15. CHANNELS AND VALLEY NETWORKS

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Channels, valleys, and related features of aqueous origin on Mars are of profound importance in comparative planetology. Martian outflow channels formed by large-scale fluid outflow from subsurface sources. Elements of cataclysmic flooding, debris and ice flowage, as modified by volcanism and wind action, seem best to explain observed morphologies. Martian valley networks show their greatest similarity to terrestrial networks formed by sapping as water emanated from seeps and springs. Exogenetic sources of recharge (precipitation) or endogenetic cycling (hydrothermal systems) would be required to achieve the indicated extensive valley development during the heavy bombardment phase of Martian history. Recent studies of valley network development associated with volcanism suggest that endogenetic cycling is likely and that it may not be necessary to invoke extensive epochs of warm, wet atmospheric conditions to explain valley network formation on Mars.

I. INTRODUCTION

Mars is the only planet besides Earth known to manifest the dynamic workings of a hydrological cycle. The channels and valleys of Mars are a bold testament to the past operation of that cycle and its profound implications for environmental change on that planet. However, to understand environmental change on Mars, one must have an exact understanding of the genesis of channels and valleys. The importance of this problem traces back to James Hutton, the acknowledged founder of geology, who ascribed the origin of valleys on Earth to the prolonged denudational action of rivers (Hutton 1788,1795).

To ascertain the origin of Martian channels and valleys, genesis is first inferred from detailed study of landforms imaged by remote sensing devices on spacecraft. Hypotheses are generated by the reasoning process of abduction (Engelhardt and Zimmerman 1988), whereby local details of morphology are used to infer probable mechanical processes, geologic controls, and ultimately the exogenetic and endogenetic environments of formation. This is a nontrivial exercise involving a type of reasoning (Gilbert 1886) that lies at the heart of geology. Application of the methodology in planetary studies is discussed by Mutch (1979) and Baker (1984,1985).

Although most generally recognized from imagery of the 1972 Mariner 9 mission (Masursky 1973; McCauley et al. 1972; Milton 1973), the fluvial forms on Mars were also inferred from studies of Mariner 6 and 7 images showing signs of crater dissection by valleys (Schultz and Ingerson 1973). Because of the incompatibility of the implied liquid water with the modern Martian environment, it was quickly hypothesized that the forms must be relict from an ancient warm, dense atmosphere (Sagan et al. 1973*a;* Sharp and Malin 1975; Mutch et al. 1976*c*). Indeed, the valley networks in particular have become a principal element of evidence that the Martian atmosphere evolved from an early volatile-rich state to its present condition (Walker 1978; Pollack 1979; Cess et al. 1980; Squyres 1984; Kahn 1985; Carr 1987; Pollack et al. 1987; see chapter 32). Because of the importance of this inference, the present review will examine the evidence in some detail.

II. OUTFLOW CHANNELS

Martian outflow channels are large-scale complexes of fluid-eroded troughs. The flows which formed these channels appear to have emanated from discrete collapse zones known as chaotic terrain (Fig. 1). The channels are immense by terrestrial standards, as much as 100 km wide and 2000 km in length. Gradients of channel floors range from nearly zero to 2.5 m km^{-1}. Many reaches of outflow channels appear to include closed depressions. Morphological attributes of the channels are extensively discussed by Baker (1982), whose conclusions are briefly reviewed and updated here.

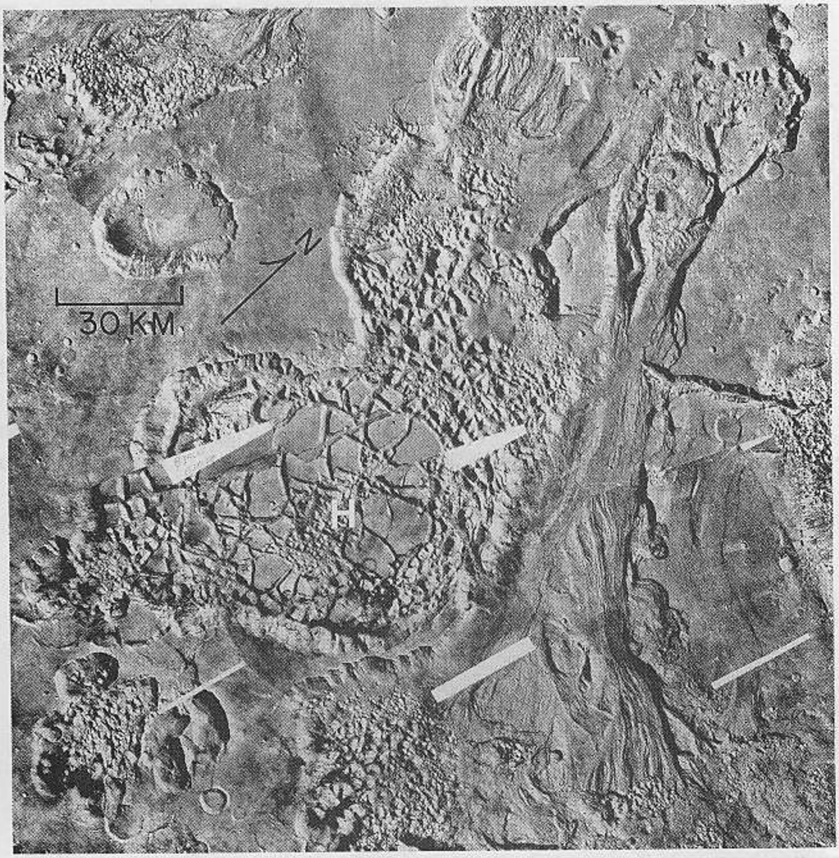

Fig. 1. Chaotic terrain at Hydaspis Chaos (H). The large outflow channel extending northward from the chaos zone is Tiu Vallis (T). (The images are a portion of JPL Viking Orbiter Mosaic 211–5556.)

Macroscale Features

The major elements of outflow-channel patterns are the large-scale flow features (Table I). A regional anastomosing pattern is noted for many of the outflow channels including Ares, Simud, Tiu (Baker and Milton 1974), Kasei and Maja Valles (Baker and Kochel 1979). Other examples of anastomosing or braiding of outflow channels are described by Carr (1974*a*, 1979), Mars Channel Working Group (1983), Komar (1980), Sharp and Malin (1975) and Hartmann (1974).

The issuance of channels from discrete source regions is a feature common to almost all outflow channels. These discrete source regions can take the form of fractures or elongate depressions (Schultz et al. 1979; Wise 1984;

TABLE I

Macro- and Mesoscale Features of Outflow-Channel Morphology and their Consistency (x) or Inconsistency (−) with Hypothesized Genetic Mechanisms[a]

MACROSCALE Channel Pattern	MESOSCALE Bedforms
1. Anastomosing pattern of channels	1. Longitudinal grooves
2. Discrete source areas, such as chaotic terrain	2. Inner channels
3. Residual uplands separating channels	3. Recessional headcuts (cataracts)
4. Many uplands partly smoothed and streamlined by fluid flows, especially on their upstream ends	4. Scabland
5. Pronounced flow expansions and constrictions	5. Scour marks near flow obstacles
6. Distinct upper elevational limit to eroded terrain	6. Pendant forms (small-scale streamlined hills, which may be either residual or depositional)
7. Transected divides and hanging valleys	7. Expansion bar complexes, or fan deltas
8. Erosion of diverse rock types thousands of kilometers from probable fluid source areas	
9. Low sinuosity	
10. High width-depth ratio	
11. Differential erosion of terrain controlled by structure and lithology	
12. Indistinct channel terminus, including the lack of obvious large-scale fans and deltas	

Morphological Features	Wind	Mud Flow	Glacier	Lava	Catastrophic Flood
Anastomosis	?	X	X	X	X
Streamlined uplands	X	X	X	?	X
Longitudinal grooves	X	X	X	?	X
Scour marks	X	?	?	?	X
Scabland	?	—	?	—	X
Inner channels	?	X	?	X	X
Lack of solidified fluid at channel mouth	X	—	X	—	X
Localized source region	?	X	X	X	X
Flow for thousands of kilometers	X	—	X	X	X
Bar-like bedforms	?	?	?	?	X
Pronounced upper limit to fluid erosion	—	X	X	X	X
Consistent downhill fluid flow	?	X	X	X	X
Sinuous channels	?	X	X	X	X
High width-depth ratio	X	X	X	—	X
Headcuts	—	?	X	—	X

[a] Table modified from Baker (1982).

Christiansen 1985; Carr 1974*a*, 1979) or they can be areas of jumbled blocks on the floors of large vaguely circular depressions termed chaotic terrain (Sharp 1973*a*). Chaotic terrain consists of slump and collapse blocks in steep-walled arcuate depressions at channel heads. It is generally assumed that removal of fluid from below the surface causes loss of support, and the material collapses under its own weight, leaving irregular blocks of varying sizes on the depression floor.

Residual uplands separating channels and partially to fully streamlined upland remnants are common in outflow channels (Fig. 2). These residual uplands take two forms: one roughly similar to a single lemniscate loop, usually wider toward the upstream end and tapering downstream narrowing to a point (Baker and Kochel 1978*b*). The other common form is rhombic or diamond shaped. Other macroscale features that have been described include expansions and constrictions to flow (Baker and Milton 1974; Baker and Kochel 1978*a*, 1979; Baker 1982), upper elevational limits on eroded terrain,

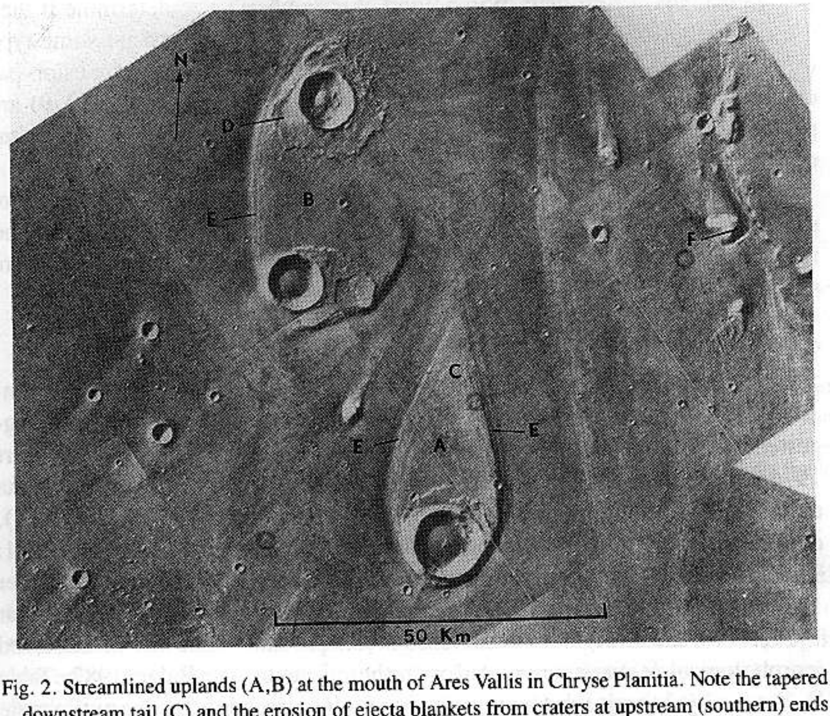

Fig. 2. Streamlined uplands (A,B) at the mouth of Ares Vallis in Chryse Planitia. Note the tapered downstream tail (C) and the erosion of ejecta blankets from craters at upstream (southern) ends of the uplands. An uneroded crater can be compared at (D). The erosive fluid flowed north in massive quantities, scouring bedrock as shown by the terrace-like benches on the upland flanks (E). A prominent crescent-shaped scour hole occurs at the upstream end of the small island at (F). The scarp surrounding the lemniscate upland (A) is about 600-m high, and that surrounding island (B) is about 400-m high. (JPL Viking Orbiter Mosaic 211–4985.)

transected divides, hanging valleys, low sinuosities, and high width-to-depth ratios (Baker and Milton 1974; Baker 1982; Baker and Kochel 1979; Mars Channel Working Group 1983).

Mesoscale Features

A variety of bedforms are located on the floors of the outflow channels (Table I). Inner channels with either residual or recessional headcuts, scour marks around presumed flow obstacles, and longitudinal grooves (Baker and Milton 1974; Baker 1979,1982; Baker and Kochel 1978*a*,1979; Carr 1979, 1986; Carr and Clow 1981; Komar 1983,1984; Lucchitta 1982*a;* Lucchitta et al. 1981; Mars Channel Working Group 1983; Cutts and Blasius 1981; Thompson 1979) which parallel the presumed flow direction are particularly common on the floors of the larger outflow channels (Baker 1978*a*,1982; Baker and Kochel 1978*a*,1979). These grooves converge in areas of channel constriction, and diverge around flow obstacles and in areas of channel expansion (Fig. 3).

Small streamlined features or pendant forms (Baker 1982) are present in many channels. From their morphology, it is impossible to determine if they are residuals similar to the larger streamlined hills, or if they are some type of depositional bar (Baker 1982). Other bedforms, such as expansion-bar complexes and fan deltas (Baker 1982; Baker and Kochel 1978*a*,1979) are probably depositional in nature. The expansion bars appear to have been generated in areas where the flow diverged and the velocity decreased.

On the floor of some channels, irregular bright patches occur that are similar to the scabland topography described by Bretz et al. (1956). Scabland results from erosional plucking wherein pieces of surface material are stripped away due to high-velocity turbulent fluid flow (Baker 1982).

Alternative Hypotheses for Channel Genesis

The morphology and scale of the outflow channels clearly indicates genesis by fluid flows of immense magnitude. Several fluids have been suggested, including liquid hydrocarbons (Yung and Pinto 1978), lavas (Carr 1974*a;* Schonfeld 1976), glaciers or ice streams (Lucchitta 1982*a;* Lucchitta et al. 1981; Lucchitta and Ferguson 1983), winds (Cutts and Blasius 1981), debris flows or mud flows (Nummedal 1978; Nummedal and Prior 1981; Thompson 1979), or cataclysmic water floods (Baker and Milton 1974; Baker 1978*a*,*b*,1979,1982; Carr 1979; Mars Channel Working Group 1983; Komar 1979,1980). Each of these mechanisms can produce some of the observed morphological features associated with the channels (see Baker 1982, Table I), but the hypothesis which most parsimoniously accounts for the entire suite of features is cataclysmic flooding.

Lava erosion was originally proposed as a mechanism of outflow channel genesis because of channel occurrences in volcanically produced terrains and apparent near contemporaneity of channelization and lava emplacement in

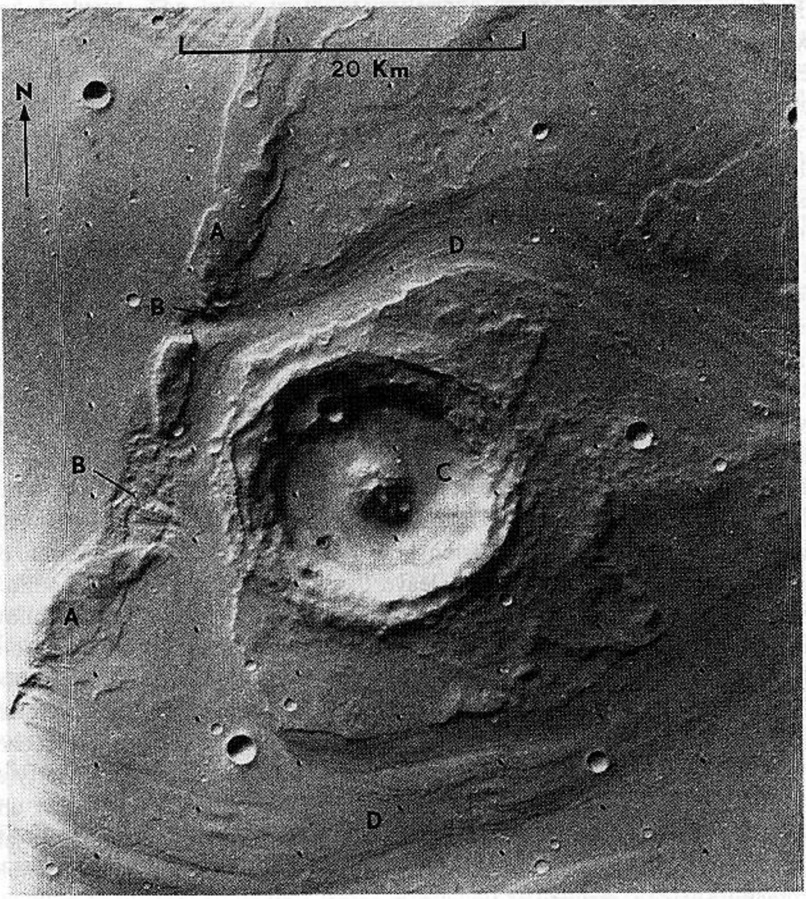

Fig. 3. A portion of Maja Vallis in Chryse Planitia. Fluid flow followed the topographic gradient sloping to the east (right on the photo). Fluid ponded upstream (west) of the mare-like ridge (A) and spilled through gaps (B) as it overflowed low points in the ridge. It then spilled around the crater (C), scouring channels and grooves (D) into the ejecta blanket. (Viking Orbiter frame 20A62.)

some regions. Typical terrestrial lava flows do not produce observed channel bedforms, but low-viscosity flows (Carr 1974*a;* Schonfeld 1976) could move turbulently and erode into underlying and surrounding material by heating it and entraining it in the flow. In theory, low-viscosity lava flows might mimic fluvially produced features. The most obvious advantages of this model include the abundance of volcanic features on Mars, the stability of lava in the current Martian environment, and local channel occurrences in volcanic terrains and even on the flanks of volcanic constructs. Disadvantages of this model include: (1) the lack of large lava deposits at channel terminations; (2)

the fact that lava channels usually have distributaries at their distal reaches which are lacking in outflow channels; and (3) regional anastomosis of the channels is not seen in terrestrial lava channels, since flowing lava, once channelized, tends not to move laterally.

Another proposal is that winds carrying saltating grains may have produced the outflow channels (Cutts and Blasius 1981). Advantages of this aeolian hypothesis include documented aeolian activity in the current environment and the lack of sediment source/sink problems (Nummedal et al. 1983; Baker 1982; Cutts and Blasius 1981; Murty et al. 1984). The source for saltating grains could presumably have been ubiquitous, although at the two Viking Lander sites there was a marked absence of grains in the proper range for saltation (Nummedal et al. 1983). The sinks for the wind-blown sediments could be the great polar ergs, or the material could have been widely dispersed by the transporting winds. Streamlined hills could be analogous to terrestrial yardangs, and longitudinal grooving similar to that within the channels is observed on Earth in desert regions like the Sahara. However, there are many problems with this model. Winds on Earth do not cut channels, and it is difficult to envision how Martian winds could be localized enough to cut a channel and yet leave the surrounding terrain unmodified (Nummedal et al. 1983; Murty et al. 1984). Also, winds do not have to follow topographic gradients, yet the outflow channels all indicate flow directions down topographic gradient. (Baker 1978*a*,*b*,1982; Lucchitta and Ferguson 1983). Finally, features like trim lines and indicated fluid spillage into craters document that the eroding agent had a free upper surface (Baker and Kochel 1978*a*,1979), which is difficult to reconcile with an aeolian origin (Murty et al. 1984; Baker 1982; Nummedal et al. 1983). Although wind is highly unlikely as an originator of outflow channels, secondary modifications to already existing channel features by aeolian processes are much more tenable (Lucchitta 1982*a;* Baker 1982).

A fluid which does have a free upper boundary and may be locally stable under current Martian conditions is ice. Lucchitta and Anderson (1980) proposed that the outflow channels were produced via glacier or ice-stream erosion, and subsequent papers extend this model (Lucchitta 1982*a;* Lucchitta and Ferguson 1983; Lucchitta et al. 1981). These authors demonstrated that glaciers may be stable and move on the surface of Mars, but flow under current conditions might be extremely slow. Additional effects, like the addition of brine (Lucchitta et al. 1981), frictional heating due to travel down a steep topographic gradient (Lucchitta 1982*a*) or subglacial water lubrication (Lucchitta 1982*a;* Lucchitta et al. 1981) are required for the increased-glacial-mobility hypothesis. A change in the Martian climate could also change the velocity at which glaciers could travel down the channels. The advantages of the glacial/ice-stream model are the surface stability of ice for present Martian conditions, the similar scales of terrestrial glacial features and Martian channel features, and similarities of channel bedforms and glacial features. The latter include similarities among streamlined hills and drumlins or Antarctic subglacial forms, longitudinal grooving similar to that seen in areas of Canada, anastomosing patterns, scouring, and U-shaped channel cross sections (Lucchitta 1982*a;* Lucchitta et al. 1981). The difficulties with an Earth-like glacial model include the slow movement of hypothesized glaciers on Mars given current conditions, a lack of cirques at the heads of the channels, the lack of glacial deposits at channel ends, and the difficulty of getting precipitation to produce the glacier on the surface. Lucchitta et al. (1981) and Lucchitta (1982*a*) suggested that the glaciers could have been grown from subsurface water feeding up onto the surface, and that this might also produce the chaotic terrain at the channel heads. Although not strictly glaciers by definition, such seepage-fed ice masses might, in theory, move downgradient and be intimately associated with impounded and released water flows (floods).

Nummedal (1978) and Nummedal and Prior (1981) proposed that mud flows or debris flows could have cut the Martian outflow channels, suggesting that the high-fluid content in Martian surface rocks may lead to induced liquefaction (loss of cohesion) via shock or strain-induced pore-pressure increase followed by rapid flow, as observed terrestrially in Scandinavian quick clays. The chaotic terrain is seen as the equivalent to quick-clay collapse depressions (Nummedal 1978; Nummedal and Prior 1981). Grooving is observed in areas where quick clays flow (Thompson 1979). An interesting related model was proposed (Komar 1979,1980) suggesting an analogy between outflow channels and features observed in submarine turbidity currents and debris flows due to the similar effects of Mars' lesser surface gravity and buoyancy in terrestrial submarine settings. Erosion by turbidity currents on Earth produces features of similar scale to outflow channels as well as many similar bedforms. Debris flows begin their movement as laminar flows, but increase in turbulence as they travel downchannel and lose transported material, producing features consistent with regional anastomosis and partial to fully streamlined residuals downchannel from the sources. In debris flows and mud flows, longitudinal roller vortices can become stable and could produce longitudinal grooves consistent with those observed in channels. The problems with this mechanism are the small size of source regions compared to the amount of mud or debris required to have eroded hundreds to thousands of kilometers, and the apparent lack of vast debris deposits (Baker 1982). Moreover, the high turbulence suggested by scabland-stripped zones, scouring around flow obstacles, and recessional inner channel headcuts are difficult to reconcile with flows saturated or supersaturated by debris and mud (Baker 1979,1982).

The Cataclysmic Flood Model

Baker and Milton (1974) pointed out the many similarities between the Martian outflow channels and the Channeled Scabland in Washington State formed by breakout flooding from prehistoric Lake Missoula. By analogy they proposed that the outflow channels were the products of catastrophic floods of immense proportions. Numerous other workers have since analyzed and expanded upon this hypothesis. Analytic studies have been made of erosion and hydraulics for channelized flows, chaotic terrain, longitudinal grooving, scouring and streamlining of residual terrain. Highly turbulent catastrophic flows are extremely effective erosion agents (Baker 1982). The entire suite of outflow-channel landforms is also observed in the Channeled Scabland, and the scale of the features, although not equal, is certainly closer than that of usual terrestrial fluvial features. Catastrophic floods are characterized by high-velocity, high-density, low-viscosity water flows with large discharges (Baker 1973,1982; Baker and Komar 1987). Usually, catastrophic floods produce channels that are not very sinuous, with high width-to-depth ratios. Features that are produced by catastrophic floods include anastomosis, streamlined remnants, longitudinal grooving, inner channels with recessional headcuts, scouring around flow obstacles, scabland-plucked erosional scars, and many other features similar to those observed in the Martian outflow channels. Komar (1983,1984) argues that some streamlining within the outflow channels could only be produced by water flow.

The advantage of the catastrophic flood model is that all of the Martian channel forms are also seen in analogous terrestrial catastrophic flood regions (Baker and Milton 1974; Baker 1982; Komar 1979). The major difficulties associated with this model include:

1. The general instability of water on Mars' surface under current climatic conditions;
2. The lack of obvious fluvial depositional areas at channel termini;
3. The source areas seem to be too small to account for the amount of water required to fill and erode the outflow channels.

In relation to difficulty (3), Baker (1982) proposed that the chaotic terrain zones probably represent only the final stages of progressive channel growth by headward growth as more and more terrain collapses. Each subsequent collapse releases new water which erodes the downchannel chaotic terrain from an earlier phase of the flood, so that evidence for the early sources of the channel water are obliterated by subsequent outflows down the same channel. Associated ice and debris must have also contributed to the genesis of the evolving channel system. Low-sinuosity channels headed by chaotic-collapse terrain have been modeled by Manker and Johnson (1982) who constructed scaled-down physical models of Martian near-surface materials infused with ground ice. When tilted and heated from below, the models yielded collapse-headed channels of low sinuosity similar to observed chaotic terrain.

Carr (1979) proposed a mechanism for outflow in which confined ground water is released when an aquifer is breached by some rupturing or fracturing event. Masursky et al. (1986*b*) proposed that the water stored as ground ice in Martian regolith might be melted by localized heating due to intrusive magma emplacement and may escape catastrophically creating the channels. Because release processes occurred in the Martian subsurface, these and other mechanisms (Clark 1978; Milton 1974*a;* Peale et al. 1975) are very difficult to confirm with present information.

The problem of maintaining surface-water flows on Mars under present conditions does not seem particularly serious for short-duration floods. Even prolonged flows could be maintained as ice-covered rivers (Lingenfelter et al. 1968; Wallace and Sagan 1979) or seepage flows (Carr 1983). Freezing-point depressants could be present (Ingersoll 1970; Brass 1980). Moreover, the low atmospheric pressure would actually facilitate the effective flood erosional process of cavitation (Baker 1979,1982; Baker and Cotta 1987). Cavitation occurs when dynamic-pressure variations in a flow produce vapor bubbles whose subsequent collapse shatters channel-bed materials which are then removed by the flow (Baker 1979). Scabland etching is probably produced by cavitation and the plucking action of kolks and is controlled by inherent bedrock structure (Baker 1979).

Komar (1979) examined Martian outflow-channel hydraulics by comparison to similar scaled terrestrial features, concluding that outflow channels and the Channeled Scablands were hydraulically similar to terrestrial turbidity currents in terms of scale, velocity, discharge, stress and sediment transport capacity. Blocks over one meter in diameter may be carried in suspension in these flows, demonstrating their enormous erosion potentials. Large amounts of sediment can be transported as wash load in the Martian floods (Komar 1980) allowing very rapid erosion. The paucity of depositional features near outflow-channel mouths and the great extent of channels compared to source-region sizes may both be as a result of extensive wash-load transport (Komar 1980).

Longitudinal grooving seen in Martian outflow channels was analyzed by Thompson (1979) and Baker (1979) who conclude that it was most likely produced by longitudinal roller vortices which are generated in high-velocity flows (Baker 1979) or flows with vertically stratified viscosities (Thompson 1979). Scouring around flow obstacles was studied by Komar (1985) in flume experiments. Scour marks produced around lemniscate-shaped flow obstacles in the flume compare favorably to scour marks observed in outflow channels on Mars. Komar (1985) indicates that flow strengths easily exceed those required to entrain channel sediments.

Cataclysmic flooding also seems consistent with the quantitative details of observed streamlined forms. Flow obstacles are modified into forms which are the least resistant to the flow by a progressive process of erosion and deposition into lemniscate (airfoil) shapes. Minimization of total drag around an obstacle is achieved by a compromise between drag due to an obstacle's form and the friction over its surface (Baker 1979). Komar (1983,1984) examined streamlining of various features such as glacial drumlins, aeolian yardangs and fluvial tear-drop islands. In general, drumlins are more elliptical than riverine lemniscates, whereas yardangs may have much greater length-to-width ratios (although these range widely) compared to those produced fluvially. Comparisons between Channeled Scabland and lemniscate forms in Martian outflow channels show good agreement, suggesting that streamlined residuals within Martian outflow channels were produced by a high-velocity, very turbulent water flow.

Because of the difficulties in explaining all the complexities of outflow channels with a single, Earth-based analogy model, it may be that the unique Martian environment modified flow processes. Conditions of enhanced wash-load sediment transport (Komar 1980) plus ice formation, with ice flowage (Lucchitta 1982*a*), may have acted in complex combination. These events could even have produced interlayered ice-rich debris flows in troughs subsequently invaded by lava and modified by wind deflation.

Fate of Flood Water and Sediment

The large channels around Chryse and northwest of Elysium debouch onto the low-lying northern plains, and all traces of the channels are lost between latitudes 45°N to 65°N. The plains in these areas display a variety of distinctive features that have been attributed to the presence of ground ice (Carr and Schaber 1977; Rossbacher and Judson 1981; Lucchitta 1981). The most pervasive and striking characteristics are a widespread polygonal pattern of fractures and a mottled appearance caused by bright crater ejecta superimposed on a dark surface. McGill (1985) and Lucchitta et. al.(1986) suggest that the distinctive features of these areas are caused by the presence of sedimentary deposits from the large floods. Parker et al. (1989) describe the regional extent of northern plains features that might reflect the accumulated ponding of outflow sediment and water discharges.

It seems clear that the floodwaters that cut the outflow channels must have pooled in low areas at the ends of the channels. If, when the floods occurred, climatic conditions were the same as at present, then the pooled water would have immediately frozen over to form an ice-covered lake. Sediment would have settled out and freezing would have continued until an ice deposit was left over frozen sediments. The amount of water that would have sublimed into the atmosphere during the flood and while the terminal lake was freezing was probably trivial compared to the size of the flood (Carr 1983). Recent modeling (Carr 1990) indicates that the ultimate fate of the water from these high-latitude ice deposits would depend largely on whether the deposit became covered with debris. If the ice were continually swept free of debris then the ice would slowly sublime into the atmosphere, and the water would be trapped at the poles. On the other hand, if the ice became covered with a few meters of material, then at these high latitudes the ice would be permanently stable and the ice deposit would remain for the life of the planet (Farmer and Doms 1979; Zent et al. 1986). The recurrence of dust storms, the likelihood of a sediment-lag accumulation on the ice, and photogeologic evidence of the accumulation of debris at high latitudes (Soderblom et al. 1974), all suggest that the ice deposits would have quickly become covered with debris and stabilized. A similar reasoning applies to the deposits at the ends of the large channels in Hellas. In contrast, the terminal ice deposits from those channels, such as Mangala Vallis, that end at low latitudes would have been permanently unstable. The ice from these deposits would have continued to sublime into the atmosphere until all the water had been lost.

III. MARTIAN VALLEYS

Martian valleys are distinguished from channels by the absence of bedforms which are direct indicators of fluid flow (Mars Channel Working Group 1983). Although the valleys may contain channels, only in rare instances can the latter be detected with the resolution of the existing imagery. Most (perhaps 98%) of the valleys are located in the old cratered-terrain regions. Younger valleys are located on the south wall of Ius Chasma and on the flanks of some Martian volcanoes. Valleys are concentrated in the 65°S to 65°N latitude belt, and decrease in number toward the high latitudes. Because most of the valley networks are located only in the oldest terrains, the valley networks themselves are thought to be old (Pieri 1976; Carr and Clow 1981), with formation ceasing just after the end of heavy bombardment, approximately 3.8 to 3.9 Gyr. Alternative explanations for the almost exclusive formation of valleys in the older terrains are that these regions were more easily eroded by valleys and that water was preferentially located in the ancient terrains (Carr 1984). A notable group of young Martian valleys is located on Alba Patera (Gulick and Baker 1989, 1990). These valleys formed well after the period of heavy bombardment and thus have important paleoclimatic implications for Mars. Many origins which are similar to those proposed for the outflow channels have also been considered for the formation of the Martian valleys, including wind, water, lava and volcanic density currents. As in the case of the outflow channels, erosion by running water is considered to be the primary mechanism responsible for the formation of valley networks (Mars Channel Working Group 1983).

General Morphology

The Martian valleys have widths ranging from < 1 km to nearly 10 km and lengths ranging from < 5 km to nearly 1000 km (Mars Channel Working Group 1983). Unlike the channels, valleys exhibit a wide variety of drainage patterns ranging from well-integrated to mono-filament networks. Cross valley profiles range from V-shaped to U-shaped to valleys with steep, nearly vertical walls and broad, flat floors. Although there is a wide variety of valley types, most individual valleys have some characteristics in common. Upper reaches frequently appear degraded and have higher drainage densities than the lower reaches (Baker and Partridge 1986). In general, drainage densities for the Martian valley networks are much lower than those for terrestrial networks (Pieri 1980*a*). Notable exceptions are the fluvial valleys located on the Martian volcanoes Ceraunius Tholus, Hecates Tholus and Alba Patera. These valleys are morphologically similar and have similar drainage densities to those formed on the Hawaiian volcanoes (Gulick and Baker 1986). Another characteristic common to most of the valley networks is that tributaries commonly have blunt, theater-headed terminations. Networks have numerous stubby first-order tributaries. Drainage patterns often are structurally controlled (Schultz et al. 1982; Gulick 1986), with valleys commonly following fractures and other linear features. Interfluves remain largely undissected, reflecting the inefficiency of valley networks at filling space within their own drainage basin (Pieri 1980*a,b;* Baker 1982). Unlike most terrestrial river valleys, tributaries often appear to be as deep as the main trunk. Most Martian valleys have steep sidewalls with a relatively constant downvalley width. Tributaries join main trunk valleys at low mean junction angles when compared to terrestrial drainages (Pieri 1980*a*).

Martian valleys have been classified in many different ways: by planimetric pattern (Pieri 1980*a*), by size and general location (Baker 1982) and by combined planimetric patterns/cross valley profiles (Brakenridge et al. 1985). For the sake of discussion, we will use a modified version of the classification scheme presented in Baker (1982).

Valley Types

Longitudinal Valley Systems. Longitudinal or elongate valley systems (Fig. 4) are by far the largest of the valley networks. Examples of this type are Nirgal, Nanedi, Bahram, Ma'adim and Al Qahira Valles. These valleys are typically several hundreds of kilometers in length and several tens of kilometers in width. Upper reaches usually have short theater-headed tributaries with the exception of Ma'adim Vallis which has long branching tributaries. Lower reaches are usually sinuous and have broad, flat floors. Valley walls in the lower reaches have a scalloped appearance which is attributed to subsequent modification by landsliding, undercutting and other mass-wasting processes (Baker 1982). There is a general lack of dissection of adjacent uplands by these valley systems which are usually located in the old cratered uplands region. It is thought that these valleys may have been initiated as small valleys and then became enlarged by wall retreat as lower courses became deeply incised (Baker 1982). The undissected nature of the interfluves and the short accordant tributaries with width-to-depth ratios equalling that of the main trunk are thought to provide strong arguments against a formation by direct rainfall and surface runoff relationships (Pieri 1980*b;* Baker 1982).

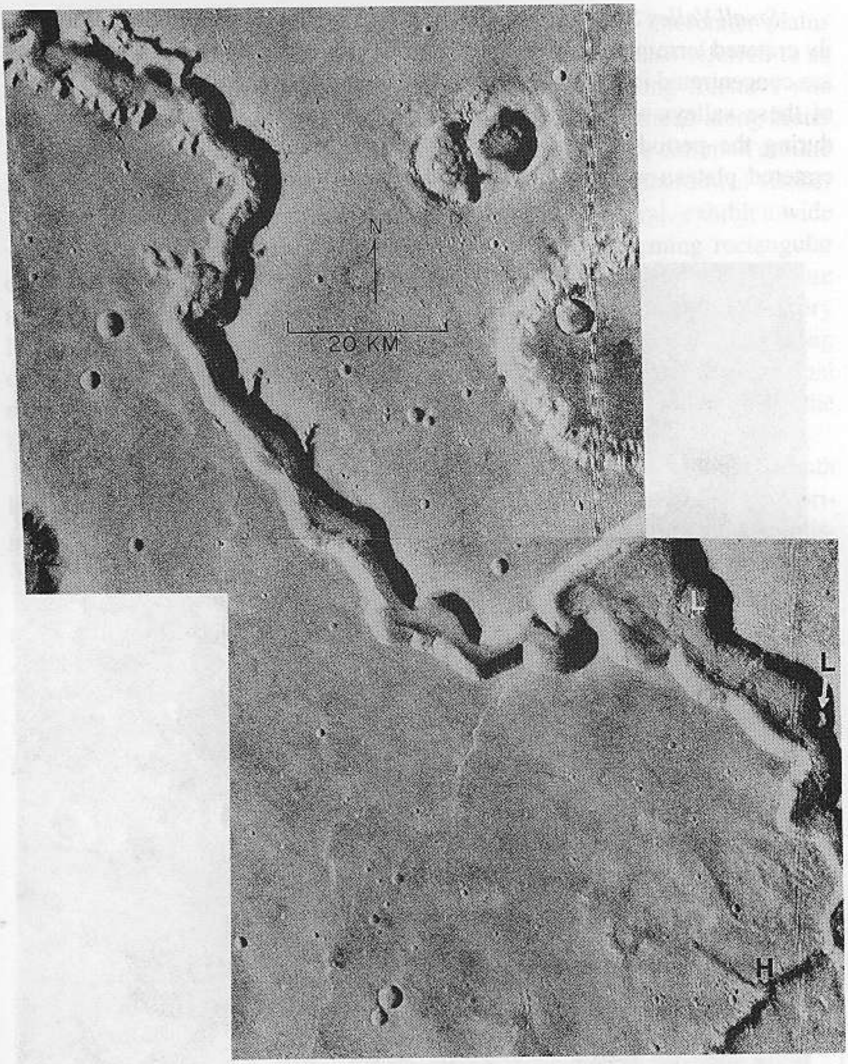

Fig. 4. Downstream portion of Nirgal Vallis, an example of a large longitudinal valley system. Note the short, stubby tributaries upstream (northwest) and the irregular widening downstream. Landslide deposits (L) and a hanging valley (H) also are present. (Viking Orbiter frames 466A61 and 466A64.)

Small Valley Systems. Small valley systems are ubiquitous in the heavily cratered terrain. However, an unusually large number of small valley types are concentrated in the Margaritifer Sinus region (Grant 1987) (Fig. 5). Many of these valleys are located in the hilly and cratered terrain, which formed during the period of heavy bombardment (Carr 1984), and drain into the cratered plateau region. The cratered plateau consists of smooth intercrater

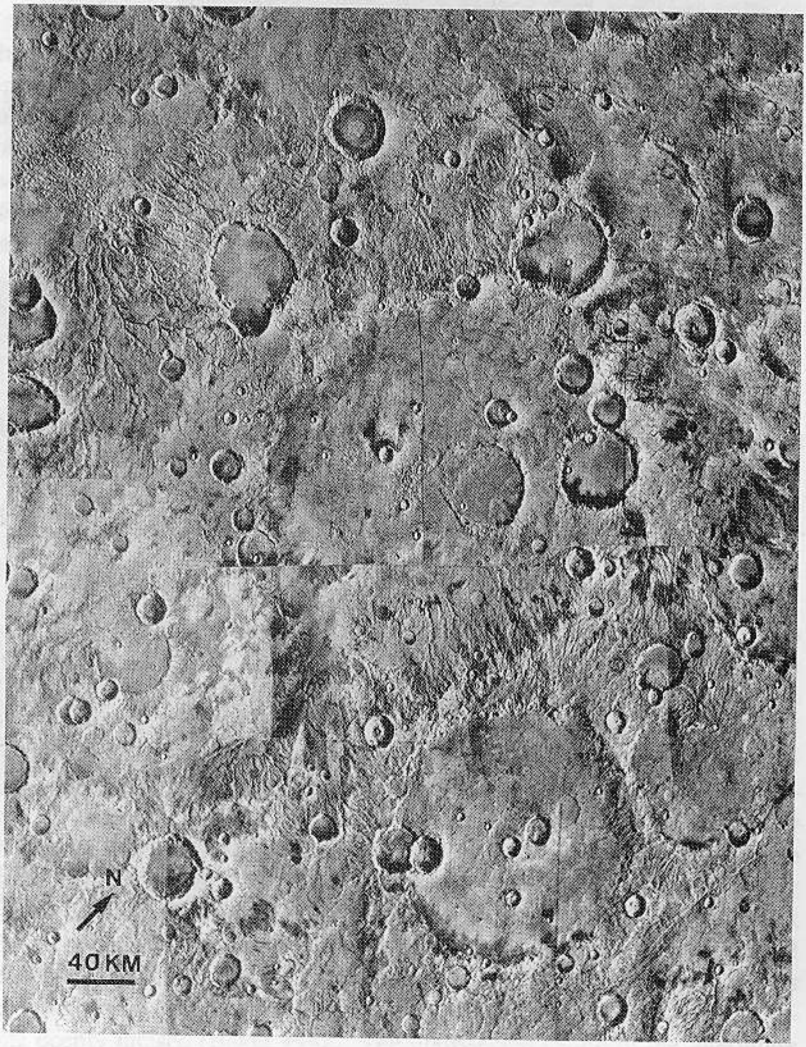

Fig. 5. Small valleys in the heavily crater terrain. Mono-filament valleys radiate out from the crater rim. Integrated systems are located adjacent to crater in the intercrater terrain. (The images are a portion of JPL Viking Orbiter Mosaic 211–5207.)

plains which postdate the hilly and cratered terrain. The intercrater plains have buried or partly buried some craters. Linear valleys, also referred to as endogenic depressions (Baker 1982), are concentrated along fractures and drain into linear troughs. Small parallel valleys are concentrated along crater rims suggesting outward drainage of fluids. These valleys are common around large craters and clearly postdate the period of heavy bombardment (Baker 1982). Small valleys in the heavily cratered terrain, in general, exhibit a wide variety of morphologic patterns (Pieri 1980*a*). Valleys forming rectangular drainage patterns are structurally controlled, probably by underlying fracture systems. These valleys generally have low tributary junction angles and very low drainage densities. Monofilament parallel valleys are concentrated along crater rims and have little or no tributary development. Digitate to transitional dendritic valleys, which are more similar to terrestrial river valleys, have the highest drainage densities in the heavily cratered terrain.

Small valley systems also dissect some Martian volcanoes (Fig. 6). Both lava (Carr 1974*a*) and volcanic density-flow (Reimers and Komar 1979) origins have been proposed for the volcano valleys. Recent studies (Mouginis-Mark et al. 1982*b*,1988; Wilson and Mouginis-Mark 1987; Gulick and Baker 1990, 1991) show that some valleys are probably fluvial. The morphology of most valleys present on the volcanoes Hecates Tholus, Ceraunius Tholus and Alba Patera is compatible with a fluvial origin (Gulick and Baker 1989, 1990). These valleys formed in regions of subdued lava-flow morphology where the surface appears to be composed of fine-grained sediments, probably ash. They are inset into the surrounding land surface, widen slightly in

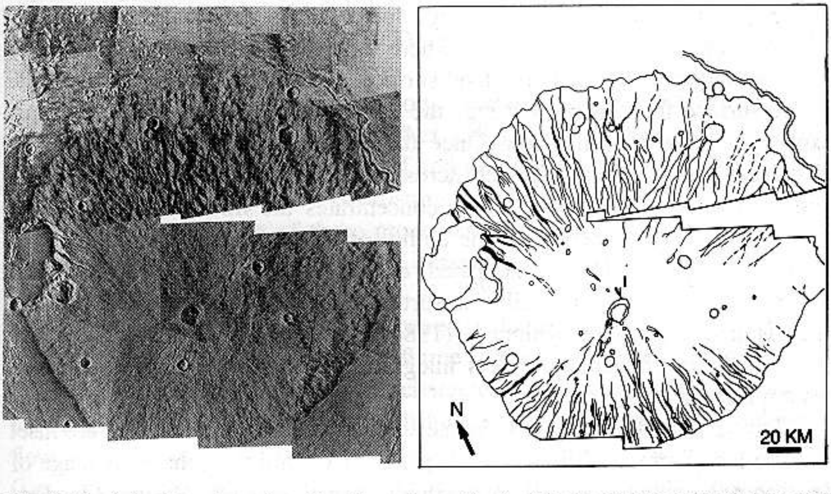

Fig. 6. Small fluvial valleys dissecting the volcano Hecates Tholus. Heavy black line denotes areas of valley enlargement, probably by sapping. The figure is taken from Gulick and Baker (1989*d*). (The images are a portion of JPL Viking Orbiter Mosaics 211–5601 and 211–5787.)

the downstream direction, form integrated tributary networks, and have drainage-density values which are comparable to those of fluvial valleys on the Hawaiian volcanoes (Gulick and Baker 1989). Valleys are also present on the volcanoes Hadriaca Patera, Apollinaris Patera and Tyrrhena Patera. The morphology of valleys on these particular volcanoes suggests that a combination of several genetic processes (water, volcanic density flows and lava) may have been important in their formation (Gulick and Baker 1990). However, extensive subsequent modification of these valleys by one or more of these processes has obscured the original morphology.

Other small valleys, referred to as slope systems by Baker (1982), formed along the walls of Valles Marineris in the Ius Chasma region. These valleys display primitive, angular, dendritic-drainage patterns. The apparent structural control of these patterns in addition to the theater-headed tributary terminations provide a strong argument for a sapping origin (Kochel et al. 1985).

Local Genetic Processes

Although it is widely accepted that the Martian valleys formed by fluvial processes (Mars Channel Working Group 1983), the way in which fluvial processes form valleys has a profound impact on the resulting morphology. Runoff valleys form from the transport of flowing water across the surface, while sapping valleys form from the outflow of ground water into the surface environment. Both processes exhibit unique morphologic characteristics. Below is a brief summary of how valleys are formed by surface runoff and ground-water sapping.

Surface Runoff. Fluvial valleys and channels, resulting from surface-runoff processes, form in locations where water runoff concentrates and flows with enough force to erode the land surface. In order to have water available at the surface for channel cutting, the rate of runoff over the surface must exceed the rate of infiltration. Once these conditions are met, water from dispersed source areas migrates across the surface and collects in topographically low areas. This action concentrates the flow of water moving downslope, thereby increasing the ability of water to erode the surface and establish a channel. Once established, channels provide a locus for the convergence of surface flow leading to further incision. This process is particularly well described by Knighton (1984). As drainage of the land surface becomes more efficient, channels integrate and eventually form a complex system of networks.

Fluvial runoff valleys and channels are erosional features which are inset into the land surface. Tributaries are present except during the early stage of development where channels form simple, trough-shaped patterns (Macdonald et al. 1983). In plan view, these drainages have tapered tributary heads, a continuous valley form and an increasing valley width in the downstream direction.

Ground-water Sapping. Sapping valleys are formed by the undermining of rock and sediment by ground-water outflow. In the model of sapping valley network formation proposed by Dunne (1980), subsurface water emerging from valley sides and headwalls disrupts ground-water flow patterns. Ground water converges and concentrates along weakness planes removing material which supports the overlying surface. This removal of support eventually results in the collapse of overburden into the valley, causing valley sides to widen and the headwall to retreat. The rates of headward erosion in these valleys are faster than the rates of valley widening because the headwall is the region of greatest flow convergence. The process is self-enhancing in that the greater the amount of erosion that takes place, the more flow convergence is produced and the higher the rate of retreat of the headwall. This process eventually halts when the flow of ground water becomes insufficient to maintain further sapping.

Sapping valleys have theater-headed tributaries (Baker 1982) and anomalously large valley width-to-depth ratios when compared to runoff valleys. Because ground-water flow tends to exploit planes of weakness in the terrain, sapping valleys will often form along fractures and joints in the bedrock, resulting in a rectilinear or structurally controlled pattern. Sapping processes will often modify the morphology of runoff valleys which have downcut into the ground-water system. This modification results in valley enlargement in which the V-shaped, cross-valley profiles typical of runoff-dominated systems on volcanic landscapes are eroded into broad U-shaped or flat-floored valleys with steep walls. The overall effect of sapping modification to an existing drainage system is to change the network pattern to one that is simpler and less integrated. Baker (1990) reviews the origin of valleys by sapping.

Numerous comparative planetology studies (Laity and Malin 1985; Kochel and Piper 1986; Gulick and Baker 1987*b*,1989,1990) have recognized the importance of both surface runoff and ground-water sapping processes in the valley development and have distinguished the morphologic characteristics resulting from each of these processes. In addition to these field-based studies, a fairly extensive series of experimental flume studies (see, e.g., Kochel and Piper 1986; Kochel et al. 1985,1988) of drainage networks formed by sapping processes have been conducted in soft and weakly consolidated, layered sediments. These flume studies have been able to replicate much of the sapping morphologic characteristics and controls in drainage networks, including, theater-headed tributaries, joint control of ground-water flow, subsequent orientation of drainages and the importance of lithology (i.e., presence of permeable and impermeable strata) in controlling ground-water sapping processes. The results of these experimental studies are being used to develop numerical simulation models of the sapping process (Howard 1988; Howard and McLane 1988).

Regional Genetic Conditions

While the morphology of the large, longitudinal valleys is generally consistent with a sapping origin, the genesis of the small valleys is less clear. Valleys in the heavily cratered terrain exhibit a complex morphology (Baker and Partridge 1986). These valley systems have relatively fresh-appearing, low-density network segments in their lower reaches and higher-density network segments in their upper reaches. Baker and Partridge (1986) concluded that the dense upper networks formed during the period of heavy bombardment and the pristine lower reaches formed after the emplacement of the intercrater plains. The pristine valley segments subsequently evolved by headward growth into the regions occupied by the degraded valleys. Baker and Partridge suggested that the relatively higher-network density of the degraded valleys may imply a greater influence of surface runoff processes in their genesis.

A similar complex morphologic pattern is exhibited by the fluvial valleys developed on older Martian volcanoes (Gulick and Baker 1990). Fluvial valleys present on volcanoes Ceraunius and Hecates Tholi, which formed during the period of heavy bombardment (Barlow 1988*c;* Neukum and Hiller 1981), exhibit a parallel drainage pattern of integrated valley systems. Some valleys on Ceraunius appear pristine and less integrated while others appear degraded and better integrated. The pristine valleys appear re-activated. Valleys on Hecates also exhibit a parallel drainage pattern, but the valley systems are better integrated than those on Ceraunius. Unlike the valleys on Ceraunius, however, only the lower reaches of some valleys appear pristine and enlarged. Fluvial valleys developed on Alba Patera, a volcano which formed well after the period of heavy bombardment (Barlow 1988*c;* Neukum and Hiller 1981), exhibit only the better-integrated upper-reach morphology (Fig. 7). Indeed, these valleys are the most integrated and most terrestrial-like fluvial valleys on the surface of Mars. The Alba valleys have tapered tributary heads that blend in gradually with the surrounding landscape and V-shaped cross-valley profiles. The Alba valleys appear morphologically similar to those formed by rainfall-runoff processes on the Hawaiian volcanoes. In addition, this same complex morphologic pattern of better-integrated upper reaches and less-integrated lower reaches is also apparent on the Hawaiian volcanoes. Recent morphologic studies of the Hawaiian valleys (Gulick 1987; Gulick and Baker 1987*b;* Kochel and Piper 1986) concluded that Hawaiian valleys formed initially by surface runoff and were subsequently enlarged by ground-water sapping. Based on comparison morphologic studies with the Hawaiian valleys, valleys on the Martian volcanoes probably evolved in a similar manner (Gulick and Baker 1990).

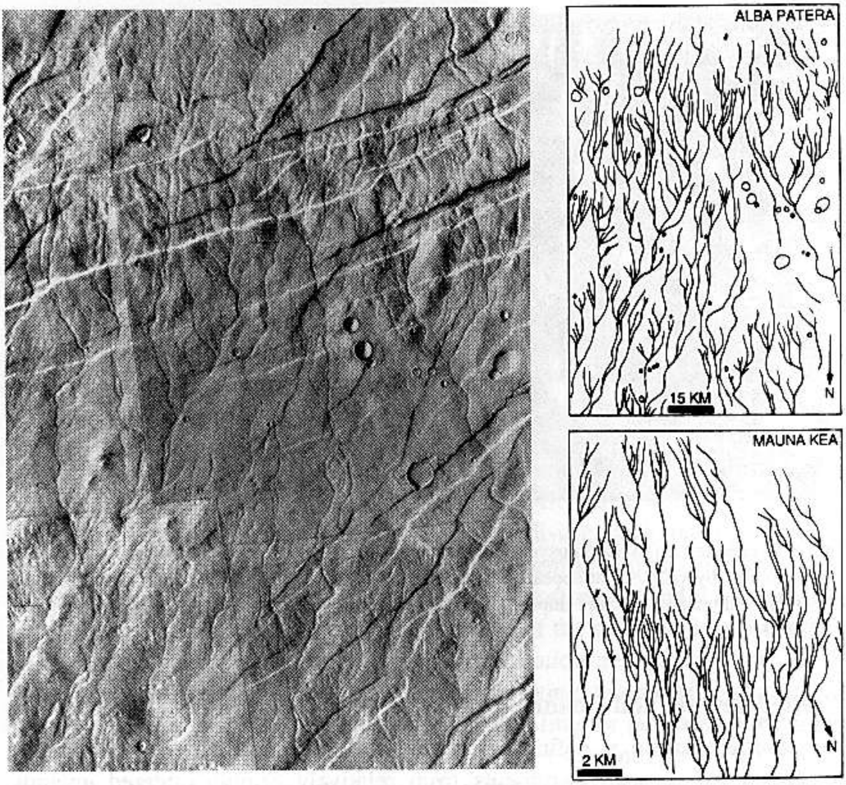

Fig. 7. Fluvial valleys on the northern flank of Alba Patera. Valleys exhibit a well-integrated drainage pattern which is morphologically similar to fluvial valleys formed by surface runoff on the Hawaiian volcanoes (e.g., Mauna Kea volcano, Hawaii). The Alba valley systems are the best developed and the most terrestrial-like fluvial networks on the surface of Mars. The figure is taken from Gulick and Baker (1990). The images are from Viking Orbiter Mosaics MTM 45107.)

IV. RELATED LANDFORMS

Erosional processes modified all varieties of Martian landforms, including channels and valleys. Mass wasting, aeolian, thermokarst and periglacial processes have subsequently acted to modify the walls and floors of channels and valleys originally carved by fluvial processes. The most extreme modification processes may have formed the fretted channels and fretted terrain (Fig. 8). The alteration of the original landforms has been so complete, however, that it is not known if the fretted channels formed by modifying preexisting valleys or not. In this section, we first examine the fretted channels and terrain and then consider the other secondary processes which have modified the channels and valleys.

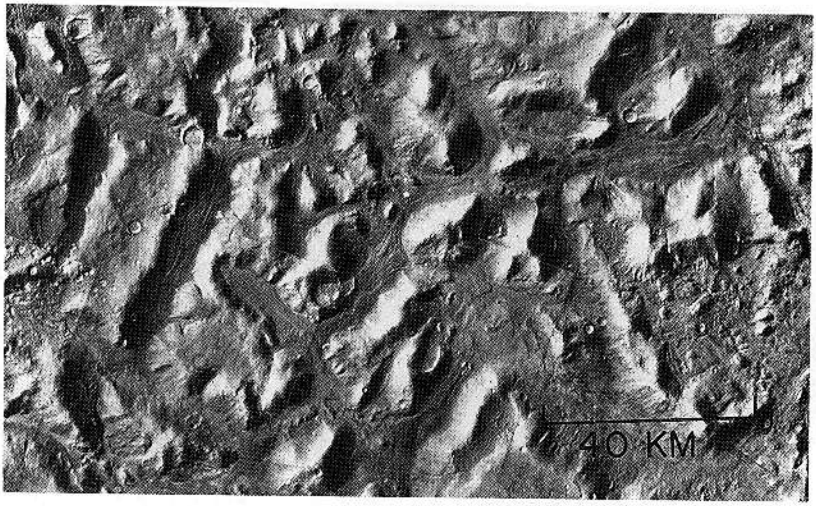

Fig. 8. Fretted terrain in the Nilosyrtis region near 34°N, 282°W. The isolated upland massifs are surrounded by valleys that appear to be filled with debris derived from adjacent slopes. Subparallel ridges and grooves imply flowage of the debris. (JPL Viking Orbiter Mosaic 211–5207.)

Fretted Terrain and Channels

Fretted terrain is defined as a complex of smooth, flat lowland areas separated by abrupt escarpments from relatively heavily cratered uplands (Sharp 1973*a*). The scarps are typically 1 to 2 km high and may indicate the depth of a geologic unit, perhaps the megaregolith or the transition to ice-rich permafrost (Sharp 1973*a;* Davis and Golombek 1989; chapter 16). The terrain is best developed along the cratered upland and northern plains boundary (Mutch et al. 1976*c*). Fretted channels are steep-walled valleys with wide flat floors and indented walls.

Both the fretted terrain and channels exhibit complex planimetric patterns with isolated butte and mesa outliers (Baker 1982). The channels exhibit structural control but no fluid flow bedforms (Baker 1982). The fretted channels are generally located in the region of the fretted terrain but occasionally extend hundreds of kilometers back from the main scarps into the older heavily cratered terrain (Baker 1982). The geomorphology of the fretted terrain is discussed in detail by Kochel and Peake (1984).

A common aspect of both the floors of the fretted channels and the lowlands of the fretted terrain is the presence of debris flows (Carr and Schaber 1977; Squyres 1978). In the fretted terrain, debris flows emanate from the scarps and flow onto the surrounding lowlands, often forming aprons around isolated mesas. The lobate debris flows characteristic of the fretted terrain are found in two planet-wide 25° latitude swaths centered at 40°N and 45°S

(Squyres 1979*a*). When unconfined, these flows may reach lengths of 10 to 20 km. There may be a correlation between scarp face orientation and apron width (Eppler and Malin 1982; Kochel and Peake 1984). The floors of some of the fretted channels are also covered by lineated deposits which appear to be debris flows. The debris aprons generally have a convex profile (Squyres 1978), although some channel and outlier escarpments exhibit a moat-like depression or swale on the apron at the base of the escarpment (Weiss and Fagan 1982). In a crater-counting study, Squyres (1978) found that both lineated and unlineated debris are younger than the uplands, escarpments and surrounding lowland. This may imply that the debris-forming process may still be operating today. It is unknown whether mass wasting and debris flowage are secondary-modification processes acting on features formed by another mechanism, or whether they are the primary mechanism that formed the fretted terrain.

The best evidence that flow of debris indeed took place is the presence of numerous striae in the debris deposits. In the fretted terrain the striae are oriented at right angles to scarp faces, and they diverge and converge around obstacles. These characteristics are generally accepted as implying flow from the scarp faces. In the fretted channels, where the striae typically run parallel to the channel length (Fig. 8), the interpretation of the striae is more problematic. Squyres (1978) interpreted the striae orientation as implying flow from the channel walls meeting in the center, the resulting compression-producing striae parallel with the walls. Squyres states that little or no downvalley flow has taken place. Lucchitta (1983*a*) argued against this interpretation, noting that side-canyon debris bends downvalley, that striae split at obstacles, and that striae are always oriented downvalley, even at the base of mesas where right-angle-oriented flow would be expected. The main problem with Lucchitta's interpretation of considerable downvalley flow is the lack of depositional features at the mouths of most channels.

Fretted channels typically extend into plateaus and have stubby tributaries. They are best developed along the boundaries of the northern terrain and other older elevated terrains. There is typically a gradation from channels cutting plateau to fretted terrains to plains with plateau outliers (Carr 1981). As distance from the cratered uplands and northern plains boundary scarp increases, the number and size of the outliers becomes smaller while the areal volume of the plains material increases (Kochel and Peake 1984). The similarity of the channels and terrain led Baker (1982) to conclude that both were formed by the same erosional process. Under such a scenario the channels would erode back into escarpments along zones of least resistance. As the channels widened and dissected the highlands, the landscape would eventually take the form of the fretted terrain. Kochel and Peake (1984) also concluded that progressive degradation and retreat of the boundary scarp was responsible for the fretted terrain. While an appealing concept, no thorough study of this fretted-terrain degradation process has yet been made.

Beyond the questions of downvalley transport and fretted-terrain evolution is the simple question of what caused the scarp-forming and debris-transport process in the first place. Mass-wasting processes probably undermined escarpments and provided materials which flowed down onto the surrounding lowlands. Sharp (1973*a*) proposed dry sapping as the principle mass-wasting process. Dry sapping proceeds as exposed ice is sublimated from scarp faces. As ice is lost, support is removed from overlying material which then collapses. Squyres (1978) proposed more conventional mass-wasting processes in which material was supplied from eroding angle of repose scarps. Both processes require an initial cliff or valley to provide the face from which the ice sublimates or mass wasting occurs. Wet sapping is another possibility that cannot be ruled out (Sharp and Malin 1975; Baker 1982). Resolution of this question will probably require much higher-resolution imagery and topography, or actual field reconnaissance.

Once the debris material has been wasted off the surrounding scarps, it must then be transported away from the scarp and perhaps down the valley. This transport removes the talus and allows mass wasting from the scarp to continue (Carr 1984). Most authors agree that ice facilitates the flow, but there is less agreement on the source of the ice. Squyres (1978) favors seasonal frost deposits being covered over by subsequent mass wasting. This process would result in interstitial ice similar to that in terrestrial rock glaciers (for discussions of terrestrial rock glaciers, see the collection edited by Giardino et al. [1987]). Lucchitta and Persky (1982) argue instead that the ice contained in the ground over which the debris flows is sufficient to lubricate the transport. The problem with this mechanism is that, in similar situations on Earth, a basal shear of approximately 1 bar is required to initiate the flow. On Mars this requires flows 2 to 5 km thick (Lucchitta 1983*a*), which is thicker than observed. However, the terrestrial criteria may be inapplicable because of the long time scale needed to maintain Martian flows (Lucchitta 1983*a*). Another possible mechanism for removing debris from scarp bases is weathering followed by aeolian deflation (Squyres 1978). Finally, the most provocative mechanism suggested for escarpment formation and debris removal is wave action at the shores of an ancient Martian ocean (Parker et al. 1989).

The fretted channels and terrain may represent important examples of an ongoing Martian geologic process. Lucchitta (1983*a*) notes that, in those latitudes where fretted terrain is located, the current climate allows for the retention of ice in the ground. Thus, ground ice is present to be exposed at scarps for dry sapping and to lubricate movement of ice-rich materials. At high latitudes with colder temperatures, it may be too cold for rock glaciers to be mobilized. Since the observed fretted terrain corresponds with those regions where present conditions allow ice both to be retained in the ground and to lubricate the flow of material, fretting may still be occurring.

Since the channels seem to be modified at low temperatures by a process which is probably associated with ice, the fretting process may be an archetypical Martian mechanism. As such, the process and the associated features deserve more study. Specifically, the mass-transport process and the question of whether downvalley transport is operating should be resolved. If these processes become better understood, they may help provide a record of the recent Martian climate. High-resolution imagery of the escarpments, tributary junctions, valley mouths and topographic information is needed in order to answer such questions.

Other Secondary Processes

Once a valley or channel has been formed, it represents a system of steep walls and valley floors which other secondary geologic processes can exploit. The results of secondary processes can be found in and along most channels and valleys. Wind, for example, can act as an erosive agent by transporting saltating particles. While inadequate to actually carve channels or valleys, aeolian processes can subsequently modify the landforms. Valleys can also provide sheltered areas for deposition by wind and the formation of dunes. In the mid to low latitudes, dunes are present mainly in the protected areas of canyons and valleys (Carr 1984). Another secondary process, mass wasting, removes material from scarps and slopes under the influence of gravity. Most slopes surrounding channels and valleys have probably been modified by these processes, as they have been on Earth. Sapping and hillslope-retreat processes may have enlarged the width of the outflow channels (Baker and Kochel 1979). Such processes are important, for if they are not taken into account when considering channel volume, anomalously large quantities of water may be derived as necessary for channel formation (Baker 1982). Mass wasting is also important in the formation or modification of the fretted landforms.

Processes which depend on the melting of ground ice can also modify channels and valleys. The very large landslides found around the margins of some of the outflow channels were probably lubricated by ground ice-derived water (Mars Channel Working Group 1983). Thermokarst topography forms when melting of ground ice produces local collapse depressions which should not be confused with channels or valleys. Chapter 16 discusses these processes in more detail.

V. IMPLICATIONS FOR PALEOENVIRONMENTAL CHANGE

Global Martian Water Budget

The channels provide a crude way of estimating the total amount of water outgassed from the planet (Carr 1986,1987). The calculation utilizes the volumes of material removed to form the various canyons, channels and chaotic terrain around the Chryse basin. Ignoring the volumes of Ius and Coprates canyons, in which faulting is most evident, then there is a negative regolith volume of roughly 5×10^5 km^3 in the remaining canyon, channels and chaotic terrain. If this volume was all removed by water, and the water carried its maximum sediment load, or 40% by volume (Komar 1980), then 7.5×10^6 km^3 of water would be required, which is the equivalent of 50-m spread over the entire planet. Clearly there is considerable uncertainty in this number since it is unknown precisely how much of the negative volume of the canyons is due to erosion and how much is due to tectonic forces. Moreover, it is unlikely that all the water carried the maximum sediment load.

Further assumptions are required to estimate how much water has outgassed from the planet. The planet-wide distribution of valley networks (Pieri 1976; Carr and Clow 1981) suggests that, early in the planet's history, water was distributed globally and not concentrated in the local areas such as around Chryse Planitia. The concentration of floods around Chryse probably represents later ground-water accumulation in the region as a result of slow migration of ground water from the surrounding higher terrains (Carr 1979). The drainage basin around Chryse Planitia can be roughly outlined from drainage patterns and topography. It constitutes roughly one sixth to one eighth of the planet's surface. Thus, assuming that water was initially distributed evenly over the planet, and that the Chryse drainage basin contained at least 50 m of water, the indicated global inventory is at least 300 to 400 m. This is a very rough estimate. It would be too high if the same water passed several times through the circum-Chryse channels. However, most of the circum-Chryse outflow channels formed after the period of intense valley formation for which warmer climatic conditions have been proposed. Climatic conditions during formation of the outflow channels are likely to have been similar to modern conditions, which cause a thick, planet-wide permafrost. As described above, the water that eroded the channels probably formed permanent ice deposits in the low-lying northern plains, and was not recycled through some global aquifer system. The estimate of 300 to 400 m may also be too low because it ignores the ground water within the Chryse drainage basin that failed to reach the surface. For comparison, the Earth is estimated to have outgassed 3 km of water (Turekian and Clark 1975).

From the volume of volcanic rocks that have accumulated on the surface, Greeley (1987) estimated that the planet has outgassed approximately 50 m of water in the last 3.8 Gyr. This number is entirely consistent with the 500 m estimated (Carr 1987) for the total outgassed water since most of the outgassing is believed to have occurred very early in the history of the planet, before the geologic record was retained. Most geochemical estimates (summarized in Pepin 1987*a*) are lower than both the Greeley and Carr estimates but can be reconciled if Mars lost part of its early atmosphere by impact erosion (Cameron 1983; Melosh and Vickery 1989) or hydrodynamic escape

(Hunten et al. 1989). The volatile inventory and evolution is reviewed in chapters 4, 6 and 32.

Paleoclimatic Implications

The valley networks have been widely cited as evidence of a former thicker atmosphere and surface temperature substantially warmer than those at present (see, e.g., Sagan et al. 1973*a;* Masursky 1973; Pollack 1979; Toon et al. 1980; Pollack et al. 1987). However, the climatic conditions required to form channels and valleys are unclear. Outflow channels could probably form under present climatic conditions. They involve floods of such magnitude that the amount of freezing under present climatic conditions would be trivial (Lingenfelter et al. 1968; Carr 1979). Cold climatic conditions with temperature well below freezing may even be required for the formation of those floods caused by eruption of ground water. A thick permafrost may have been needed to contain the ground water, thereby allowing artesian pressures to build and eventually cause massive outflows (Carr 1979).

The valley networks are much smaller than the flood outflow channels and are presumed to have formed from correspondingly smaller discharges. Moreover, the valleys divide into smaller valleys upstream. If temperatures were well below freezing, it is reasonable to assume that, irrespective of the source of the water, small streams in the distal parts of the networks would freeze, thereby cutting off flow into the larger channels downstream and arresting further development of the valleys. The presumed warmer conditions are believed to have occurred mainly very early in the planet's history. The valley networks are almost entirely restricted to the oldest terrains on Mars, in contrast to the outflow channels which cut into materials with a wide range of ages (Carr and Clow 1981; Baker and Partridge 1986). The simplest explanation of the almost complete restriction of valleys to the oldest terrains is that the valleys themselves are old, and that conditions required for valley formation were commonly met early in the planet's history but only rarely throughout the planet's subsequent history. These two inferences, that warm conditions are required for valley formation and that the valleys are mostly old, have led to a perception that early Mars was warm and wet, and that climatic conditions then changed such that conditions similar to those prevailing today were maintained for much of the later planetary history. We caution that these conclusions are by no means proven.

The climatic conditions required for valley formation are not clear. The assumption of warm wet conditions is based on the premise that the valleys formed by slow erosion of running water. The arguments for water as the erosive agent are based on analogy with terrestrial valleys and the abundant corroborative evidence, in addition to the valleys, for the presence of water at the Martian surface. Agents other than water or ice, such as wind, carbon-dioxide ice and lava are very unlikely for a number of reasons (for summaries, see Baker 1982; Carr 1981). However, as described above, many of the networks have a distinctly different appearance from terrestrial river valleys. Because of these differences and the difficulty of maintaining liquid water at the Martian surface, and, given the abundant evidence for water and ice, the possibility should be left open that many, perhaps most valleys formed not by Earth-like fluvial processes, but by some other mechanism, such as mass wasting, that involves water or ice.

If the valleys were formed by running water then the water could come from two possible sources, the ground or the atmosphere. Many valley networks have characteristics suggestive of ground-water sapping (Pieri 1980*a;* Baker 1982, Higgins 1982; Kochel et al. 1985). One requirement for ground-water sapping is that liquid water is stable sufficiently close to the surface that it can seep onto the surface. This implies that temperatures are above the freezing point at depths comparable to the scale of local topographic relief. This can be achieved in two ways: (1) the surface temperature may be above freezing; or (2) if surface temperatures are below freezing, the temperature gradient is large enough to allow liquid water close to the surface. Steep thermal gradients are likely on Mars. Heat of accretion and core formation would have resulted in heat flows more than a factor of 10 larger than at present during the first few hundred Myr of the planet's history (see chapter 5). This, combined with low conductivities expected of the brecciated megaregolith could have resulted in liquid water at shallow depths early in the planet's history irrespective of the surface temperatures.

If surface temperatures were below freezing and water reached the surface, then it may have been possible for streams to survive and cut the valley networks. Modeling of the freezing of small streams (Wallace and Sagan 1979; Carr 1983) suggests that if a stream 1-m deep or more can be started then the water could flow for a few 100 km, depending on slope and other factors, before it froze solid. However, such calculations are highly idealized and do not take into account the vagaries of natural systems. Streams from terrestrial springs under arctic conditions are typically arrested within a relatively short distance of the spring as a result of formation of icings (Carey 1973; Childers et al. 1977; Sloan et al. 1976). Thus, while theoretical calculations suggest streams could survive and cut channels under present conditions, analogy with terrestrial streams indicates the contrary. Moreover, if temperatures were permanently below freezing, then the water released onto the surface could not readily re-enter the ground-water system and recharge. Any recharging would have to be from juvenile sources.

If the valleys formed from precipitation, then surface temperatures most likely had to be above freezing. While it has been suggested that ice could be precipitated at low latitudes during periods of high obliquity (Jakosky and Carr 1985) and that the ice could melt and generate runoff despite low temperatures (Clow 1987), the amount of liquid water generated under these conditions is very small, and it is doubtful that it would be sufficient to cut the valleys.

Endogenetic Hydrologic Cycling

Local water budgets must be balanced to achieve the development of valley networks. For water to flow at the surface, it must be available at source areas. For spring sapping, a hydraulic gradient is ultimately achieved for terrestrial examples as solar energy generates an atmospheric hydrological cycle that transports evaporated or transpired water as vapor back to river headwaters, where it precipitates and infiltrates to recharge aquifer systems. However, it may also be possible to supply energy geothermally to recycle water in the subsurface.

The ancient valleys of the heavily cratered terrains are associated with very high cratering rates and with widespread volcanism. The former association led Brakenridge et al. (1985) to hypothesize that some of the valleys might have originated through the interaction of ground ice and hot springs located along the semi-permeable fringes of slowly cooling impact melts. They concluded that, with widespread operation of impact-related hydrothermal systems, the possible existence of which was proposed by Newsom (1980) and Schultz et al. (1982), it is not necessary to infer major atmospheric change to explain valley formation. Gulick (1992) estimated lifetimes of hydrothermal systems associated with impact crater and volcano formation. She concluded that hydrothermal systems could remain active for long enough periods to form the valley networks that are associated with these features.

Hydrothermal systems are an inevitable consequence of volcano or crater formation, tectonism, sill intrusion, or other plutonic intrusions into permeable, liquid water or ice-rich subsurface environments. Such geologic events generate local thermal anomalies which induce density perturbations in the ground water as heat is dissipated out into the surrounding country rock. Water near the anomaly migrates upward in response to buoyancy forces. This action results in the flow of ground water towards the thermal anomaly. Hydrothermal systems can focus, transport and recirculate large amounts of ground water to dynamic surface water environments for periods in excess of 10^5 yr (Gulick 1992). As pointed out by Gulick and Baker (1987*a*) water flow transported to the surface, via seeps and springs, may initiate valleys if there is a low permeability, erodible surface (e.g., ash) and sufficient runoff. However, if the surface is permeable (e.g., basalt flows), water will infiltrate and recharge near surface (perched) aquifers. These aquifers may eventually intersect with the surface farther downslope and initiate sapping at these locations. However, some of the upward migrating water may recharge high-level aquifers without ever flowing on the surface.

Squyres (1989*b*) proposes that the higher heat flow and steeper thermal gradient that existed early in Mars' history would have resulted in the melting of ground ice to a minimum depth of 350 m (see also chapter 32). However, in order to form sapping valleys, water would have to intersect with the surface environment and exploit a pre-existing (runoff) valley, joint, fault or fracture system as indicated by several terrestrial sapping process studies. The remaining problem, then, is in getting the subsurface water up into the surface environment, so that valleys can be initiated and eroded down to depths of the melted-ground ice reservoirs. The generation of hydrothermal systems, as discussed above, may provide a way of transporting and circulating this water through the surface environment.

The complex sequences of ancient cratered plateaus on Mars apparently include interstratified impact breccia, reworked aeolian sediments, lava flows or sills, and ice (Tanaka 1986; Wilhelms and Baldwin 1989*b*). The coincidence of very heavy cratering, extensive volcanism and regional valley formation suggests an association with endogenetic energy sources driving a dynamic hydrological system. As with the outflow channels, the subsurface character of this hypothesized system precludes its direct study.

VI. CONCLUSIONS

The spectacular relict channels and valley networks of Mars represent ancient hydrological conditions greatly different from those seen to be active on the planet today. The outflow channels are relatively young, late Hesperian or Amazonian in age (Tanaka 1986). They formed by immense outbursts of fluid from subsurface sources. Complexity in outflow-channel morphology was generated by varying amounts of sediment and ice in the aqueous-fluid flow systems. The overall cataclysmic-flood morphology thus may be locally transitional to morphologies generated by ice and debris flowage. Moreover, secondary processes, including wind, lava flows and mass movement, extensively modified some channel systems.

Although local areas of valley networks, such as on Alba Patera, formed coevally with outflow channel activity, regionally extensive networks dominate in the heavily cratered terrains. These networks are Noachian and early Hesperian in age. The morphology of many valleys suggests genesis by ground-water sapping; for some valleys, surface runoff may have been more important. The morphologic evidence is consistent with but does not require atmospheric or climatic change for its explanation. Endogenic cycling of water, as in volcanic or impact-related hydrothermal systems, provides an alternative explanation.

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