 300 m) sub-parallel transverse waveforms that were formerly locally much more extensive, but have been erased by postglacial river meandering (Plate 5).
These bedforms occur immediately downstream of a hill that would have caused a standing wave to develop in its lee if floodwaters overtopped the hill, or alternatively the location is one where flow would have been channelled into a narrower constriction and possible transcritical flows would have developed locally. Riverside exposures of the internal structure of these bedforms show clearly that the core of the structures is a weakly stratified coarse cobble and boulder glacial diamicton with many angular fragments as well as well-rounded clasts. This diamicton is around 10 to 15 m thick within the highest bedforms. The caps of the bedforms, however, consist of a conformable drape of finer pebble-sized gravel up to 2.5 m thick that is thickest on the crests of the large bedforms and thins and pinches out down both flanks. In the caps many particles are angular, but sub-rounded particles also exist and, although no clear bedding is discernable, the discontinuity between the diamicton and the cap indicates a separate depositional event. On the stoss side of one large bedform, the cap sediments are deformed into a few short-span dune-forms up to 2 m high with wavelengths up to 4 m. Taking the dunes as evidence for deposition from water flow, it is concluded that the large-scale bedforms in the diamicton are erosional antidunes developed beneath standing waves within transcritical flows, and that the finer-grained caps were deposited during subcritical waning flow. Transverse boulder ridges with amplitudes less than those of antidunes are usually termed ‘transverse ribs’. They may form in the same manner as antidunes but this is disputed (see Carling, 1999). Good examples from Earth have been reported from Modrudalur in Iceland, where the spacings were used to estimate the palaeoflood velocity and water depth (Rice et al., 2002).

Streamlined residual landforms Large-scale bedrock macroforms that might reflect flood sculpting usually are crudely formed. Lozenge-shaped residual outcrops of basalt in the Cheney-Palouse Scablands may represent flood-formed residual massifs or en masse a group of these with their associated intervening channels give an impression of a braided or anastomosed channel (Plate 4). However, although such an association of macroforms might be formed by catastrophic flood flow, similar associations are also noted where structural influence on channel macroforms is important. A particularly good example of the latter is the ‘Four Thousand Islands’ reach of the Lower Mekong River within Laos, close to the border with Cambodia (Plate 6). Here Quaternary dykes, as young as 5720 yr BP, have been eroded into a myriad of bedrock-cored sandy islands and rocky islets by Late Pleistocene and Holocene discharges that today can exceed 56,000 m³ s⁻¹ during each annual wet season. The resulting landscape is reminiscent of both furrow complexes and streamlined landforms with an underlying structural control on channel alignment.

Equally striking are the streamlined residual hills cut into scabland loess (Plate 7). Bretz (1923b) recognised that hundreds of isolated loess hills in the eastern scablands were characterised by steep margins with distinctive prows at the up-flow margins. An origin owing to fluvial erosion was posited by Bretz et al. (1956), and Baker (1973b) argued that many had been formed sub-fluvially by floodwater emanating from Lake Missoula. Regardless of overall size, many of the streamlined hills are approximately three times as long as they are wide. Measures of maximum length, maximum breadth and total planform area can be used to demonstrate that each hill has a close resemblance to an airfoil in shape, for which form drag is minimised, which supports a hypothesis of streamlining by water (Baker, 1979; Komar, 1983).

Obstacle marks Foremost amongst composite features are obstacle marks, which consist of a concave current-scor rugose crescentic curving around the upstream side of a convex obstacle (Peabody, 1947; Karcz, 1968). These scour elements may have some depositional features but are here treated as erosional bedforms. Typical patterns of scour are reproduced at a variety of scales from the very small (e.g. 10⁻² m: Bunte and Poesen, 1994) to the very large (10² m: Fay, 2002). Baker (1978c, his Figure 4.7) provides an example of the very large from the Channeled Scabland where Lenore Canyon and Long Lake occupy scour depressions that wrap around an obstacle formed by High Hill and Pinto Ridge. A scour hole at a boulder is also described from the Lake Missoula flood (Baker, 1978b; Baker et al., 1987). Other possible examples are considered below.

Earth scientists have used a variety of expressions to describe these features and these include the term ‘obstacle mark’ (e.g. Karcz, 1968; Reineck and Singh, 1980; Allen, 1984; Russell, 1993) as used herein, as well as ‘scour mark’ (e.g. Allen, 1965; Richardson, 1968; Baker, 1973b; Elfrström, 1987), ‘current crescents’ (Bridge, 2003), ‘crescentic/hairpin erosional marks’ (Shaw, 1994), ‘obstruction-formed pool’ (Hassan and Woodsmith, 2004), ‘scour hole’ (Baker, 1973a) and ‘comet mark’ for special large-scale obstacle marks (Werner et al., 1980). The variety of erosional obstacle marks incised into bedrock has been illustrated and considered by Richardson and Carling (2005) and an overview of the variety of current crescents and shadows formed by unidirectional currents is given by Allen (1984). In engineering sciences, scour around isolated obstacles, such as bridge piers, that protrude through the water surface have been studied intensively (e.g. Shen, 1971; Breusers et al., 1977; Breusers and Raudkivi, 1991; Hoffmans and Verheij, 1997; Melville and Coleman, 2000;
Richardson and Davis, 2001). Less well studied are fully submerged obstacles (Carling et al., 2002a, b).

Many factors influence the final shape of the erosional hollow. These include: obstacle shape and alignment to flow, flow speed and flow steadiness, Reynolds number, flow depth, sediment grade or bedrock type and time of development (e.g. see Melville and Coleman, 2000 and Herget, 2005). Unfortunately, the complex interaction of different controlling factors means that it is frequently difficult to recreate the hydraulic conditions associated with obstacle marks within palaeochannels. Some flume-derived relationships for submerged or free-surface obstacles are difficult to apply to obstacle marks in the field as they are valid only for the range of flume conditions (e.g. Johnson, 1995) and scale effects are frequently not regarded (Ettema et al., 1998). On the other hand, Johnson (1995) has argued that, by comparing several different approaches, often consistent data sets emerge that tend to point to a narrow range of possible hydraulic conditions. Finally, guidelines and computer software to address the problem are available (e.g. Richardson and Davis, 2001; Landers et al., 1996).

Among the largest obstacle marks on Earth are those related to cataclysmic ice-dammed lake outburst floods from Pleistocene times, such as those from Lake Missoula in the northwestern USA (e.g. Baker and Bunker, 1985; Baker et al., 1993) or in the Siberian Altai Mountains (e.g. Carling et al., 2002c; Herget, 2005). These floods inundated valleys and basins to depths of up to hundreds of metres, while submerged flood-resistant bedrock hills acted as obstacles inducing large-scale local scour around them. One example is the bedrock hill located in the western Chuya Basin of the Altai Mountains, Siberia. During Late Pleistocene times, valley glaciers blocked the course of the River Chuya downstream, generating an ice-dammed lake with a depth of up to 350 m (2100 m a.s.l.) above the bedrock hill (Herget, 2005). During the failure of the ice dam, the currents draining the Chuya Basin formed a connected scour hole with a maximum depth of 8.1 m, a length of 91.5 m and a wall width of about 400 m in front of the hill with a height of about 50 m. (Plate 8). A palaeohydraulic interpretation of the obstacle scour at Chagun-Uzun is problematic but one has been presented by Herget (2005). Obstacle marks are also found along the pathway of the Lake Missoula Flood in the northwestern USA. A particularly large example is the volcano outcrop ‘Rocky Butte’ in the eastern parts of the city of Portland (Allen et al., 1986) but no hydraulic interpretations have been advanced beyond general descriptions. According to Alt (2001), Locust Hill, located about 100 km northwest of the city of Missoula in Montana, divided the flow reaching it from the east into two branches. The current system in front of the obstacle scoured the bedrock and generated a depression. This scour hole, called Banana Lake due to its characteristic shape in front of the hill, is still filled with water today (see Pardee, 1942).

2.5 Example of a very large bedrock channel complex: Kasei Valles, Mars

The remainder of this review describes an immense landscape on Mars that has probably been subject to water erosion. It is one of many intriguing features on the Martian surface that were discovered in 1971 by the American Mariner 9 probe and termed outflow channels, in accordance to their strong resemblance to terrestrial rivers. Outflow channels commonly start abruptly and fully developed, usually from circular to elliptical depressions called chaotic terrains, large collapse depressions in which the collapsed material lies tilted and broken in a chaotic assemblage on the depression floor. Some start equally developed at deep fissures, mainly in the vicinity of large volcanotectonic rises. They have no or few tributaries, a large and relatively constant width to depth ratio and a low sinuosity.

Baker and Milton (1974) were the first to notice the resemblance of the Martian outflow channels to the Channeled Scablands in Washington and Oregon, USA, that have been shaped by Pleistocene catastrophic glacial floods. The sudden outburst of the glacially dammed Lake Missoula stripped away surface materials in a high-velocity turbulent flow, shaping a bizarre and quite unique landscape with an assemblage of erosional features similarly similar to what are found in the outflow channels on Mars. It is now widely believed (yet not generally accepted) that the outflow channels were formed by similar processes of sudden and catastrophic outbursts of large amounts of water. A common scenario introduced by Carr (1979) is as follows (in summary form):

Surface water, abundant in the very early history of Mars, oozed away forming a large groundwater system. Globally falling surface temperatures and a developing and growing global permafrost layer trapped the groundwater in a system of large aquifers. The increasing depth of the permafrost basis increased the pore pressure in the aquifers. Sudden groundwater outbursts might have occurred as a result of the pore pressure exceeding the lithospheric pressure which would destabilize the aquifers and open faults and cracks in the overlaying permafrost-rock layers, allowing the highly pressurised water to escape onto the surface. Tectonics or meteorite impacts could also have broken the permafrost seal. Emptying of the aquifers and decreasing discharge rates would refreeze the remaining water, thus closing the disrupted permafrost seal. Carr suggested that the aquifers could have refilled, leading to a repeating cycle of outflow activity and groundwater recharge. As the outflowing water could not have been reintegrated into the groundwater cycle due to the permafrost seal and would have quickly been lost to the thin Martian atmosphere.
by evaporation, the loss of water from the system would have been great and the ability of the cycle to repeat itself strongly limited.

Other processes of outflow channel generation have been proposed involving e.g. glacial activity (Lucchitta and Anderson, 1980; Lucchitta et al., 1981; Lucchitta, 1982, 2001), liquefaction (Nummedal, 1978) or gas-supported density flows (Hoffman, 2000) similar to pyroclastic flows on Earth but under cryogenic temperatures and with carbon dioxide being the active agent in the "floods." A combination of different processes is imaginable and even likely, especially, regarding the extreme environment on Mars, a combination of fluvial and water-ice processes possibly in the form of ice-covered rivers.

The Kasei Valles lie in the western circum-Chryse region, separating the Luna Planum Highlands from the Tharsis and Tempe Terra volcanotectonic provinces (Plate 9). They are 3000 km long, in parts up to 400 km wide and 3–4 km deep, making them the largest outflow channel system on Mars. They originate in a relatively shallow north–south-oriented depression that adjoins a large tectonic graben closely connected to the Valles Marineris canyon system. Two branches of Kasei Valles are distinctive, North Kasei Vallis (NKV) and South Kasei Vallis (SKV), to the north and south of the Sacra Mensa. The Kasei Valles are one of the most interesting places to study on Mars due to the fact that those parts which are not covered by young lava flows exhibit many typical erosion features mentioned above, though at an immense scale.

2.5.1 Topography

The most prominent topographic features of Kasei Valles are several large terraces (Plate 10). These provide a unique opportunity to study different phases of channel development in more detail. The Mars Orbiter Laser Altimeter (MOLA), an instrument onboard the American Mars Global Surveyor (MGS) mission, measured the Martian topography in great detail and generated a global topography model with a pixel resolution of 3 km. The along-track resolution of single MOLA tracks is even better, with only 300 m between single laser spots and a vertical resolution of 1.5 m (with a relative error in altitude along MOLA profiles given to be 1–10 m). These data allow the generation of cross-profiles and long-profiles of Kasei Valles (Figures 2.4 and 2.5).
The two branches of Kasei Valles, NKV and SKV (Figure 2.6), show a distinct morphology. NKV is very broad and mostly shallow, showing numerous examples of typical outflow morphology, whereas SKV is rather narrow, very deep (up to 4 km) and morphologically more uniform. In Figure 2.4 several cross-profiles of NKV and SKV are shown, indicating terraces and other channel features, providing a general overview of the different channel shapes of the two Kasei Valles branches. A correlation of terraces based on MOLA data shows that, despite the different appearance of NKV and SKV and a less distinct terracing in SKV, the development of both branches probably was connected and contemporaneous as terrace heights correlate strongly (in the error range given by the instrument). The widths of the deep central parts of both valley branches (Plate 9) are similar and both channels show a noticeable increase in the width to depth ratio towards the channel mouth. The depths of NKV and SKV differ slightly with the channel floor of SKV being generally 200 m deeper than the NKV floor indicating that activity in SKV might have lasted longer.

The long-profiles of Kasei Valles give another insight in channel development. The good MOLA coverage allows generation of long-profiles of the channel floor as well as along prominent valley terraces. Figure 2.5 shows long-profiles of the valley floor of NKV and SKV (lower lines), two main terraces (middle lines) and the adjacent highland surfaces to the north and to the south (upper lines). The long-profiles of the terrace floors are very uniform with steep slopes parallel to the adjacent highland surfaces, whereas along the channel floors the slopes are generally low except along two prominent knickpoints visible in both
NKV and SKV at approximately the same longitude that developed at later stages during the channel evolution. The knickpoints might indicate a change in bedrock resistance or the position of tectonic faults as they lie quasi-parallel to tectonic features that dominate the adjacent highlands. They are not an indication of a change in the erosion level during channel development, as is often the case on Earth.

Based on MOLA data, information can be inferred regarding channel slopes and water depths (bankfull stages) of assumed floods, allowing better calculations of maximum discharge rates of the palaeofloods. Older calculations based on Viking data gave discharge rates between $10^9$ m$^3$ s$^{-1}$ (Baker, 1982) and $2.3 \times 10^9$ m$^3$ s$^{-1}$ (Robinson and Tanaka, 1990) for the Kasei Valles floods (during the initial stages). Calculations based on MOLA data show that the discharge rates were probably more in the range of large floods on Earth or only slightly higher, with values between $5 \times 10^9$ m$^3$ s$^{-1}$ and $10^9$ m$^3$ s$^{-1}$ for the initial stages of channel development and values as low as $5 \times 10^8$ m$^3$ s$^{-1}$ during the end of the outflow activity (Lanz, 2004; Williams et al., 2000).

2.5.2 Morphology

The most prominent morphologic characteristics of Kasei Valles, the streamlined islands and longitudinal grooves (Figure 2.6), have been used as arguments for both glacial and fluvial activity. High-resolution imagery from the Mars Orbiter Camera (MOC) onboard MGS shows details that seem to support the flood hypothesis. Figure 2.7A shows, for example, possible scour marks along the upstream side of a streamlined island in Kasei Valles. Scour marks are typical of erosion around obstacles by flowing water, though they do not necessarily rule out a glacial origin. Another interesting feature is seen in Figure 2.7B. This MOC image shows the downstream side of a streamlined island. It has a ‘fretted’ appearance with small erosional alcoves on both sides of the island, merging at its tail-end. The fact that these features appear only at the tail-end of the island and no talus deposits can be found rules out younger denudation processes, as they would affect the whole island and should leave clearly visible landslide deposits. It is therefore proposed that they are the result of erosion in turbulent high-velocity flows. Gouges or potholes shaped by glaciers can be similar in appearance but these are usually randomly spaced features rather than aligned along the downstream side of an obstacle. Another fact inconsistent with a purely glacial origin of the outflow channels is that to date no glacial deposits have been identified clearly in any of the outflow channels on Mars and it seems unlikely, especially imagining the size of the glaciers needed to carve these enormous valleys, that they simply vanished leaving no signs of the millions of cubic metres of material they eroded along the way.

![Figure 2.7](image)

Figure 2.7. (A) Possible current-scour around upstream side of a streamlined island (MOC-Image E1401150). (B) 'Fretted' island in Kasei Valles; the erosion features appear only at the tail end and no talus deposits are visible (MOC SP245505).

2.5.3 The time scale of outflow events in Kasei Valles

Besides the question of what shaped the Martian surface and which agents were involved (e.g. water, ice, carbon dioxide), the timing and sequence of depositional and erosional processes and events needs to be established. Impact crater statistics provide a tool for determining
relative and, within model-dependent limitations, absolute time scales (Neukum and Hiller, 1981). The method of defining relative ages of surface units by crater statistics is based on the assumption that all solid planetary surfaces, or rather geologic units inside these surfaces, accumulate more and more craters with time. By measuring and comparing the frequencies of impact craters superimposed on these surfaces it is then possible to determine the relative ages of units. Units exhibiting high crater frequencies are generally older than those with lower crater frequencies as they have been exposed longer to impact cratering processes. If the time dependency of the cratering rate of a planet is known, as is the case for Mars, absolute ages can then be deduced from the crater frequencies (for a detailed discussion of the method see e.g. Neukum and Wise, 1976a, b; Hartmann, 1977; Hartmann and Neukum, 2001; Ivanov, 2001).

The dating of different Kasei Valles units based on Viking, MGS and Mars Express imagery gives results regarding the evolution of the outflow channel. Outflow channel activity started approximately 3.7 billion years (Ga) ago and ended possibly as late as 1.0 Ga ago in the Upper Amazonian period of Mars history (Lanz, 2004; Neukum et al., 2007), giving a time span of up to 2.7 Ga during which the Kasei Valles have been the site of recurring episodic outflow events. The outflow activity seems to centre around two main phases of increased activity. The first phase between 3.6 and 3.1 Ga ago eroded the broad reaches of the upper terrace floors. The second phase eroded the grooved terrace floors and deep inner channels and started approximately 2.2 Ga ago. Outflow activity during both phases was not continuous but happened in recurring pulses of presumably short time-spans. Nevertheless, these studies show that the erosional activity in the Kasei Valles lasted much longer than previously believed (Neukum and Hiller, 1981; Nelson and Greeley, 1999).

The long (episodic) activity of the Kasei Valles channel system requires a process that periodically triggers the outflow events. Catastrophic outbursts of groundwater as described by Carr (1979) partly meet these criteria. However, these processes are temporally limited due to the loss of water from the cycle by the refreezing permafrost seal and it is unlikely that they could have lasted for almost 3 Ga. In addition, a period of enhanced outflow activity towards the end of the cycle, as indicated by the crater frequency analyses, can also not be explained by these processes. Another possibility is that the fluvial activity was linked to pulses of volcanic activity (e.g. Lanz, 2004; Neukum et al., 2007). Neukum et al. extensively re-mapped and dated volcanic and fluvial units in the Kasei Valles and other regions on Mars based on the latest Mars Express data. They found that both volcanic and fluvial activity show common episodic pulses in intensity throughout Mars history. They believe that these pulses are related to the interior evolution of the planet when convection in the asymptotic stationary state changes from the so-called stagnant-lid regime to an episodic behaviour. Therefore, it appears likely that volcanic activity may have triggered the outflow events, e.g. by the melting of ground-ice and/or the mobilisation of subsurface waters.

2.6 Conclusions

Erosional bedrock forms occur over an immense range of scales, from small flutes and furrows of bedrock rivers to the streamlined hills, cataracts, and anastomosing channels of the Channeled Scabland, and their immense counterparts on Mars. All of these forms contain information about generative hydraulic processes and considerable progress has been made in understanding some of these relationships. However, despite the detailed classification of channel-scale bedforms that can be developed for small modern river systems (Richardson and Carling, 2005), relatively few types of channel-scale erosional bedforms have been described previously and no detailed process-based typology has been developed to define features unambiguously. There is considerable opportunity to advance understanding of channel-scale features using high-resolution study of the erosion of materials that mimic bedrock, aided by computer modelling of the development of such features given different hydraulic conditions. Aeolian, glacial and lava-flow processes also can produce erosional landforms that have characteristics very similar to fluvial and megaflood features (e.g. Evans, 2003) and consequently care is required in identification and interpretation. This is especially so when the only information available is drawn from morphological planform data obtained using satellite or aerial photography. Recent studies often allow height data to be derived using techniques such as photogrammetry, or shape-shading (Beyer et al., 2003; Burr et al., 2004). Where self-similar spatially contiguous groups of erosional bedforms occur then suites of morphological geostatistics, such as fractal properties, might help resolve the genetic origins of scale-specific landforms (Evans, 2003). Such a geostatistical approach has yet to be applied to megaflood erosional landforms as usually the population of self-similar features is small (see Carr and Malin (2000) for a perspective).

References


