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WORKSHOP ON THE MARTIAN NORTHERN PLAINS: SEDIMENTOLOGICAL, PERIGLACIAL, AND PALEOCLIMATIC EVOLUTION

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WORKSHOP ON
THE MARTIAN NORTHERN PLAINS:
SEDIMENTOLOGICAL, PERIGLACIAL, AND
PALEOCLIMATIC EVOLUTION

Edited by

J. S. Kargel, J. Moore, and T. Parker

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Fairbanks, Alaska

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Preface

This volume contains papers that have been accepted for presentation at the Workshop on the Martian Northern Plains: Sedimentological, Periglacial, and Paleoclimatic Evolution, August 12–14, 1993, in Fairbanks, Alaska. The Program Committee consisted of Jeffrey S. Kargel (*USGS, Flagstaff*), Jeffrey Moore (*NASA Ames Research Center*), and Timothy Parker (*Jet Propulsion Laboratory*).

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Contents

Abstracts

HYDROLOGICAL CONSEQUENCES OF PONDED WATER ON MARS. V. R. Baker, Department of Geosciences, University of Arizona, Tucson AZ 84721, USA.

Although geomorphological evidence for ancient highly active water cycling has long been obvious for Mars [1–4], geochemical analyses have only recently been found to be consistent with this fact [e.g., 5]. One approach to understanding the geomorphological evidence has been to estimate the total volume of the martian hydrosphere [e.g., 2,6,7]. Here I will discuss an approach focused on the history of changing water processes on the planet, and on the understanding of those processes in a global sense.

Although the arguments for the various full-scale Mars ocean hypotheses [e.g., 8–11] are still much debated [12], there is abundant evidence for smaller-scale lacustrine basins, both early [13,14] and late [15–17] in martian history. Like the so-called "pluvial" lakes of Earth's arid regions, these martian water bodies had a transient existence, during which a sufficient water volume V persisted in a basin through time t according to the hydrological balance

$$\frac{dV}{dt} = \frac{d(P+R+I)}{dt} - \frac{d(E+O)}{dt}$$

where P is precipitation input to the water body, R is runoff from tributaries feeding the water body, I is groundwater inflow to the basin, E is evaporation loss, and O is groundwater outflow from the basin. For Earth studies there is considerable experience with estimating various parameters in the balance, simplifying for equilibrium

$$\frac{dV}{dt} = 0$$

and evaluating runoff in terms of coefficient K for transforming precipitation over tributary catchments P_T

$$R = K\,P_T$$

Mifflin and Wheat [18], Street-Perrot and Harrison [19], and Benson and Paillet [20] illustrate how paleolake levels can be used to indicate paleoclimatic parameters prevailing during periods that allowed lake persistence.

Baker et al. [11] presented a very rough and preliminary calculation demonstrating the paleohydrological possibility of such a balance, in which inflow was given by cataclysmic channel flood discharges. However, long persistence of the water bodies is indicated by glacial landforms [21]. There is an obvious need to evaluate long-term balances.

There are many issues raised by these long-term balances. For example, the inflow discharges have for the most part been only crudely approximated. Ideally, it would be best for the outflow channels to calculate model discharges as done for terrestrial flood paleohydrology [22]. These procedures, which have documented the largest known flood discharges on Earth [23,24], will best be applied to high-resolution Mars Observer topographic profiles. Intrabasin water transfers, as evidenced by lake spillways, are important checks on the overall balance. Scott and Chapman [15] recognize at least one such spillway leading from the Elysium Basin.

It is likely that martian flood channels may have developed a regime relationship to the scale of their water reservoirs. This behavior has been documented for catastrophic floods on Earth (Fig. 1), and it seems probable for Mars.

The association of the northern martian plains of ancient lakes or seas, glaciers, permafrost, and related phenomena is reminiscent of certain full-glacial Quaternary phenomena on Earth. Marginal lakes and spillways of the Laurentide ice sheet [25] and the newly discovered catastrophic paleofloods of central Asia [24] provide a rich new treasure for geological reasoning about the responsible processes. Regional relationships for the Asian lakes are shown in Fig. 2.

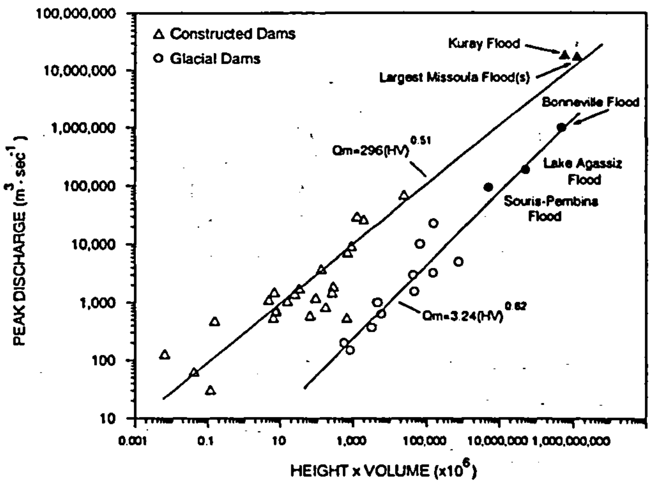

Fig. 1. Relationship of peak discharge to the product of dam height × lake volume for terrestrial cataclysmic floods (see [24]).

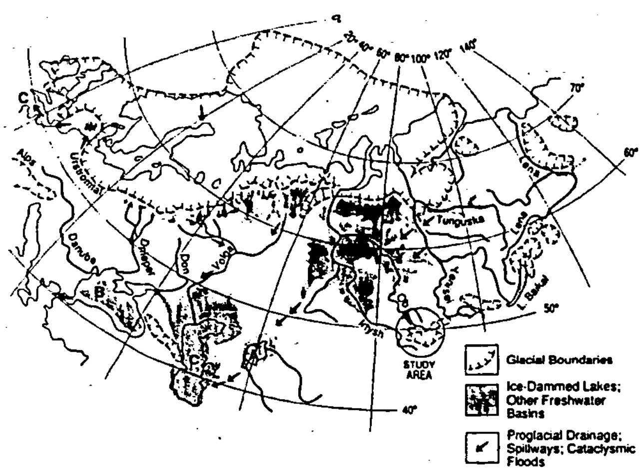

Fig. 2. Late Pleistocene Eurasian Ice Sheet and related drainage features showing the location of the Altay Mountain flood area described by Baker et al. [24]. Glacial spillways carried outflow floods to the Pleistocene predecessors of the Aral Sea (A), Caspian Sea (C), and Black Sea (B). These are potential analogs to the martian northern plains paleolakes (seas, oceans) and associated floods and glaciation. From Rudoy and Baker [26].

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MORPHOLOGIC AND MORPHOMETRIC STUDIES OF IMPACT CRATERS IN THE NORTHERN PLAINS OF MARS.

N. G. Barlow, Lunar and Planetary Institute, 3600 Bay Area Boulevard, Houston TX 77058, USA.

Fresh impact craters in the northern plains of Mars display a variety of morphologic and morphometric properties. Ejecta morphologies range from radial to fluidized, interior features include central peaks and central pits, fluidized morphologies display a range of sinuosities, and depth-diameter ratios are being measured to determine regional variations. Studies of the martian northern plains over the past five years have concentrated in three areas: (1) determining correlations of ejecta morphologies with crater diameter, latitude, and underlying terrain; (2) determining variations in fluidized ejecta blanket sinuosity across the planet; and (3) measurement of depth-diameter ratios and determination of regional variations in this ratio.

Ejecta morphologies are subdivided into six major categories: single lobe (SL), double lobe (DL), multiple lobe (ML), pancake (Pn), radial (Rd), and diverse (Di). Single-lobe, DL, and ML morphologies are lobate structures with a distal ridge marking the edge of the outermost lobe. Pancake morphologies are found primarily around craters <5 km at high latitudes; this morphology lies outside the diameter range (≥8 km diameter) used in this study and is not discussed further. Radial morphologies are represented by linear ejecta structures radiating outward from the crater; although slightly different in appearance, these morphologies approximate the radial ejecta patterns surrounding fresh impact craters on the Moon and Mercury. Diverse morphologies represent patterns where elements of both radial and fluidized morphologies are present.

Two major theories have been advanced to explain the different ejecta morphologies surrounding martian impact craters. The first model proposes that volatiles trapped in the near-surface region are vaporized during impact and ejecta entrained in this vaporized material is emplaced in a flow pattern [1]. The second model proposes that the martian atmosphere is thick enough to provide the fluidizing medium in which the ejected material is entrained [2,3]. The observed latitude dependence of several ejecta morphologies in this analysis strongly supports the buried volatile model over the atmospheric entrainment model [4].

The ejecta morphology analysis utilized 3819 craters ≥8 km in diameter distributed across the entire martian surface [4]; 1150 of these craters were located within terrains encompassing the northern hemisphere plains. Single-lobe craters dominate all other ejecta morphologies and are found to be correlated with diameter: SL craters occur in the 8–66-km-diameter range, but dominate between 8 and 25 km diameter. However, the SL diameter range varies with latitude: Near equatorial regions, the SL diameter range is smaller (8–20 km), but at higher latitudes it covers a greater extent (8–40 km). The opposite trend is observed for ML craters: These craters are primarily found in the 20–45-km-diameter range near the equatorial regions, but are associated with only larger crater diameters (40–60 km) at high latitudes. Computation of crater depths from depth-diameter relationships [5] and comparison of these results with the proposed distribution of subsurface volatiles on Mars [6] suggest that this observed distribution of ML and SL craters can be explained by ML craters forming from excavation into liquid reservoirs and SL craters forming by impact into ice [4].

The majority of DL morphologies are <50 km in diameter and are found within the 40°–65°N latitude range. Previous studies have proposed that the DL morphology results from impact into a volatile-stratified surface [7]. A current study of DL craters within the northern plains will determine if a relationship exits between the distribution of DL morphologies and ancient impact basins.

Little correlation between Rd or Di morphologies with latitude were noted in the ejecta morphology study. Radial morphologies dominate the ejecta structure for large craters (generally >50 km) at all locations across Mars and have been suggested to form from impact into volatile-poor materials. Diverse morphologies are more regionally confined, occurring most often along the highlands-plains dichotomy boundary. They commonly occur in the30–55-km-diameter range and may represent a transition morphology between the smaller lobate crater patterns and the larger radial morphology.

A second study has quantified the sinuosities of the SL, DL, and ML morphologies [8]. This study was inspired by the results of a more regionally restricted study of martian lobate crater sinuosity [9] that reported strong correlations between sinuosity and crater diameter, latitude, and terrain. Our study distinguished between different lobate ejecta morphologies and, for DL and ML craters, measured the sinuosity of all complete ejecta lobes. Results of our study indicate that SL and DL craters have similar low sinuosities, but the larger ML craters display much more sinuous ejecta patterns. In addition, sinuosity increases from inner lobe to outer lobe for the DL and ML craters. No correlation between sinuosity and diameter, latitude, or terrain was detected within the morphology classes; we believe the previously reported correlations resulted from not distinguishing between the ejecta morphology classes.

A third, and currently continuing, study of impact craters in the northern plains of Mars is using photoclinometry to determine depth-diameter ratios of fresh impact craters. Studies of Maja Valles and the highland region of Arabia indicated that crater depth-diameter ratios were approximately constant between these areas, and therefore measurement of these ratios for all craters could provide constraints on the amount of degradation experienced by the region [10,11]. However, studies of craters within the Medusae Fossae Formation deposits in the Memnonia-Amazonis region of western Mars indicate that fresh craters are deeper than predicted based on comparisons with Arabia, Maja Valles, and the heavily cratered region adjacent to the Medusae Fossae Formation [12,13]. This increase in crater depth is attributed to properties of the target material, which consists of thick, fine-grained, easily erodible deposits. The current study is therefore expanding to other regions of the martian northern plains to determine if the Medusae Fossae Formation is anomalous or representative of regional variations in target properties across the plains units.

Impact craters within the northern plains regions of Mars display a wide variety of morphologic and morphometric characteristics. Continuing studies of these features are helping to better constrain the properties of the surface and near-surface regions of the northern hemisphere. The results of such studies, combined with new information from both orbital and surface spacecraft missions to Mars, will dramatically increase our understanding of the current state of the northern plains and lead to enhanced understanding of the geologic history of this region.

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CALDERAS PRODUCED BY HYDROMAGMATIC ERUPTIONS THROUGH PERMAFROST IN NORTHWEST ALASKA. J. E. Begét, Department of Geology and Geophysics, University of Alaska, Fairbanks AK 99775-0760, USA.

Most hydromagmatic eruptions on Earth are generated by interactions of lava and ground or surface water. This eruptive process typically produces craters 0.1–1 km in diameter, although a few as large as 1–2 km have been described. In contrast, a series of Pleistocene hydromagmatic eruptions through 80–100-m-thick permafrost on the Seward Peninsula of Alaska produced four craters 3–8 km in diameter. These craters, called the Espenberg maars, are the four largest maars known on Earth.

The thermodynamic properties of ground ice influence the rate and amount of water melted during the course of the eruption. Large quantities of water are present, but only small amounts can be melted at any time to interact with magma. This would tend to produce sustained and highly explosive low water/magma (fuel-coolant) ratios during the eruptions. An area of 400 km^2 around the Alaskan maars shows strong reductions in the density of thaw lakes, ground ice, and other surface manifestations of permafrost because of deep burial by coeval tephra falls.

The unusually large Espenberg maars are the first examples of calderas produced by hydromagmatic eruptions. These distinctive landforms can apparently be used as an indicator of the presence of permafrost at the time of eruption.

MILANKOVITCH INSOLATION FORCING AND CYCLIC FORMATION OF LARGE-SCALE GLACIAL, FLUVIAL, AND EOLIAN LANDFORMS IN CENTRAL ALASKA. J. E. Begét, Department of Geology and Geophysics, University of Alaska, Fairbanks AK 99775-0760, USA.

Continuous marine and ice-core proxy climate records indicate that the Earth's orbital geometry modulates long-term changes. Until recently, little direct evidence has been available to demonstrate correlations between Milankovitch cycles and large-scale terrestrial landforms produced during worldwide glaciations. In central Alaska large areas of loess and sand fill valleys and basins near major outwash streams. The streams themselves are bordered by sets of outwash terraces, and the terraces grade upvalley into sets of moraines. The discovery of the Stampede tephra (~175,000 yr ago) reworked within push moraines of the Lignite Creek glaciation suggests that this event correlates with the glaciation of marine isotope stage 6. A new occurrence of the Old Crow tephra (~140,000 yr ago) on the surface of the oldest outwash terrace of the Tanana River, correlated with Delta glaciation, suggests this event also occurred at this time. The penultimate Healy glaciation apparently correlates with marine isotope stage 4, while radiocarbon dates indicate the latest Pleistocene moraines correlate with marine isotope stage 2. Recognition of the importance of orbital forcing to the cyclical formation of glacial landforms and landscapes can help in interpretations of remotely sensed glacial and proglacial landforms.

THE FATE OF WATER DEPOSITED IN THE LOW-LYING NORTHERN PLAINS. M. H. Carr, U.S. Geological Survey, Menlo Park CA 94025, USA.

Many large outflow channels terminate in the low-lying northern plains. If the outflow channels formed by running water, as appears likely, then standing bodies of water must have accumulated at the ends of the channels. Most of the observed channels, and hence the bodies of water, are post-Noachian. They formed after the period for which we have the most abundant evidence of climate change. While it has been speculated that the post-Noachian period has experienced large, episodic, climatic excursions [1], this paper takes the more conservative view that the climatic conditions on Mars, at least from mid-Hesperian onward, have been mostly similar to the climatic conditions that prevail in the present epoch. Thus obliquity variations are taken into account, but massive climate changes induced by the floods are considered so improbable that they are ignored.

Under present climatic conditions a lake on Mars will immediately freeze over. The rate at which the lake freezes will depend on the rate of loss of heat from the surface of the ice and the rate of loss of heat by conduction into the lake floor. Both loss rates decline rapidly with time. The sublimation declines because the vapor pressure of water varies logarithmically with temperature, and the temperature on the ice surface falls as the ice thickens. Heat loss through conduction declines as the temperature gradient from the lake bed into the ground becomes less steep. The fate of the water depends largely on the sublimation at the surface. Conduction into the surrounding rocks will not be discussed further, although it was taken into account in the calculations.

Sublimation rates were calculated from the heat balance at the surface of the ice [2]. The important terms are heat radiated from the surface, adsorbed solar radiation, adsorbed long-wave radiation from the atmosphere, heat conducted up through the ice, and heat lost by sublimation. These terms vary both systematically with time and periodically with season, and depend on a variety of factors such as albedo, winds, obliquity, and optical depth of the atmosphere. In a typical result, a 10-m-deep lake takes 1 yr to freeze, by which time the sublimation rate has fallen from an initial 10^{-5} to close to 10^{-8} g cm^{-2} s^{-1}. After 6 yr 2–3 cm of ice would have sublimed from the surface. Assuming a lake with an area of 10^6 km^2, and that all the sublimed water froze out at the poles, then the polar layered terrain would have accumulated a layer of ice roughly 10 cm thick in the 6 yr. If the ice remained bare, it would continue to slowly sublime at all latitudes lower than about 80°. Although the sublimation rates would be low, typically in the range of 0.01–0.1 cm yr^{-1} by the time the thermal anomaly represented by the ice deposit had largely dissipated, the rates are sufficient to eliminate the deposit in a geologically short period of time (10^4–10^5 yr). However, Mars is a dusty planet, and the ice deposit is likely to accumulate a dust cover that would inhibit sublimation.

The loss of water from the ice through the overlying dust cover was determined by first estimating the temperature, and hence the vapor pressure, of water at the soil/ice interface, and then applying the techniques of Clifford and Hillel [3] to determine the rate of diffusion of water upward through the soil. The rate of water loss depends mainly on the thickness of the soil, the soil properties, the opacity of the atmosphere, and the obliquity. If the soil cover is thinner than the depth of penetration of the diurnal wave (a few centimeters), then the sublimation rates are similar to those of bare ice, and the deposit will slowly sublime. However, with a thicker dust cover sublimation loss is severly impeded. For example, at 50° latitude with a 1-m dust cover, sublimation will occur only at very high obliquities and then only at very low rates. Ice deposits at latitudes greater than about 40° buried under 1 m or more of dust are essentially permanent under the present climatic conditions. However, lower-latitude deposits would be only temporary. Irrespective of the depth of their cover, they would continue to sublime into the atmosphere until they had completely dissappeared.

Most of the large outflow channels around the Chryse Basin, those that originate in Elysium, and those that extend into Hellas almost all end at latitudes higher than 40°. In addition, thick deposits of dust are suspected at these high latitudes, particularly the high northern latitudes [4]. We should therefore expect to find permanent ice deposits in the lowest-lying areas. Several characteristics of these areas suggest the ice deposits are indeed present.

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EVIDENCE FOR AN ICE SHEET/FROZEN LAKE IN UTOPIA PLANITIA, MARS. M. G. Chapman, U.S. Geological Survey, Flagstaff AZ 86001, USA.

Previous workers have noted evidence for a lacustrine basin in Utopia Planitia, Mars [1]. Geomorphic features within the basin that collectively suggest that water or ice may once have been present include (1) channels within the basin, (2) channels peripheral to the basin, (3) etched basin floor, (4) "thumbprint" terrain (whorled patterns), (5) polygonal outlines, (6) smooth floors (infilled), (7) shoreline indicators (terraces, platforms, lineaments), and (8) small cratered cones (pseudocraters or pingos) [1]. The authors interpret these data to suggest that the basin may have been the locus of a large paleolake in the northern lowlands of Mars. Alternatively, the area has been proposed to be part of an ancient circumpolar ocean [2].

The hypothetical paleolake was probably frozen to some depth. In fact, features now present at its boundaries suggest that the edges may have been frozen solid. Plains units at the southwestern boundary of Utopia Planitia show ridges that have been compared with those that form at the mouths of Antarctic ice streams [3], thumbprint terrain, and young, high deposits having lobate margins that suggest mudflows [4]. Thumbprint terrains have been interpreted as recessional moraines [5], ice-pushed ridges [6], or subglacially eroded tunnel valleys with eroded eskers [7]. At the east boundary of Utopia Planitia, geologic mapping at 1:500,000 scale of the Granicus Valles area (MTM quadrangles 30227, 30222, and 25227) indicates the presence of a basal scarp around the northwest flank of Elysium that formed the east boundary of an ancient ice sheet [8].

The northeast-trending scarp, formed by Early Amazonian faulting, separates the shield of Elysium Mons and the varied material that fills lower-lying Utopia Planitia [9]. Upper Hesperian lavas of the shield [10] are cut by the scarp (between latitude 28.1°N, longitude 223.1° and latitude 28.8°N, longitude 222.4°), but younger shield lavas [10] partly bury it. The scarp is interpreted to be the expression of a hinge fault (west side down) that may have been associated with extensional regional stress along a preexisting zone of weakness [9], because the scarp lies along the trend of a major wrinkle ridge in Lower Hesperian ridged plains at about latitude 16°–19° N, longitude 230°, about 600 km southwest of the Granicus Valles area [11]. The scarp was instrumental in the growth of Elysium Fossae northwest of the volcano [9]. Six major theater-headed fossae, including Granicus Valles, lie along the scarp.

The basal scarp is interpreted to have been the east boundary of an ice sheet that covered the area during a prolonged period in the Early Amazonian [8]. Although some workers have interpreted all rough-textured deposits northwest of Elysium as lahars [12] or erosional plains [13], many high-standing, rough, and flat-topped mounds trend along fissures of the Elysium Fossae west and north of the scarp; the mounds resemble Icelandic table mountains and moberg ridges. Previous workers have suggested that these types of subice volcanos occur in other areas of Elysium [14] and Utopia [15,16] farther north beyond the scarp and have found evidence of volcano-ground ice interactions in the general area [13,17]. The mounds are associated with low-standing, much smoother appearing deposits containing nested concentric (collapse) pits [17,18] and irregular channels. These deposits are locally bounded by lobate scarps and have been termed lahars [12], but they are more likely jökulhlaups formed by eruption of hot lava beneath the ice sheet [8]. The Granicus Valles are suggested to have been initiated by subglacial fluvial activity [14]; this interpretation is supported by the occurrence of an inner channel ridge [8] that bears a striking resemblance to some channel eskers noted elsewhere on Mars and Earth [7].

The ice sheet west of Elysium may have been a frozen remnant of the east edge of the paleolake in Utopia Planitia [1] or an ancient ocean [2] that lapped against the basal scarp of Elysium Mons. The northern part of the mons, north of the scarp, may have been covered by a glacier that fed into the paleolake, because there the ice-related features [13,17], table mountains, and moberg ridges lie on a steeper slope than those west of the scarp.

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A WET-GEOLOGY AND COLD-CLIMATE MARS MODEL: PUNCTUATION OF A SLOW DYNAMIC APPROACH TO EQUILIBRIUM.

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Introduction: It has been suggested that Mars may have possessed a relatively warm, humid climate and a vigorous hydrological cycle involving meteoric precipitation, oceans, and continental ice sheets [1,2]. Baker et al. [1] hypothesized that these geologically active conditions may have been repeated several times; each of these dynamic epochs was followed by a collapse of the climate and hydrologic cycle of Mars into essentially current conditions, completing what is here termed a "Baker cycle." The purpose of this brief report is to present an endmember possibility that martian glacial landscapes, including some that were previously considered to have formed under warm climatic conditions [1–5], might be explained by processes compatible with an extremely cold surface.

Two aspects of hypothesized martian glacial terrains have been cited as favoring a warm climate during Baker cycles: (1) The formation of some landscapes, including possible eskers, tunnel channels, drumlins, and outwash plains, appears to have required liquid water and (2) a liquid-surfaced ocean was probably necessary to feed the glaciers. The requirement for liquid water, if these features have been correctly interpreted, is difficult to avoid; it is entirely possible that a comparatively warm climate was involved, but it is not clear that formation of landforms by wet-based glaciers actually requires a warm climate. Even less certain is the supposed requirement for liquid oceans. Formation of glaciers only requires a source of water or ice to supply an amount of precipitation that exceeds losses due to melting and sublimation. At martian temperatures precipitation is very low, but so are melting and sublimation, so a large body of ice that is unstable with respect to sublimation may take the role of Earth's oceans in feeding the glaciers. Recent models suggest that even current martian polar caps, long thought to be static bodies of ice and dust, might actually be slow-moving, cryogenic continental glaciers [6; D. A. Fisher, unpublished data]. Is it possible that subglacial processes beneath cryogenic (but wet-based) ice sheets formed the hypothesized martian glacial landscapes?

The Problem of Feature Freshness: Kargel and Strom [2] noted an important aspect of martian glacial features: They commonly are almost pristine, and they have not been worn down by postglacial fluvial processes associated with the warm, humid, postglacial terrestrial climate. Many Pleistocene glacial features on Earth also appear pristine at resolutions comparable to those of Viking images, but this is because only 10,000 yr of warm-climate degradation have elapsed since glaciation. The freshness of martian glacial features might be explained by freeze-drying caused by a rapid transition from warm to cold conditions. Alternatively, perhaps Mars has always possessed the cold, dry conditions conducive to freeze-dried preservation of landforms.

PSDAES: The possibility that the martian climate might have always resembled that of today, with small fluctuations, is an old and well-regarded possibility. PSDAES (Punctuation of a Slow Dynamic Approach to an Equilibrium System) is a glacier-oriented conceptual merger of Clifford's proposed model [7; S. M. Clifford, unpublished data] of martian hydrology with recent interpretations of present martian ice caps [6; D. A. Fisher, unpublished data] and recent observations, interpretations, and theoretical advances in terrestrial glacial geology [8–10].

PSDAES starts with a Mars similar to that envisioned by Clifford [7; S. M. Clifford, unpublished data], one that resembles the current conditions below and on the surface. Ice-rich permafrost exists almost globally, although where ice occurs at latitudes less than about 30°–45° it is unstable over geological time due to sublimation (but it may persist metastably) [11]. The melting geotherm exists at depths of several kilometers, and groundwater flows beneath it in a global aquifer, allowing desiccated permafrost to be replenished with water derived from basal melting of polar ice caps and ice-rich permafrost at higher latitudes.

In the PSDAES concept, the present martian polar caps (composed of polar layered deposits and the residual "white" ice caps) constitute genuine, albeit cryogenic, continental glacial ice sheets [6; D. A. Fisher, unpublished data]. The lateral glacial transport of ice away from cold accumulation zones to warmer locales and the consequent ablation try to balance precipitation. Precipitation rates in the cryogenic martian accumulation zones are very low (hundreds of micrometers per year, perhaps), so the ice caps do not need to flow very quickly to balance accumulation. The chief source of ice for the glaciers is sublimation of low-latitude periglacial ice (some of which may have been deposited by glacial outbursts, as described below) and vapor emitted at the sublimating fronts of polar layered deposits. The water exchange between the permafrost and atmosphere and the hydraulic flow through a global subpermafrost aquifer [7; S. M. Clifford, unpublished data] result in a slow approach toward an equilibrium distribution of ice on the surface and in the subsurface of Mars [11].

Because of dynamical effects on glacier movement caused by basal water [12], basal melting often occurs only in relatively small regions, and even then only a few millimeters of ice melts per year at most. This water is normally accommodated by hydraulic flow away from the glacier through the substrate [7; S. M. Clifford, unpublished data] and by a forward transport of slurry at the glacier sole [13], where it then refreezes near the cold, thinned terminus of the glacier. In general, a state of dynamic near-equilibrium prevails. So long as a glacier remains frozen over most of its base, the glacier has little ability to modify its bed; a largely wet-based glacier interacts strongly with its bed and leaves abundant geomorphic evidence of its activity.

If subglacial hydraulic flow and expulsion of basal slurry and meltwater do not balance water production, then deep subglacial lakes or thick accumulations of water-saturated basal till may form; in either case, glacier behavior is fundamentally altered by high hydraulic pressures at the base of the glacier. A range of non-steady-state behavior ensues, including glacier surging (ice streaming and lobe formation) and spasmodic thickening and thinning in response to surging; at the most catastrophic scales, large water outbursts sometimes form extensive proglacial lakes. When deep proglacial lakes form, or when a glacier meets the sea, the water lifts the leading edge of the glacier, and ice shelf phenomena also become active, causing an acceleration of ice streaming and sometimes prompting disintegration of the ice sheet.

The West Antarctic ice sheet is presently in this non-steady-state regime [14], as was the Laurentide ice sheet for several thousand years during its demise. It is thought that the isostatically depressed Hudson Bay may have filled with subglacial water [10]. According to one range of interpretations, megajökulhlaups emitted from the Hudson Bay area and other subglacial lakes formed expansive glacial landscapes, including eskers, tunnel channels, moraines, bedrock and till flutings, and drumlins. Some subglacial glacio-lacustrine-fluvial landscapes are thought to have been as much as 1800 km long and 280 km wide [8,9]. One such outburst in Alberta may have released over 84,000 km^3 of water and eroded thousands of cubic kilometers of rock [9]. Under related interpretations, such landscapes may have formed by water-saturated till streams and surging ice instead of basal meltwater floods [15,16]. In any case, glacier dynamic behavior caused by large accumulations of basal meltwater was a chief agent in such events. The similarity of some martian glacial landscapes to terrestrial ones that were formed by ice surging, rapid basal till flow, and/or glacial outburst floods suggests that similar processes also occurred on Mars.

Megajökulhlaups: The "Punctuation" of PSDAES: Regardless of the mechanisms of late Pleistocene glacial surging and outburst flooding on Earth and for the rapid disintegration of the Laurentide ice sheet, a major factor certainly was a warming climate. It is uncertain whether, in general, a warming climate is a necessary factor, or whether warming was simply the particular aspect that triggered late Pleistocene glacial activity and, ultimately, deglaciation. At least until a theoretical analysis or future observations prove otherwise, it is reasonable to suppose that glacier surges, rapid flowage of subglacial till streams, and mega-outburst flooding events might have occurred even for cold-surfaced martian ice sheets.

Large basins on Mars form ideal traps for voluminous subglacial lakes. Rapid drainage of these lakes and accompanying glacial surges may drastically and fairly suddenly reduce the confining pressure of sedimentary and other rocks on the basin floor, causing slumping, fracturing, and general disintegration of those rocks. This may have occurred in the Valles Marineris area, producing the great outflow channels and chaotic terrains [17] and large surface pondings in the Chryse Basin [S. Rotto and K. L. Tanaka, unpublished data]. Rapid drainage of a voluminous subglacial lake in the South Polar Basin could have created the sinuous ridges (eskers) of Dorsa Argentea, the great channels that open into southern Argyre, and the sinuous ridges and other glacial topography of Argyre. Associated surged glaciers may have stagnated and formed the south polar thermokarstic etched terrain [20]; water and ice that debouched into Argyre would have frozen and stagnated, and eventually partially sublimated, possibly forming the great thermokarstic structures of Argyre. While the scale of such hypothesized events seems almost too incredible, the scale is no greater than similar events hypothesized to have occurred in Canada [8–10].

Closure of the PSDAES Cycle: When glacial surges and outburst floods punctuate the course of martian geologic events, water and ice move from areas of high ice stability to areas down the hydraulic gradient; sinks for water and ice are generally areas where ice has been unstable over geologic time (otherwise an ice sheet would have been there, and the water would have flowed somewhere else). Water released from these outflows rapidly freezes and then slowly sublimates; the vapor is deposited back near the source of the ice sheet, allowing the ice sheet to reform and perhaps to repeat the cycle.

Conclusions: Martian glacial terrains show good evidence of formation by glacial surging and water outbursts. Ice shelves were probably important in the formation of some terrains in the northern plains. There may be a natural tendency for continental glaciers (on Earth and Mars) to evolve and eventually self-destruct by subglacial melting, regardless of surface climatic conditions. Individual outbursts, especially the large ones, are probably triggered by discrete perturbations, such as climate changes, increased subglacial volcanic activity and increased geothermal gradients, and perhaps asteroid impacts.

Outburst and glacier surge behavior described here appear to have been important on Earth, under terrestrial glacial climatic conditions; it is assumed, but not certain, that the same mechanisms can work at the warm bases of ultra-cold-surfaced martian glaciers. If these processes cannot work on Mars under cold conditions, then one is driven to the concept of a warm ancient Mars, and must find a mechanism that would have allowed Mars to revert quickly to cold, dry conditions so that glacial landscapes were preserved.

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POSSIBLE OCCURRENCE AND ORIGIN OF MASSIVE ICE IN UTOPIA PLANITIA. J. S. Kargel[1] and F. M. Costard[2], [1]U.S. Geological Survey, Flagstaff AZ 86001, USA, [2]Centre National de la Recherche Scientifique, Centre de Geomorphologie, Caen, France.

Introduction: F. M. Costard recently discovered a large field of possible thermokarst depressions near latitude 45°N, longitude 270°, in western Utopia Planitia (Costard and Kargel, unpublished data). Oval to circular pits are typically 300–1000 m across and ~25 m deep; larger depressions, 3–5 km across, are compound and seem to have formed by coalescence of smaller pits (Fig. 1). Small domical hills occur on the floors of two pits. These depressions characteristically have steep, scalloped edges and one or more inner benches. Truncation relations of the benches suggest a discontinuous, lenslike stratification of the material in which the pits are developed.

Based on a close analogy in form and scale with coalesced thaw-lake basins (alases) on Earth [1–3], the martian pits may have formed by thermophysical interactions of pooled water with ice-rich permafrost. This interpretation is not unique; sublimation of ice-rich permafrost or possibly even eolian processes acting on ice-free material might have formed the pits. However, the regional setting contains many other indications of massive ice. Below, we examine some of these features.

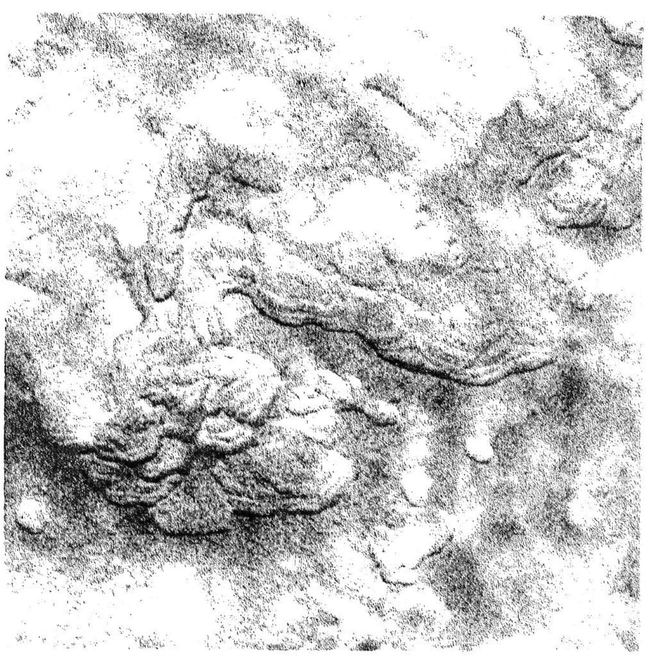

Fig. 1. Possible coalesced thaw-lake basins (alases), discovered by Costard, in Utopia Planitia. Solar illumination from below. Scene width 7.5 km. Viking Orbiter 2 frame 466B85.

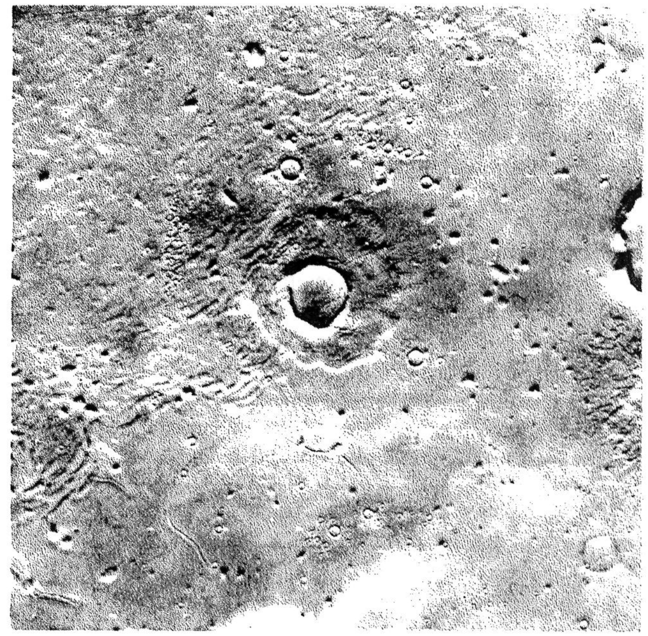

Fig. 2. Thumbprint terrain interpreted as possible moraines and eskers. Solar illumination from below. Scene width 150 km. Viking Orbiter frame 572A03.

Regional Landform Associations: Typical "terrain-softened" crater rims and "concentric crater fill," in the terminology of Squyres and Carr [4], occur in the pitted area. These features have been taken to indicate flow processes related to the presence of ground ice.

Rampart craters occur across almost the whole face of Mars, suggesting a global occurrence of ground ice or groundwater. Morphometry of rampart crater ejecta has indicated to some researchers that Utopia Planitia may be unusually ice rich [5].

"Thumbprint terrain" and sinuous ridges occur in a field that roughly parallels the highland/lowland transition a few hundred kilometers south-southwest of the depressions described above (Fig. 2). This terrain type has been interpreted as moraines [6–8] or as DeGeer (glaciolacustrine) moraines with associated eskers [9,10]; Lucchitta (personal communication) has suggested a similarity to ice shelf features. Nonglacial interpretations of thumbprint terrain include paleoshoreline terraces and near-shore spits or bars deposited in ancient seas or lakes [11,12], compressional features near the former shore of a massive mud ocean [13], and eolian deposits marking the former frontal positions of ancient sedimentary mantles [14].

Low domical mounds, many bearing summit depressions, ranging between 200 and 3000 m in diameter, occur widely across the northern plains, including parts of Utopia Planitia. Some pitted mounds are probably wind-eroded rampart craters, and others appear to be true volcanic cones. Some pitted domes occur in chains, arcs, or other types of clusters; some such clusters appear to be associated with basins and probably are pingos. At least two small pitted domes occur on the floors of alaslike pits in the area described above; this morphologic association is similar to the common occurrence of pingos on the floors of terrestrial alases [3].

Alaslike pits form an annular or arcuate trough at the bases of some large (3-km-diameter) pitted mounds. The flanks of the mounds taper toward adjoining surfaces, forming a volcanolike, inverted cone. These mounds might be volcanos or modified pedestal craters; alternatively, in keeping with the ice-rich sedimentary nature of the region, such pitted cones could be either pingos or mud volcanos; mud volcanos are important features of some terrestrial permafrost landscapes [15]. Alaslike moats can be explained by thermal interactions of water draining from mud volcanos; alternatively, moats might be explained by geothermal heating associated with mud or silicate volcanism, or, if the cratered mounds are modified impact craters, they can be explained by water drainage from ejecta.

Cryohydrovolcanic features, including possible table mountains and lahars, have been described from western Elysium and adjoining parts of eastern Utopia Planitia [16,17].

Polygonally cracked terrain is well developed several hundred kilometers to the east and southeast of the alaslike pits. Polygonal terrain has been interpeted as giant ice-wedge polygons [6] or as fractures induced by subsidence of wet, rapidly emplaced sediments [18].

Fretted terrain, believed to be ice-related landscapes modified by gelifluction or glacial flow of ice-rich debris [19], occurs along the highland transition near Utopia Planitia.

Alaslike pits are developed on locally smooth plains that were mapped by Greeley and Guest [20] as a part of the mottled member of the Vastitas Borealis Formation. The widespread occurrence of the mottled member across the northern plains, particularly in Utopia Planitia, suggests that whatever formed this terrain did so on a very large scale. Some of the mottled appearance of this geologic unit is due to impact crater ejecta, but high-resolution views show that considerable mottling consists of low rolling hills, the pitted domes described above, and arcuate to linear ridges probably representing poorly organized examples of thumbprint terrain. If these features are periglacial and glacial in origin, the wide extent of the mottled terrain suggests that much of the northern plains, particularly Utopia Planitia, may contain abundant ground ice, consistent with the conclusions of Chapman [17] and Lucchitta [19].

Origin of Massive Ice in Utopia Planitia: Pingos and ice wedges represent formation of massive segregated ice in permafrost, whereas thermokarst requires destabilization of massive ice. Ice interbeds, lenses, and wedges in permafrost do not form simply by diffusive infiltration and condensation of water vapor in pore spaces; rather, they require (1) intercalation with sediments of former snow or frost layers, stagnant glacier ice, or frozen lake water or (2) postdepositional surficial or interstratal melting, percolation, and segregation of water, then refreezing. The scale of purported thermokarst and pingos suggests thicknesses of massive ice on the order of tens of meters, similar to the thickness of massive ice in some permafrost areas of Arctic Alaska, Canada, and Siberia [21].

Many researchers have suggested that Utopia Planitia and adjoining areas were inundated by seas, large lakes, mud oceans, and/or ice sheets [7–13,17,18,22–25]. Brakenridge [25] recently interpreted features in nearby Elysium Planitia as sea ice, suggesting that ancient seas still exist, but in a frozen condition. Collectively, these studies suggest that Utopia Planitia could be very ice rich. Smooth plains and thermokarst in Utopia Planitia may be stagnant, debris-mantled glacier ice. Alternatively, formation of ground ice lenses in Utopia Planitia may have resulted from postdepositional periglacial segregation of ice.

Viking 2 Site Reinterpreted: Viking Lander 2 landed in eastern Utopia Planitia 1900 km east of the alaslike pits described above. The spacecraft revealed a poorly sorted, rock-strewn scene that has been likened to degraded, wind-eroded volcanic terrains in the Mojave Desert and to surfaces of debris flows [26–28]. An intersecting network of small troughs observed from Lander 2 has been interpreted as possible ice-wedge or sand-wedge polygons. Although there are wind-related features at the landing site, and the rocks may be volcanic, we suggest that the Lander 2 site might be a periglacially modified ground moraine or ice-cored moraine.

Conclusions: Utopia Planitia is very ice rich and seems to have been modified by glacial and periglacial processes. The northern plains may have formed under warm, wet conditions similar to those of the late Pleistocene in northern parts of Europe and North America. Alternatively, all the features described here may have been produced by processes active even in today's cryogenic climate. Future work should help develop explanations involving processes that might work under cryogenic conditions, and future observations should help to discriminate between cold- and warm-climate-based formational mechanisms.

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ICE IN THE NORTHERN PLAINS: RELIC OF A FROZEN OCEAN? B. K. Lucchitta, U.S. Geological Survey, 2255 North Gemini Drive, Flagstaff AZ 86001, USA.

Viking images revealed many features in the northern plains and along their boundary that early investigators believed to be formed by ice-related processes. The features are possible pingos [1], pseudocraters [2], table mountains and moberg ridges [3,4], thermokarst depressions [1,5,6], moraines [5], patterned ground [1,5], and lobate aprons that suggest viscous flow such as that of ice or rock glaciers [7–9]. More recently, many of these features have been reinterpreted as related to sedimentation in hypothetical former polar lakes, oceans, or alluvial plains [9–11] or as shoreline features of associated water bodies [12,13]. Here I review some evidence that points toward the existence of former bodies of standing water in the northern plains, but is also consistent with the idea that these bodies were ice covered or completely frozen.

Patterned ground formed of large polygons occurs mostly poleward of the mouth of large outflow channels in the low-lying northern plains [10]; it is thought to have formed in alluvial plains, lakes, or oceans by settling and contraction of frozen sediments [11,14]. Closely associated with this patterned ground are pedestal craters whose ejecta, rising above the general level of the ground, are surrounded by a scarp. If these ejecta had armor-plated underlying ice, they may have shielded it from sublimation that lowered the surrounding ground. Many small mounds topped by depressions occur in the same areas. They have variously been interpreted as cinder cones, pseudocraters [2], or pingos [1], but they could also be secondary craters standing in inverted relief, and thus would be smaller versions of the pedestal craters. Three of these interpretations suggest the former presence of ice. In addition, curvilinear ridges at the mouth of Ares Vallis resemble ridges in an ice shelf at the mouth of an Antarctic ice stream. The similarity suggests that Ares Vallis emptied into an ice-covered body of water [10]. The lack of deltas at the mouths of most martian outflow channels is also consistent with the former existence of ice shelves at these locations. Other recent work pertaining to features in the northern plains possibly caused by ice includes the recognition of possible tunnel channels, eskers and kames [15], lahars [16], ice sheets and jökulhaups [17], and sea ice [18].

Moreover, the northern plains locally show curvilinear ridges or hummocky to ridged and lobate patterns that resemble those of terrestrial moraines [5,19]. The orientation of these patterns suggests a local source or a source from the north, yet no local ice caps have been identified in these areas, nor do connections appear to exist between these features and the presently observed polar caps. Could these ridges and lobate patterns possibly reflect flow from thick bodies of ice left behind by polar lakes or oceans? The "moraines" occur in areas far enough north that ice may have persisted in the ground during certain past climatic cycles [20], yet they are far enough south that the relatively warm ice was able to flow [21,22].

The boundary between the southern highlands and the northern lowlands is extraordinarily complex in the region of the fretted terrain analyzed in detail by Parker et al. [12]. Their excellent and well-documented observations show that the boundary underwent repeated resurfacing, which they interpret to have been due to sediments settling out of a paleo-ocean. My alternative interpretation of their observations envisions the following sequence of events related to frozen lakes or oceans: A very early frozen body of water bordered the highlands in the region of the fretted terrain. The border region was flooded by plateau lavas, which did not melt the underlying ice [23] but left behind massive ground ice in the subsurface. The surface ice sublimed farther north, and a scarp developed near the present fretted terrain where the ice was protected by volcanic caprock. Eventually the ice disintegrated to form the fretted mesas, fretted channels, collapse features, and early versions of the glacierlike debris aprons whose last vestiges we see on mesa remnants today [21]. The possible thermokarst depressions on mesa tops elsewhere [1,5], and the moats surrounding some mesas in Kasei Valles reflecting now-vanished debris blankets [24], suggest that such massive ground ice may have also existed in other regions of the lowland-highland boundary. Thus, this frozen early paleo-ocean would have yielded the massive ground ice required by Sharp's [25] model for the origin of the fretted terrain. Later flooding in the northern plains inundated the already dissected scarp and formed the draping sediments of the older lowland unit of Parker et al. [12]. This unit is characterized by small-scale patterned ground similar to that of ice-wedge polygons [26]. The resemblance suggests that these sediments contained ice [12]. Even later, the flow lobes of Parker et al.'s younger lowland unit extended southward and penetrated the older fretted valleys. The lobes were thick enough during emplacement to overtop mesa and channel borders, but the material eventually collapsed so that now only a thin veneer is left. If these lobes contained ice that eventually sublimed, the disappearance of the material is explained. The ice bodies from which these lobes came may have been the remnants of the slowly subliming frozen polar ocean.

This hypothesis agrees with other recent ideas of standing bodies of water in the northern plains [12,13], but it differs in that the water is thought to have been partly or wholly frozen. Thus many of the features we see in the northern plains and along their border could be due to ice rather than water, and past climates need not have been very different from those of the present.

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OBSERVED CLIMATIC ACTIVITY PERTAINING TO THE EVOLUTION OF THE NORTHERN PLAINS. L. J. Martin[1] and P. B. James[2], [1]Lowell Observatory, Flagstaff AZ 86001, USA, [2]University of Toledo, Toledo OH 43606, USA.

Although most of the landforms observed by Viking were probably formed long before the advent of present climatic conditions, it is possible that some modifications have occurred more recently and may continue to the present. Also, present climate activity provides another dataset for hypothesizing earlier climatic conditions. This paper summarizes observations of climatic activity in the northern plains from 1905 to 1993. Data from Earthbound telescopes and from spacecraft, including the Hubble Space Telescope (HST), were used.

The best-documented seasonal events in the north are the recessions of the polar cap [1,2]. However, since martian northern springs occur during the less-favorable, aphelic apparitions (when Mars is more distant from the Earth), the north cap recessions have not been as well observed as those for the south cap. Even the Viking observations happened to be more favorable for the south cap recessions than they were for the north [3]. Averaged values for any point on the recession curve are accurate to within less than 2° of latitude, but the data for particular years is less dependable and thus the interannual variability of the recessions is not well known.

Throughout fall and winter, the northern plains are mostly cloud covered. These north polar hoods are the most dynamic phenomena in the martian atmosphere. During these seasons, this area is tilted away from Earth much of the time and at least partially hidden by the polar night. Few comprehensive studies of these hoods have been made, and their role in the forming of the seasonal surface cap is not understood. Daily maps of the positions of these clouds, compiled from 1969 International Planetary Patrol images, showed rapid, hourly motions that suggest diurnal influences [4]. Although the hoods are optically thin and transparent to red light, the seasonal polar cap is not visible either beneath them or during times when the hood clouds part to latitudes well north of cap-covered areas [5]. However, Viking data explicitly show that a CO_2 cap forms during the fall and winter seasons [6]. This CO_2 cap is evidently too patchy, thin, and/or dark to show on the red-filter images. The bright surface cap does not appear until the end of winter, when the hood clouds begin to disperse. This seasonal polar cap may have been brightened at that time by an H_2O component that has been cold-trapped by the subliming CO_2 [7].

Bright condensate clouds have also been observed around the periphery of the subliming spring cap. These clouds are most commonly seen near the terminator and/or bright limb of the planet, sometimes widely separated from the surface cap. When these clouds are close to or adjoining the edge of the cap, they may distort measurements of the cap's edge on blue-filter images [8].

Dust activity is less frequent in the northern plains than in other parts of Mars but is still an important component of the climate. The Tempe region (70°W) has been known as a location for suspected activity, although few of these events have been confirmed [9]. Condensate clouds, some of which probably have included dust, have repeatedly been observed over Alba Patera. Dust clouds have also been tentatively identified at the edge of the receding polar cap on HST images from May 1991. These were probably diurnal clouds since they were seen in similar relative positions near the bright limb on images of three different faces of Mars about 120° longitude apart. Two Viking sequences taken two days apart show a cloud front moving across an area that had been recently freed from surface ice by the spring retreat of the seasonal cap. These clouds have been tentatively identified as dust [10], based upon the characteristic billowing appearance of dust clouds that have been identified on earlier Viking observations [11]. Since these sequences were taken using a clear filter only, however, the composition of the clouds is not certain. Because the sequences of this area were not repeated during the same orbits, there is no record of possible diurnal activity. Another Viking sequence (from the day between the two above) shows additional cloud activity over the receding seasonal cap. Those clouds, however, may be condensates since they do not show the same visual characteristics established for dust. Color filters were not used for the more northerly sequence either.

Most, if not all, of the above activity plays a role in the continuing changes observed in the albedo features of the northern plains. While some of these changes in the surface albedo may be seasonal, others are not. Although these changes may be temporary and possibly superficial, they are evidence that the surface is not static. The data are insufficient to determine if topographic features might still be evolving; however, it is been suggested that the layered deposits may be actively forming [12]. It has also been pointed out that the possible presence of atmospheric hazes must be taken into account when evaluating the perceived amount of surface detail visible on some images [13].

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A MODEL FOR THE ORIGIN OF MARTIAN POLYGONAL TERRAIN. G. E. McGill, Department of Geology and Geography, University of Massachusetts, Amherst MA 01003, USA.

Extensive areas of the martian northern plains in Utopia and Acidalia Planitiae are characterized by "polygonal terrain." Polygonal terrain consists of material cut by complex troughs defining a pattern resembling mudcracks, columnar joints, or frost-wedge polygons on the Earth [1–4]. However, the martian polygons are orders of magnitude larger than these potential Earth analogs, leading to severe mechanical difficulties for genetic models based on simple analogy arguments [5,6]. Stratigraphic studies show that the polygonally fractured material in Utopia Planitia was deposited on a land surface with significant topography, including scattered knobs and mesas, fragments of ancient crater rims, and fresh younger craters [7–9]. Sediments or volcanics deposited over topographically irregular surfaces can experience differential compaction [10,11] producing drape folds. Bending stresses due to these drape folds would be superposed on the pervasive tensile stresses due to desiccation or cooling, such that the probability of fracturing is enhanced above buried topographic highs and suppressed above buried topographic lows. Thus it has been proposed that the scale of the martian polygons is controlled by the spacing of topographic highs on the buried surface rather than by the physics of the shrinkage process [7,9].

Polygonally fractured material provides few direct clues to its origin. Although both the sedimentary and volcanic models will work mechanically, the sedimentary alternative has been selected primarily because results of recent regional geological studies indicate that polygonally fractured materials are best interpreted as sediments [12,13]. Stresses developed during desiccation and compaction of materials with properties appropriate for water-deposited sediments on Mars are of the right order of magnitude to account for the fractures defining the giant martian polygons, but the analysis could, at least in principle, also be applied to a thick blanket of volcanics.

The geological and mechanical arguments for the origin of polygonal terrain by a combination of desiccation and bending have been published [7,9]. The calculations in [9] deal with the specific case of differential compaction over the rims of buried craters. Clearly, most of the fracture patterns present in areas of polygonal terrain cannot be explained as related to buried craters, but some can be, and it is relatively easy to develop a model based on the well-known geometry of craters. If the model works for crater rims, it should also work for other forms of positive topography that cannot easily be modeled quantitatively.

Despite the general success of the model, there are a number of problems and questions that remain, and this paper will focus on these. One of the more interesting corollaries of the model is that desiccation, compaction, and fracturing must occur very rapidly (on the order of thousands of years or less); otherwise, the material would simply flow without the development of fractures or faults. This is intuitively acceptable if the layer involved is thin. But the total thickness of polygonal terrain material is on the order of 600 m [9]. If this thickness were deposited by many small events, one would expect desiccation fractures of "normal" dimensions and spacing to form. We thus seem forced to infer that fewer events occurred, each depositing layers scores to hundreds of meters thick. Such "catastrophic" events are somewhat counter-intuitive. The martian outflow channels are large enough to account for rapid deposition of this much sediment [14], and these channels were active during the right time interval [15]. Even so, it is difficult to understand how such large volumes of material could desiccate so quickly. Obviously, we do not know enough about the physical properties of the polygonal terrain materials, but there is little chance that these uncertainties can be removed without much-higher-resolution orbital imagery or perhaps even returned samples. Furthermore, we have only the most rudimentary understanding of how the materials were transported and deposited; in particular, the number of depositional events is completely unknown. Our approach to these uncertainties has been to test a range of possibilities and make simplifying assumptions that are conservative.

Our hypothesis for the origin of large polygons also does not account for all the landforms and material units found within polygonal terrain. For example, if circular troughs result from drape folding over the rims of fresh craters, then why are circular troughs not more uniformly distributed within polygonal terrain? Careful mapping of these features in the Utopia Planitia area demonstrates that they occur only within the southern part of the polygonal terrain, and in immediately adjacent plains materials to the south and west that are not otherwise polygonally fractured. Some curved segments of polygon boundaries farther north might be related to segments of the rims of buried, degraded craters, but clear-cut, complete circles indicative of buried fresh, complete craters are absent. It is possible, of course, that the density of fresh craters superposed on the surface buried beneath polygonal terrain decreases northward, but this is an untestable hypothesis. Furthermore, many of the circular troughs are double, even though their diameters are far too small for the buried craters to be double-ringed impact structures. This suggests that the three-dimensional effects of the asymmetry of crater-rim profiles can create a more complex fold geometry than the single hinge line of maximum curvature that we assume.

A number of small-scale features within areas of polygonal terrain appear to be volcanic in origin. These include small cone-shaped landforms, areally restricted units that resemble lava flows, and narrow, aligned ridge-graben features that are probably related to shallow dike intrusion. All these features appear to be younger than the polygon troughs. How they bear on the origin of polygons, if they do at all, is not known at present. Broad, low-relief ridges are also common in Utopia Planitia, and these locally display a latticelike pattern with a horizontal scale comparable to the polygons. The age and structural significance of these are not known. I believe that these unresolved questions and unexplained observations are not fatal to our hypothesis; rather they represent opportunities for further study needed to fully understand the nature and origin of polygonal terrain.

The proposed combination of desiccation and differential compaction indicates that stresses caused by drape folds are of sufficient magnitude to control where failure will occur near the surface. Most of the extension expressed as grabens bounding polygons must be due to desiccation shrinkage. However, the loci of fracture initiation and fracture suppression are determined primarily by buried topography rather than by random distribution of strength inhomogeneities in the fracturing layer, as would be the case for pure desiccation fracturing. This effectively removes the scale barrier against desiccation or cooling cracking erected by Pechmann [6].

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THE DISTRIBUTION OF GROUND ICE ON MARS. M. T. Mellon[1] and B. M. Jakosky[2], [1]Department of Astrophysical, Planetary, and Atmospheric Sciences and Laboratory for Atmospheric and Space Physics, University of Colorado, Boulder CO 80309-0392, USA, [2]Department of Geological Sciences and Laboratory for Atmospheric and Space Physics, University of Colorado, Boulder CO 80309-0392, USA.

A wealth of geologic evidence indicates that subsurface water ice has played an important role in the evolution of martian landforms. Theoretical models of the stability of ground ice show that in the near-surface regolith ice is currently stable at latitudes poleward of about ± 40° and below a depth of a few centimeters to a meter, with some variations with longitude. If ice is not previously present at a particular location where it is stable, atmospheric water will diffuse into the regolith and condense as ice, driven by the annual subsurface thermal oscillations. The lower boundary of this ice deposit is found to occur at a depth (typically a few meters) where the annual thermal oscillations give way to the geothermal gradient. In the equatorial regions near-surface ice is currently not stable, resulting in the sublimation of any existing ice and subsequent loss to the atmosphere. However, subliming ice might be maintained at a steady-state depth, where diffusion and loss to the atmosphere are balanced by resupply from a possible deeper source of water (either deeper, not yet depleted, ice deposits or groundwater). This depth is typically a few tens to hundreds of meters and depends primarily on the surface temperature and the nature of the geothermal gradient, being deeper for a higher surface temperature and a lower geothermal gradient. Such an equatorial deposit is characterized by the regolith ice content being low nearer the surface and increasing with depth in the deposit.

Oscillations in the orbit will affect this picture of ground ice in two ways: (1) by causing periodic changes in the pattern of near-surface stability and (2) by producing subsurface thermal waves that may be capable of driving water ice deeper into the regolith.

Periodic changes in the pattern of near-surface ice stability result from oscillations in surface and near-surface temperatures due to orbitally induced changes in insolation and from oscillations in the atmospheric water content due to similar changes in polar insolation affecting polar cap summertime sublimation. The effects range from near-surface ice being stable at all latitudes during periods of high obliquity, when the equatorial regolith is colder and the atmosphere is wetter, to near-surface ice being restricted to the polar regions during periods of low obliquity, when the equatorial regolith is warmer and the atmosphere is dryer. These periodic oscillations in the obliquity may cause a cyclic saturation and desiccation of the near-surface regolith with ground ice over orbital timescales (with characteristic periods of 10^4–10^6 yr). However, estimates of the timescales for the diffusive exchange of water vapor between the atmosphere and the regolith are similar to the timescales for orbital oscillations, thus the near-surface ice configuration will depend on the previous orbital history, rather than the stability pattern only at that instant.

Orbitally induced subsurface thermal oscillations may be capable of driving ice deeper into the regolith by the molecular diffusion of vapor. Water vapor will diffuse from icy regions of higher temperature to icy regions of lower temperature following a gradient in the saturation vapor pressure. The geothermal heat flow alone produces temperatures higher at depth and lower at the surface, which in turn will drive water toward the surface. Orbital oscillations will create climate oscillations, specifically in the surface temperature, which will in turn produce a subsurface thermal wave penetrating kilometers in depth. This climatological effect can in some instances overwhelm the geothermal gradient, reversing the direction of thermally driven vapor diffusion over short depth intervals. These regions might trap small pockets of ice and drive them deeper into the regolith as the thermal wave propagates downward. Locations where this process might be active, the amount of ice transported, and how deeply the ice might be carried will depend on the magnitude of these thermal oscillations in relation to the magnitude of the geothermal gradient of Mars and the rate at which water vapor diffuses through the regolith.

These patterns of ice transport and stability may have effects on the geomorphic processes active on Mars on similar length scales. Near-surface ice, having depth scales on the order of meters to tens of meters, might affect the surface on a relatively small areal extent, producing features such as meter-scale patterned ground, frost creep, and small-scale thermokarst terrain. The deeper ice transport may affect geologic processes on hundred-meter to kilometer scales, possibly influencing features such as terrain softening, chaotic terrain, and the formation of rampart craters. The extent of the effect on martian geomorphology will depend on the magnitude of ice transport and deposition at each geographic location and will probably vary considerably from one location to the next.

A MARINE SEDIMENTARY MODEL FOR THE EVOLUTION OF THE NORTHERN PLAINS. T. J. Parker and D. S. Gorsline, Department of Geological Sciences, University of Southern California, Los Angeles CA 90089-0741, USA.

For the purpose of focusing the following discussion, we will begin with the reasonable assumption that landforms interpreted to be shorelines [1,2] can be considered to be as diagnostic of former water levels within the northern plains of Mars as lava flows are of former volcanism. Once (or if) this interpretation of these features becomes more generally accepted and the compositional and topographic datasets for Mars improve, their distribution can be used to make valuable independent estimates of the planet's water budget by providing quantitative measures of surface reservoir volumes at specific instances in martian history. We will approach the problem from a consideration of the basin's volume and its implied requirement on juvenile water sources and channel discharges, rather than the other way around.

1. *What is the volume of the basin contained within each proposed shoreline?* Surface area measurements of the northern plains are the easiest to make, but are still approximations because neither the "gradational boundary" [3] nor the "interior plains boundary" [2] can be traced to complete closure due to subsequent geologic modification or lack of images of sufficient resolution. (Note: Clear identification of shorelines in Lake Bonneville requires resolutions probably on the order of, at the very least, 50 m pixel or better, although some of the very largest barriers can

be identified in 80 m/pixel Landsat images.) The plains area contained within the gradational boundary is approximately 46×10^6 km^2, that within the interior plains boundary is approximately 27×10^6 km^2. Extending these estimates into the third dimension is very difficult due to the large errors inherent in the topographic data currently available [4]. This difficulty will probably be compounded as new data become available by effects commonly encountered (and indeed anticipated) in terrestrial paleolake studies. These include subsequent crustal warping, isostatic rebound upon removal of the water, and sediment compaction upon desiccation of the basin floor sediments. If we assume an average depth of 500 m as a reasonable upper value for the basin within the interior plains boundary, which is contemporaneous with or postdates the circum-Chryse outflow channels, we get a volume of 13.5×10^6 km^3. If the average depth was closer to 100 m, we get volumes in the neighborhood of less than 3×10^6 km^3. These values are comparable to Carr's [5] and Rotto and Tanaka's [6] estimates for the circum-Chryse channel floods (if all channels were active simultaneously). Average depths greater than 1 km may be required for the gradational boundary, which predates the latest circum-Chryse floods. This depth implies volumes above 46×10^6 km^3. Much better estimates of the basin volume can be expected once high-resolution topography from the Mars Observer Laser Altimeter (MOLA) becomes available beginning in 1994 [7].

2. *Why don't we see clear evidence for fans or deltas at the mouths of outflow channels and valley networks?* Simply stated, fluvial sedimentation into any basin on Mars would have been either subaerial or subaqueous, depending on the volume of transported material, the sediment/water ratio, the rate and duration of discharge, the number of simultaneously active channels draining into the basin, and the stability of liquid water and ice at the martian surface at the basin's latitude at the time.

Evidence of flow from the circum-Chryse channels can be traced for several hundred kilometers beyond the channel mouths to as far north as 40° latitude in Acidalia Planitia, where their deposits gradually blend with the morphology of the northern plains. The difficulty in delineating fans or deltas from the outflow channels, which represent the most recent style of fluvial discharge into the plains, and therefore the most likely to exhibit depositional landforms today, does not preclude their presence at the surface, since such deposits are difficult to delineate in terrestrial analogs due to the extreme sediment transport energies involved [8–10].

The first catastrophic flood to reach the northern plains would have cut a channel leading to the lowest point, at least locally, within the basin. That none of the outflow channels are cut far into the northern plains suggests either a base level elevated above the floor of the basin or subsequent burial. On Earth, ultimate base level is defined by sea level, where the drop in current velocity allows suspended sediment to be deposited in deltas along continental margins. (Note: Though many terrestrial rivers exhibit submarine extensions, these are leveed channels within the aggradational or progradational offshore deposits, with the exception of the relatively steep continental slope, which has no martian analog.)

Did Mars have a similarly well-defined base level? This question can be addressed by considering the differences in the ratios of channel influx to basin volume between the two planets. All terrestrial rivers and lakes combined contain less than 0.002% of the volume of water contained in the oceans [11]. Changes in sea level associated with even the largest conceivable catastrophic floods on Earth, therefore, would be insignificant. The martian outflow channels, however, transported up to 100% of the volume of any hypothetical northern plains ocean contained within the interior plains boundary. Lower percentages would require a sea or ocean (either liquid, ice-covered, or frozen) in the plains prior to flooding. This could substantially reduce the flood volume required during any given flood and cannot be ruled out, since geomorphic evidence of an earlier sea would have been erased by subsequent flood events.

In any case, catastrophic floods into the northern plains of Mars would have experienced a rapid rise and headward transgression of base level as the flood progressed. The distal ends of the channels would become buried and their mouths back-filled as the locus of deposition migrated up-channel with the rise in base level. Classical terrestrial-style delta morphology, which requires a near-equilibrium base level and a more or less steady discharge for a sustained period, could not develop under these conditions.

For the small valley networks, in which the transport energies are small but the flow durations long compared to those of the outflow channels, deposits should resemble small- to modest-scale terrestrial alluvial fans or deltas. The amount of water contributed to the northern plains by the valley networks is difficult to estimate due to the large number of individual valleys, the large topographic uncertainties and lack of knowledge of the prechannel topography (for estimating volume of transported sediment), and the difficulty of resolving and measuring the channel thalweg on the valley floors. Most of the valley networks were active during Noachian time, though several investigators are proposing that reduced activity probably continued throughout Hesperian time.

It is important to note that many valley networks and some of the older outflow channels (e.g., Mawrth, Al Qahira, and Ma'adim Valles) terminate abruptly at the gradational boundary, although regional topographic gradients continue far beyond them into the plains. To us, this suggests a common base level for these networks or a subsequent highstand of a sea or ocean within the northern plains that reworked the distal ends of the networks.

3. *How much surface water would have been available during Noachian and Hesperian time?* Terrestrial sea level fluctuates over long periods of time, due largely to astronomical forcing of climate (which is more extreme on Mars [12]). Variations in sea level are limited to a few hundred meters, however, because only a very small percentage of the 1.37×10^9 km^3 of water in the oceans can be stored in continental and polar reservoirs at any given time (Earth's continental and polar reservoirs contain $\sim 2.3 \times 10^8$ km^3 of water [11]).

It is clear that Mars' volatile reservoirs, the megaregolith and polar caps, are of sufficient volume to contain its entire water inventory (i.e., it seems there is no liquid water present at the surface today). Assuming the pore volume of $\sim 10^8$ km^3 for the megaregolith estimated by Clifford [13] is reasonable and that very little of the original water budget has been lost to space, confining his 600-m global layer equivalent to the northern plains within the earlier, gradational boundary yields an average depth of over 1.5 km. This amount is more than adequate to fill the northern plains to the gradational boundary, but only if this estimate of the planet's total inventory is approximately correct and it was on the surface at once.

The total martian water inventory is still poorly known, but the final figure is likely to be (or have been) between 1 and 2 orders of magnitude smaller than that contained within the Earth's oceans and continental/polar reservoirs. How much water could have been present at the surface during Noachian and Hesperian time depends on the rate at which juvenile and/or recycled water was produced through volcanic outgassing and heavy bombardment relative to the rate at which it would be lost to the megaregolith and cryosphere. Both processes were probably important during Noachian time, but volcanism would have been the only significant contributor during Hesperian and Amazonian time. The question of the permeability of the megaregolith is a huge unknown [14], but it probably fell with the meteorite impact flux over geologic time due to chemical weathering and cementation of surface rocks in the presence of liquid water (or ice) and formation of permafrost seams. Percolation would have been further retarded in a cold martian paleoclimate because outgassed water would simply snow out onto the surface and remain there until removal through basal melting or sublimation.

4. *What happened after mid-Amazonian time?* The development or modification of characteristic morphologies in water-lain sediments in the northern plains could have been profoundly influenced by the prevailing climate or subsequent changes in the climate. The giant polygons, for example, have been characterized as indicative of a deep permafrost layer in the northern plains [15]. Lucchitta et al. [16] favored thick sedimentary deposits, laid down in an ice-covered ocean, to explain the giant polygons, pointing out that they are found in low areas that would have collected sediment from the outflow channels. Recently, McGill and Hills [17] argued that the scales involved point to desiccation rather than ice-wedging as the formative mechanism, which could occur in either a warm or cold climate. But in a cold or cooling climate, pingos would probably develop on the plains surface once the water evaporated or sublimated and the wet sediments froze [1,18].

Conclusions: Mars lacks large surface water/ice reservoirs today (and probably at least since middle Amazonian time) mainly for these reasons: (1) Volcanic and impact production and recycling of water on Mars fell over time until it could no longer keep pace with loss to subsurface reservoirs through percolation, chemical weathering, and permafrost/polar cap formation. (2) The total water inventory remaining on the planet is less than the pore volume and water-volume equivalent of chemical weathering products in the megaregolith and the ice content of the polar caps. (3) The planet's lack of plate tectonism means that Mars has no mechanism to "rejuvenate" topographic basins, such that the basin broadens and shallows (via erosion at the margins and deposition in the interior) whenever a lake is present, resulting in eventual extinction as a playa.

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CRYOPLANATION TERRACES OF INTERIOR AND WESTERN ALASKA. R. D. Reger, Alaska Division of Geological and Geophysical Surveys, Fairbanks AK, USA.

Cryoplanation terraces are step- or tablelike residual landforms consisting of a nearly horizontal bedrock surface covered by a thin veneer of rock debris and bounded by ascending or descending scarps or both. Among examples studied, rubble-covered scarps range in height from 3 to 76 m and slope from 9° to 32°; nearly vertical scarps exist where bedrock is exposed or thinly buried. Simple transverse nivation hollows, which are occupied by large seasonal snow banks, commonly indent the lower surfaces of sharply angular ascending scarps. Terrace treads slope from 1° to 10° and commonly cut across bedrock structures such as bedding, rock contacts, foliation, joints, faults, and shear zones. Debris on terrace treads is generally 0.8–2.5 m thick. Permafrost table is generally present from 0.5 to 2 m below the tread surface. Permafrost is shallowest in the floors of nivation hollows and deepest in the well-drained margins of terrace treads. Side slopes of cryoplanation terraces are shallowly buried bedrock surfaces that are littered with a variety of mass-movement deposits.

Cryoplanation terraces are common throughout interior and western Alaska in uplands above and generally outside areas glaciated during the Wisconsin glaciation. The most sharply angular terraces are present on massive volcanic rocks, greenstones, and hornfels. Relatively few terraces with planar treads are found on finely crystalline schists, phyllites, and slates, and on fine- to medium-grained sedimentary rocks, although they are common rock types. These alpine terraces show close relations to climatic indicators such as permafrost, lowest active solifluction lobes, treeline, modern snowline, and past snowline.

Field evidence indicates that cryoplanation terraces form by a complex of mechanisms acting in unison, including nivation, mass movement, frost action, piping, and wind. The retreat of ascending scarps into alpine ridges and hills is accomplished by nivation attacking the backwalls and sidewalls of transverse nivation hollows. The debris of nivation and frost weathering is transported across the terrace tread and down side slopes by gelifluction and frost creep. Agents of terrace destruction are frost wedging, solifluction, and frost creep. These processes attack the margins of terrace treads, where the permafrost table is deepest, and progressively round the planar terrace tread. Terrace form is a function of the dynamic interaction of formational and destructional processes.

Today, the cryoplanation terraces in interior and western Alaska are inactive. Sharply angular terraces are generally considered to be Wisconsinan in age, although an angular terrace scarp at 1000 m elevation on Indian Mountain in west-central Alaska was modified by nivation at least as late as 2000 yr ago. Cryoplanation terraces are indicators of a periglacial climate and demonstrate the former presence of shallow permafrost. Comparison of terrace distribution with modern climatic data in nearby lowlands indicates that when the terraces were active, average midday temperatures during the summer were colder than 10°C, mean summers were probably about 2°–6°C, and the mean annual temperature was probably about –12°C.

THE THUMBPRINT TERRAIN: WHAT WILL MARS OBSERVER TELL US? M. W. Schaefer, University of Maryland, College Park MD 20742, USA, and NASA Goddard Space Flight Center, Greenbelt MD 20771, USA.

Some of the more puzzling features seen on Mars are those known as curvilinear features or "thumbprint" terrain, types of patterned ground found in the northern plains. The thumbprint terrain, named for its resemblance to the lines of a human thumbprint [1], is found on what appears to be level, relatively low-lying ground near the crustal dichotomy boundary. It is often found near the termini of large channels [2].

There are three types of thumbprint terrain in the classification of Rossbacher and Judson [3]. The first consists of ridges and depressions about a kilometer wide, separated by a few kilometers, and with an apparent relief of 40–100 m. The second consists of steep-sided, flat-floored depressions about a kilometer wide, separated by a few kilometers, and with an apparent relief of 10–290 m. The third consists only of albedo markings, with a typical scale of about a kilometer.

Many models have been proposed to explain the origin of this terrain. It has been suggested that it was caused by lava flows [4], removal of debris mantles [4], glaciers [6,7], or karst [8]. However, the most popular models at present involve the action of subsurface ice to form such thermokarst features as striped ground, solifluction lobes, and/or linear, ice-cored ridges [3,9].

There are several instruments on the Mars Observer spacecraft that will be able to provide us with information useful in distinguishing between these models. Obviously, the improved images of the thumbprint terrain that the Mars Observer Camera (MOC) will be able to provide will be of great use. Because this terrain is made up of roughly kilometer-scale features, only the higher-resolution Viking Orbiter images (better than 200 m/pixel) can be used to study it at present. MOC will provide images at resolutions up to about 1.4 m/pixel [10], enabling a better comparison between the martian thumbprint terrain and its possible terrestrial counterparts.

The Gamma Ray Spectrometer (GRS) may be able to detect the presence of ground ice within tens of centimeters of the surface, at latitudes within 30°–40° of the poles, and may even be able to make maps of the distribution and concentration of this ice [11] at scales of 400 km/pixel. Information about the location of ground ice, if correlated with the presence of curvilinear features, would be supportive of the thermokarst model for thumbprint terrain.

The Thermal Emission Spectrometer experiment (TES) should be particularly useful in determining the apparent mineralogy and perhaps even the rock type on which the thumbprint terrain has been developed. It is capable of identifying primary rock-forming silicates, as well as silicate and nonsilicate weathering products, salts, carbonates, nitrates, phosphates, and incorporated volatiles [12]. By this means we will be able to distinguish between such similar landforms as karst, developed on carbonate geology, and thermokarst. Thermokarst structures such as pingos that involve the presence of subsurface lenses of ice tend to develop primarily on permeable, sandy sediments [13]. It may also be possible for us to distinguish between debris blankets formed from sedimentary materials from those formed from pyroclastic deposits.

It is more difficult to distinguish between some of the other models. Lava flows may have the same mineralogy and chemical composition as rubble that glaciers have transported or that is sorted to form patterned ground. The Mars Observer Laser Altimeter (MOLA) may be helpful here. Over the regions where the thumbprint terrain is primarily found, about 25°–55°N, the average MOLA track spacing, by the end of the 687-day primary mission, is as small as 1.25 km. The along-track spacing is only 330 m, with a spot size of 140 m. The vertical precision of the altimeter measurements will be 1.5 m or less over baselines less than 100 km, and up to 30 m for longer baselines [14].

With this quality of topographic data available, detailed comparisons should be possible between the thumbprint terrain and the candidate terrestrial landform analogs, for example, terrestrial striped ground forms on surfaces with a slope of 5° or more [15]. The martian curvilinear features, on the other hand, seem to occur on level plains.

However, our present determination of the slope of those plains is not very accurate. By combining the new data available from this array of instruments, we hope to solve the mystery of the thumbprint terrain at last.

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STRATIGRAPHY OF THE MARTIAN NORTHERN PLAINS. K. L. Tanaka, U.S. Geological Survey, Flagstaff AZ 86001, USA.

Introduction: The northern plains of Mars are roughly defined as the large continuous region of lowlands that lies below martian datum, plus higher areas within the region that were built up by volcanism, sedimentation, tectonism, and impacts. These northern lowlands span about 50×10^6 km² or 35% of the planet's surface [1,2]. The age and origin of the lowlands continue to be debated by proponents of impact and tectonic explanations (see review in [3]). Geologic mapping and topical studies indicate that volcanic, fluvial, and eolian deposition have played major roles in the infilling of this vast depression [e.g., 4–6]. Periglacial, glacial, fluvial, eolian, tectonic, and impact processes have locally modified the surface. Because of the northern plains' complex history of sedimentation and modification, much of their stratigraphy has been obscured. Thus the stratigraphy developed herein is necessarily vague and provisional; it is based on various clues from within the lowlands as well as from highland areas within and bordering the plains. The results are summarized in Table 1.

Noachian and Early Hesperian: The lowlands were largely formed either during the Early Noachian by impact [7] or during the Late Noachian by tectonism [8]. In either case, considerable sedimentation during the Late Noachian and Early Hesperian is demonstrated by extensive patches of fretted and knobby terrain and many runoff and outflow channels along the highland/lowland boundary [9,10]. The volume of material eroded along the highland boundary can be roughly estimated by assuming that the eroded zone is 1 km deep [cf. 11,12], 100 km wide, and as long as the diameter of the planet; these assumptions yield $\sim 7 \times 10^5$ km^3 of eroded material, or 20 m of fluvial sediments if they were evenly deposited over the lowlands. In addition, an unknown quantity of other sedimentary material may have been derived from erosion of interior parts of the highlands. If mass wasting had been extensive within the northern plains, as indicated by vast patches of knobby terrain at Elysium and Acidalia Planitiae and Scandia Colles [5,6], then as much as a few hundred meters of Upper Noachian/Lower Hesperian sediment may have been deposited.

Evidence for volcanic resurfacing in the northern plains includes Lower Hesperian ridged plains material and lava-flow fields at or near southwestern Elysium Planitia, western Arcadia Planitia, Phlegra Montes, Acheron Fossae, and Alba Patera [4,5]. The eruptive vents include fissures and central volcanos. If such exposures were once continuous, they may have covered an area of about 10×10^6 km^2 [2]. Elsewhere, the lowlands are obscured by younger deposits. Because of the emplacement of patches of Lower Hesperian sheet flows throughout the martian highlands [9], and because of the likely prospect for relatively voluminous lowland volcanism due to magma buoyancy [cf. 13], it appears likely that Lower Hesperian flood lavas covered a sizeable fraction of the lowlands.

Late Hesperian: The northern plains were dramatically resurfaced during the Late Hesperian by catastrophic floods and debris flows into Chryse Basin. The extent of the deposition is uncertain, because neither paleoshorelines nor deposit edges can be traced very far [14,15]. Estimates of the extent of flooding, based on the topography of Chryse and Acidalia Planitiae, flood-volume estimates, and the occurrence of landforms related to deposition (such as shorelines and polygonal grooves), are as great as 5×10^6 km^2 [14,15]. The estimated amount of sediment is about 4×10^6 km^3 [16], with a thickness of about 800 m if it were distributed evenly across the proposed maximum extent of the basin (or about 100 m if it were deposited in an ocean filling the entire northern lowlands [1,17]).

Meanwhile, volcanism continued to build voluminous, broad shields on preexisting volcanic strata at Elysium Mons and Alba Patera. In addition, Late Hesperian pyroclastic volcanism at Alba Patera produced channeled slopes [18]. This volcano may have spewed large volumes of ash into the atmosphere that were deposited downwind in lowland areas.

Fluvial and volcanic deposition resurfaced much of the southern northern plains during the Late Hesperian; however, the northernmost plains were also largely resurfaced at this time [2,6]. If, as indicated by [14,15], an ocean did not form, then resurfacing of the northernmost lowlands must have been accomplished by other means. Is another explanation conceivable?

The Late Hesperian may have been a period of highly active fluvial and eolian deposition and erosion; eolian deposition is indicated by a variety of remnants of degraded and dissected deposits throughout the highlands [e.g., 19–21]. This period may have included temporary climate changes associated with release of volatiles into the atmosphere from flooding and volcanic activity. If abundant fine material had been produced by fluvial and volcanic activity, and if eolian activity had temporarily increased, then eolian deposition in the northern plains may have been extensive during the Late Hesperian.

Early and Middle Amazonian: At Elysium, complex, rugged flow and channel features, thought to consist of lahars [22] or products of subglacial volcanism [23], and associated grooved terrain cover about 2×10^6 km^2 [2]. The proposed lahars appear to be at least several hundred meters thick. Volcanism declined at Alba Patera but resurfaced most of the Olympus Mons and Amazonis Planitia regions [4]. Floods from Maja and Shalbatana Valles resurfaced part of Chryse Basin [24]. At Mangala Valles, roughly 3×10^3 km^3 of material was deposited into southern Arcadia Planitia [25]. Along the highland boundary, a thick sequence of friable material, the Medusae Fossae Formation, was emplaced ($\sim 2.5 \times 10^6$ km^2). These materials have been interpreted to be ignimbrites [26], paleopolar deposits [27], and eolian deposits [5].

Late Amazonian: Alluvium and lava flows were deposited in a broad, shallow basin in southern Elysium Planitia [5,28]; flooding breached a gap along the east edge of the basin and flooded Amazonis Planitia and possibly part of Arcadia Planitia [14]. Polar mantle and dune deposits, perhaps transient over millions of years, presently encircle the north polar layered deposits.

Summary: Since the Noachian, the northern plains have been resurfaced by fluvial, mass-wasting, eolian, and volcanic processes (Table 1). Extensive mass wasting and volcanism during the Noachian and Early Hesperian were followed by episodic fluvial and volcanic resurfacing originating from the Tharsis and Elysium regions. The latter resurfacing was extensive during the Late Hesperian, but was more sporadic and on a smaller scale during the Amazonian. Eolian reworking of deposits, particularly near the north pole, may have temporarily increased following fluvial events.

TABLE 1. Proportions of total area (50×10^6 km^2) resurfaced in the northern plains of Mars.

	Process		
Epoch	Fluvial/ mass-wasting	Eolian	Volcanic
Noachian/ Early Hesperian	nearly 1.0?	?	~0.5?
Late Hesperian	>0.15	>0.5?	~0.2
Early and Middle Amazonian	<0.05?	?	0.25
Late Amazonian	0.08	0.05	0.02

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SEISMIC-TRIGGERING HISTORY OF THE CATASTROPHIC OUTFLOWS IN THE CHRYSE REGION OF MARS.

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Introduction: Much attention has recently been focused on the Chryse outflow channels as the source for an ocean [1,2] or large lakes [3,4] in the northern plains of Mars. A critical question is whether the channels formed quickly and in concert or sporadically. Crosscutting relations demonstrate multiple episodes of channel formation for some of the channels; however, for most channels, the absence of high precision in the densities of superposed impact craters prevents precise reconstruction of the duration and sequence of channeling history [3]. Another approach to addressing the groundwater discharge history is to evaluate the hypothetical storage of the Chryse aquifer system and its recharge and triggering histories. Here we discuss how outflow breakouts may have been triggered by relatively frequent, large marsquakes caused by faulting and impact.

Previous Work: The preferential occurrence of most outflow channels around Chryse Planitia is commonly explained by invoking a regional-scale confining layer that was capable of sustaining a significant hydraulic head at low elevations. Carr [5] suggested that such a confining layer would have naturally evolved with the progressive freezing of the crust in response to the decline in the planet's early surface temperature and heat flow. The development of such an impermeable seal would have permitted the gradual buildup of hydraulic pressure as the groundwater became increasingly confined by the thickening layer of frozen ground. In addition, the tectonic uplift of Tharsis and Valles Marineris near the areas where the channels originate may have increased subpermafrost pore pressures at lower elevations. Eventually, pore-water pressure may have become superlithostatic, resulting in rupture of the confining layer and catastrophic discharge of groundwater into Chryse Basin [5].

Basis for and Description of the Proposed Model: Although the development of superhydrostatic pore pressures by tectonic uplift or progressive freezing probably has taken place, the attainment of superlithostatic pressure was not necessarily required for outbreaks to have occurred. Failure of the confining layer could also be caused by impact excavation, tectonic faulting, or volcanic eruption. Alternatively, the confining layer could be disrupted by some remote seismic event, if it were of sufficient intensity [see 6,7]. As P waves pass through the crust, they can forcefully compress water-filled fractures and pores, temporarily generating high pore pressures that can reduce intergranular friction and destroy the integrity of any confining layer. During the great Alaska earthquake of 1964 (M = 8.4), this process resulted in water and sediment ejection from shallow aquifers as far as 400 km from the earthquake's epicenter; some eruptions rose more than 30 m into the air. Seismically induced water-level fluctuations were also observed in more than 700 wells worldwide; the most notable observation was of a 2.3-m variation recorded in an artesian well in Perry, Florida, some 5500 km from the earthquake.

We propose that the Chryse outflow channels were triggered by such seismic events. The events locally disturbed a system of confined and perched aquifers in the Chryse region that had developed superhydrostatic pore-water pressures [5]. The disturbed aquifers locally underwent liquefaction or developed transient superlithostatic pore pressures, disrupting vast expanses of confined aquifer and allowing the fluidized material to be released as catastrophic floods and debris flows. Their downhill movement would then be sustained by turbulence or low internal friction at the base of the flow. Elsewhere, water (or slurry) apparently erupted through fissures, resulting in channels (commonly braided) that emanated from troughs and pits [8]. Note that in both cases, the high pore pressures associated with the seismic disturbance serve only to trigger the outflow event. All subsequent discharge is driven by whatever artesian pressure existed within the confined aquifer prior to its disruption. As the discharge of groundwater continued, the resulting decline in the local hydraulic head would have eventually reduced the flow rate to the point where the water-saturated crust would refreeze at the point of breakout, a process that would continue until the original ground-ice thickness was reestablished. Thereafter, another cycle of groundwater recharge and catastrophic outbreak could occur. Subsequent outbreaks would probably occur retrogressively along the headwalls of preexisting canyons.

Seismicity Near Chryse: Valles Marineris and Tempe Terra: The displacement of rock masses by faulting causes seismic activity, whether the displacement actually breaks the surface or is entirely restricted to the subsurface. The larger the displacement along the fault plane, the greater the seismicity. Near the outflow-channel headwater regions, the largest faults along which Late Hesperian seismic activity probably occurred are in Valles Marineris and Tempe Terra. Golombek et al. [9] estimated the frequency of martian seismicity on the basis of terrestrial oceanic earthquakes, which have an observed maximum magnitude of ~6.7. However, given the length of potential fault lines within Valles Marineris, more powerful temblors are clearly possible. Indeed, normal faulting at Valles Marineris was probably the dominant source of tectonic seismicity on Mars during the Late Hesperian. Large marsquakes ($M > 6$) probably took place once every few hundred years on average, whereas the recurrence interval for possible $M = 8$ and $M = 9$ earthquakes would have been tens to hundred of thousands of years respectively. Note that the frequency of smaller marsquakes is closely tied to the frequency and magnitude of the very largest marsquakes, because more of the accumulated crustal stress is dissipated by the larger events. Because Valles Marineris has the largest faults and cumulative fault displacement on the planet, we conclude that most of the largest marsquakes should have occurred there. In addition, the faults of Tempe Fossae could have produced large marsquakes that may have affected lower Kasei Valles in the northwestern part of the Chryse region.

Impact Seismicity: To assess the potential effect that impacts would have on subsurface water, Leyva and Clifford [6] investigated the seismic response of a basalt aquifer to the propagation of P waves generated by impacts in the 33–1000-km-diameter size range [craters with final diameters (D) of ~33 km release an amount of seismic energy equivalent to that released by the 1964 Alaska earthquake]. Their calculations of the resulting one-dimensional changes in effective stress and pore pressure that occur below the water table are based on the following assumptions: (1) All the seismic energy radiated by an impact is transmitted as a single compressional wave. (2) Both the host rock and groundwater are compressible. (3) No net flow occurs between the water-filled pores.

Their results indicate that at the maximum distance (400 km) at which water and sediment ejections were observed during the Alaska earthquake, pore-pressure changes of about 2 bar were produced in near-surface silt and clay sedimentary deposits. On Mars, an impact of equivalent seismic energy ($D \sim 33$ km) would cause this same change in pressure in a basalt aquifer out to a distance of ~100 km, whereas an impact with a diameter of ~200 km (generating an $M = 10$ marsquake) would generate similar pressures out to distances of 1000 km and a pressure of more than 10 bar at a distance of 250 km. Given a more realistic representation of seismic-wave propagation through a planetary body, these pore-pressure changes are likely to be amplified significantly as the seismic waves converge toward the impact's antipode.

On the basis of the crater-density distribution and duration of the Late Hesperian [10], we calculate that impacts produced seismic disturbances of $M > 6.2$ ($D > 5$ km) every 150,000 yr, whereas events with $M > 8.4$ ($D > 33$ km) occurred approximately every 6.4×10^6 yr. Another way to view the crater data is by the mean distance to a 33-km-diameter or larger Late Hesperian impact, which is about 300 km. Also, interspersed throughout the region of the Chryse channels are five relatively young craters (Hesperian or Amazonian) that are 100–150 km in diameter; they indicate that powerful seismic events induced by impact probably occurred during the period of channel activity. However, the only known post-Noachian multiring basins on Mars—Lowell (latitude 52°, longitude 81°), Lyot (latitude 50° N, longitude 330°), and possibly Galle (latitude 51° S, longitude 31°), about 200 km in diameter—are each more than 2000 km away from (but not antipodal to) the Chryse channels. Lowell and Lyot have been dated at Early Hesperian and Early Amazonian respectively [10].

Implications and Future Work: The source areas of the Chryse outflow channels were subjected to considerable seismic shaking during the period of channel formation. Large impacts distributed over the planet are estimated to have produced seismic disturbances of $M \geq 8$ every 10^6 to 10^7 yr, whereas those produced by faulting, particularly in the region surrounding Valles Marineris, were probably 10–100× more frequent. This rate of seismic activity suggests that thousands of possible triggering events occurred throughout the Late Hesperian, if the depleted local inventory of groundwater could be replenished at a sufficient rate. Depending on such variables as the total volume of water discharged by the channels, the number and size of the discharging regions, the duration of the discharge, the average value of crustal permeability, the extent of hydraulic continuity, the large-scale permeability of the crust, and the size of the available reservoir of groundwater that can be tapped by the channels, we find that the timescale for recovery of the local water table to its predischarge level could range anywhere from less than 1 yr to as long as 10^6 yr. The huge size of the channels indicates that the events generally involved fully charged aquifers. Thus, our preliminary assessment is that hundreds to thousands of seismically induced outflow events were possible.

This history makes possible the following scenarios for outflow channel formation: (1) multiple floods in which moderate (vs. hyperconcentrated) sediment loads were carried and (2) origination and enlargement of the channels chiefly by a succession of retrogressive debris flows. In either case, present channel forms probably result from the cumulative erosion of many events over a prolonged period of volcanotectonic activity at Valles Marineris and, less importantly, Tharsis Montes and Tempe Terra. These scenarios are more consistent with the associated development of periodic ice-covered lakes and ice sheets in the Chryse lowlands, rather than a northern plains-filling ocean.

The semiquantitative model presented here is still being formulated. Future refinements to the seismic modeling portion of this analysis will include a more realistic representation of seismic-wave production and propagation, as well as a consideration of the effects of flow in a heterogeneous aquifer with both intergranular and fracture porosity. There are other questions we hope to address: Why did chaotic terrain and channel troughs develop in some areas but not others? How are structural controls manifested? Is there any geomorphic evidence, beyond the channels themselves, that either supports or constrains the model presented here? And, finally, if this model is correct, why was channel activity not even more pronounced earlier in martian geologic history, when the frequency and magnitude of tectonic and impact-generated seismic events was probably orders of magnitude greater? Work on these and other questions is currently in progress.

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A FORMATIONAL MODEL FOR THE POLYGONAL TERRAINS OF MARS: TAKING A "CRACK" AT THE GENESIS OF THE MARTIAN POLYGONS. M. L. Wenrich and P. R. Christensen, Arizona State University, Tempe AZ 85287-1404, USA.

The mechanism for the genesis of the polygonal terrains in Acidalia and Utopia Planitia has long been sought; however, no completely satisfying model has been put forth that characterizes the evolution of these complexly patterned terrains. The polygons are roughly hexagonal but some are not entirely enclosed by fractures. These polygonal features range in widths from ~5 to 20 km [1]. Several origins have been proposed that describe the polygon borders as desiccation cracks [2], columnar jointing in a cooled lava [3], or frost-wedge features [4]. These tension-induced cracking hypotheses have been addressed by Pechmann [5], who convincingly disputes these mechanisms of formation based on scale magnitude difficulties and morphology. Pechmann [5] suggests instead that the cracks delineating the 5–20-km-wide polygons on the northern plains of Mars are graben resulting from deep-seated, uniform, horizontal tension. The difficulty with this hypothesis is that no analogous polygonal forms are known to have originated by tectonism on Earth. McGill and Hills [1] propose that the polygonal terrains on Mars resulted from either rapid desiccation of sediments or cooling of volcanics coupled with differential compaction of the material over a buried irregular topographic surface. They suggest that fracturing was enhanced over the areas of positive relief and was suppressed above the topographic lows. McGill and Hills [1] suggest that the spacing of the topographic highs primarily controls the size of the martian polygons and the physics of the shrinkage process is a secondary concern.

Ray et al. [6,7] conducted a terrestrial study of patterned ground in periglacial areas of the U.S. to determine the process responsible for polygonal ground formation. They developed a model for polygon formation in which convection of seasonal melt water above a permafrost layer, driven by an unstable density stratification, differentially melts the permafrost interface, causing it to become undulatory (Fig. 1). They suggest that circulation cells produce the subsurface undulations whose spacing drives the wavelength spacing of the overlying polygons. They tested their model at 17 field locations throughout the northwest U.S. by trenching to observe polygons in three dimensions. The result was that the dimensions of the polygons calculated by the free convection model agreed remarkably well with those in the field.

The Ray et al. free convection model requires (1) a saturated porous medium, (2) a free upper surface that is maintained at a constant temperature of 277 K (mathematically this is consistent with a saturated melt water surface at or below the ground surface), and (3) an impermeable lower boundary that moves downward as melting occurs. The lower surface represents the melt water/permafrost phase-transition interface and is therefore the 273 K isotherm. Other model requirements are that the lower boundary is allowed to melt differentially and no heat is conducted into the solid phase (i.e., all the heat is used for the phase transition).

Rayleigh convection describes the movement of a fluid that is unstably stratified as a result of thermal gradients. In this case, water at a temperature of 277 K is 0.0132% more dense than at 273 K [6,7]. As solar radiation warms the surface of a frozen soil, the denser, warmer surface water will sink, if adequately perturbed, displacing colder, less dense water to the surface. The exchange of fluid is most efficient when small convection cells are established; thus, many convection cells are formed rather than just one large cell [8]. Melting of the permafrost will be enhanced beneath the downwelling warm water relative to the area beneath the cooler, upwelling water. This differential melting creates a scalloped texture on the subsurface permafrost layer that defines the scale of the surficial polygons. This two-dimensional free-convection model is applicable to large polygons related to seasonal fluctuations as well as to smaller polygons resulting from diurnal freeze-thaw cycles. The model predicts a width-to-depth ratio of the polygonal features to be 3.81 with a critical Rayleigh number of 27.1 (discussed in detail below). This width-to-depth value corresponds exceptionally well with the field data. The field ratio was determined to be 3.57, remarkably similar to the theoretical value.

The first step of the terrestrial model is to examine linear stability theory to determine if a small disturbance of the unstable system will die out or grow into a periodic form. If the perturbation will neither grow nor decay, it is considered to be neutrally stable. This behavior is traced on Fig. 2 by the neutral stability curve; the "stable region" represents the conditions under which decay of the perturbation will occur. Conversely, conditions that plot within the "unstable region" will cause the perturbation to grow and initiate convection of the system. The system is described by a set of

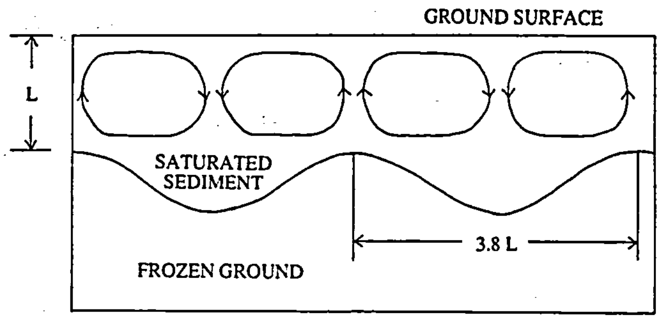

Fig. 1. A schematic two-dimensional cross section of hexagonal Rayleigh convection cells in an active layer. (After [6].)

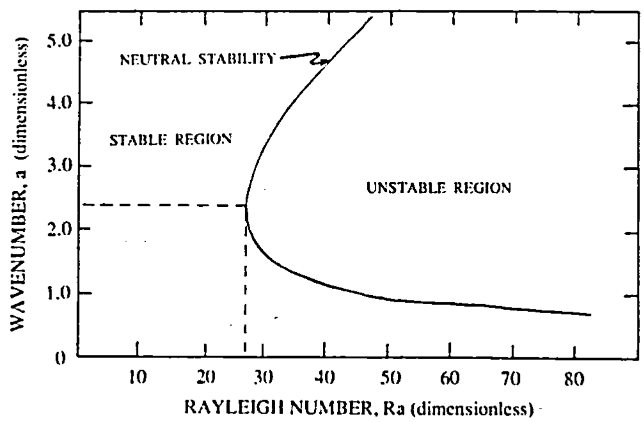

Fig. 2. Wavenumber vs. Rayleigh number. Critical Rayleigh number is 27.1 and critical wave number is 2.33. (After [6].)

boundary conditions that define the relationship between the dimensionless Rayleigh number, Ra, and the dimensionless wave number, a. The critical Rayleigh number is the ratio of buoyancy forces to viscous forces and defines the unique point at which convection initiates. The critical wave number is the unique point that determines the dimensions of the convection cell. In an isotropic medium, a can be defined as

$$a^2 = l^2 + m^2 \tag{1}$$

where l and m are dimensionless horizontal wave numbers. The symbols l and m are proportional to the depth (L) to width (W) of the threshold disturbance size at neutral stability in the x and y directions (W_x, W_y) respectively. These parameters are defined as

$$l = \frac{2\pi L}{W_x} \tag{2}$$

and

$$m = \frac{2\pi L}{W_y} \tag{3}$$

In an isotropic medium, l = m. Therefore, substituting these values into equation (1) gives

$$a = \sqrt{2}\left(\frac{2\pi L}{W}\right) \tag{4}$$

For the specific boundary conditions of an impermeable, isothermal upper boundary and a impermeable, constant heat flux lower boundary, the free convection model determines the values of a and Ra to be 2.33 and 27.1 respectively. Substituting the value of a into equation (4) allows us to calculate the width-to-depth ratio for the convection cells as

$$a = \sqrt{2}\left(\frac{2\pi L}{W}\right)$$

and so

$$W = \sqrt{2}\left(\frac{2\pi L}{2.33}\right) \tag{5}$$

Thus W = 3.81 L.

Using the width values for the Mars polygons and the aspect ratio predicted by the model, the depth of the active layer for onset of convection was determined to be ~1.31–5.25 km. These width and depth dimensions represent the wavelength spacing of the undulations and the depth to the subsurface permafrost layer.

With the depth of the active layer calculated to be ~1.31–5.25 km, an investigation into the viability of the free convection model was conducted. In order to use the model a periodic freeze-thaw cycle on Mars is required. Fanale et al. [9] estimate that the annual thermal wave for ~45°N latitude (where the polygons are approximately located) penetrates the surface to approximately 1 m. Nummedal [10] estimates this value to be 1–10 m. These values are 3 orders of magnitude less than required for the active layer depths predicted above by the free-convection model. Additionally, the thermal wave resulting from Mars' obliquity cycle of 1.2×10^6 yr has been modeled by Coradini and Flamini [11]. They found the thaw front to reach a maximum depth of 100 m. These data suggest that no periodic freeze-thaw cycles penetrate to the depths required for melt-water convection necessary for the free-convection model.

We suggest, however, that the requirement of thaw to kilometer-scale depths predicted by the Ray et al. model can be met by introducing a preexisting "thawed" active layer that is derived from a catastrophic process on Mars. Outflow channels on Mars are thought to have formed by catastrophic release of groundwater when artesian pressure exceeded the lithostatic pressure supplied by a superposing impermeable permafrost layer [10]. The morphology of the outflow channels suggest northward transport of immense volumes of fluid and sediment that may have pooled within Mars' northern basins such as Acidalia and Utopia Planitia [12]. Estimates of the sediment thickness of the outflow deposits are suggested to be roughly on the order of hundreds of meters [1]. We suggest that this warm, wet package of sediment satisfied the requirement of a saturated kilometer-scale active layer necessary for convection to initiate. Convection within the active layer would have produced an undulatory subsurface geometry of the underlying permafrost layer as described by the Ray et al. model above. After deposition the overlying sediments would have undergone compaction (presumably due to loss of water from the system) and bending stresses (which will produce drape folds on the flanks of the undulations). The greatest tensile stresses in the sediment cover produced by gravity and bending would have occurred above the crests of the scallops due to sediment moving downslope. Thus weaknesses in the sediments were maintained above the topographic highs of the subsurface. Desiccation of the wet sediments will cause soil to shrink uniformly; however, we suggest that because of the preexisting weaknesses in the sediment cover, the resultant shrinkage fractures were concentrated above the crests of the undulations. To summarize, we suggest that the scalloped character of the permafrost defined the wavelength spacing of the polygonal features on the surface.

In conclusion, using the Ray et al. model [6,7], the active layer for the 5–20-km-diameter martian polygonal features is calculated to have been 1.31–5.25 km thick. The active layer thickness required is much greater than even the obliquity-driven freeze-thaw cycle can explain. If, however, the outflow channels deposited kilometer-thick sediment deposits in the northern basins, the initial requirements of a saturated active layer would have been satisfied for convection to initiate. The wavelength spacing of the undulatory subsurface permafrost boundary would have controlled the position of the fractures that resulted from desiccation of the overlying sediment package. The undulatory permafrost interface predicted by free convection would have promoted fracturing of the ground surface preferentially above the subsurface topographic highs. This mechanism is similar to McGill and Hills' irregular subsurface topography model; however, in the free-convection model, the spacing is influenced by the dimensions of the convection cells rather than crater locations.

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BALLOON EXPLORATION OF THE NORTHERN PLAINS OF MARS NEAR AND NORTH OF THE VIKING 2 LANDING SITE. F. R. West, 520 Diller Road, Hanover PA 17331-4805, USA.

The next stage of exploratory surveying of the northern plains of Mars after the Mars Observer mission may best be done by Mars Balloon Exploration Vehicles (MBEVs) deployed in the atmosphere above the northern plains near and north of the Viking 2 lander (VL-2) landing site.

This region of Mars is favorable for exploration by MBEVs for the following reasons: (1) The low surface elevation [~–2km, lower than the standard (zero) elevation of the martian surface] provides atmospheric pressure sufficient to allow a MBEV to successfully operate and explore this region at the present stage of MBEV design and development. (2) The relatively smooth nature of the northern plains terrain (few mountains, ridges, valleys, and craters) indicated so far by Mariner 9 and Viking Orbiter data seems adequate for operation of a MBEV when its SNAKE extension is in contact with the surface of one of these plains.

For some details of the design, testing, and planned construction of joint Russian (IKI)-French (CNES)-Planetary Society MBEVs now scheduled for first deployment in the martian atmosphere in September 1997, see [1].

Science Objectives: 1. To map in detail Utopia Planitia and Vastatis Borealis near and north of the VL-2 landing site (225.7°W longitude, 47.96°N latitude). One of the main objectives of this mapping should be to determine if there are fundamental differences between Utopia Planitia and Vastatis Borealis, such as the abundance of rocks and boulders, the existence of sand dunes, etc. If these regions are found to be quite different, is there evidence for a definite boundary between them or is there a gradual transition? Are there surface formations that can be attributed to water ice or permafrost?

2. To study the formation and disappearance of the seasonal north polar ice cap during autumn, winter, and spring. Does the CO_2 condense out of the atmosphere directly onto the surface of the plains or does a CO_2 snow form in the martian atmosphere that then falls onto the surface? Is this seasonal ice cap essentially all frozen CO_2 or is a small part of it frozen H_2O, which appears to be the main component of the frost observed at the VL-2 landing site in autumn and winter [2]?

What is the relationship of the seasonal north polar ice cap to the atmospheric polar hood observed in autumn and winter? Finally, if parts of Vastatis Borealis that are sometimes covered by the seasonal ice cap (north of 65°N) are studied by MBEVs, are there observed seasonal changes on the surface that might be caused by interactions between the surface and either the atmosphere or the ice cap?

3. In spring and summer, to survey the region transversed for such evidence of near-surface water ice or permafrost as morning fog, infrared spectroscopic absorption bands of water ice, etc. If its mass (23.2 kg) can be included in the scientific payload, a gamma ray spectrometer (GRS) of the type onboard Mars Observer, which is capable of detecting the gamma rays with 2.223 MeV energies emitted upon capture by H nuclei in surface or near-surface water ice of neutrons produced by cosmic rays, should be carried by at least one MBEV. A GRS onboard a MBEV could locate concentrations of near-surface water ice on Mars much more accurately than the GRS on Mars Observer, 370–430 km above the surface of Mars [3].

4. To search for evidence of deep sand dunes between the VL-2 site and circumpolar sand dunes that are already known.

5. To study atmospheric properties and motions in the lowest kilometer of the martian atmosphere at different seasons.

Specific observations are needed to determine (1) the near-surface wind speed and direction in this region of Mars during different seasons; (2) how the polar hood in northern hemisphere autumn and winter affects insolation, diurnal temperature change, and visibility over a broad wavelength range near the surface; and (3) the amount of dust in the atmosphere above different regions at different seasons, and how these data correlate to dust and sand observed on the surface.

There are potential obstacles to the effective operation of MBEVs over the high-latitude northern plains of Mars: (1) The duration of daylight will be short in northern-hemisphere autumn and winter. A MBEV will be heated by sunlight for only a few hours each day, causing its flight duration to be brief and of short range, followed by extended contact of its SNAKE with the martian surface at night. (2) In these seasons, the polar hood will further reduce the diurnal amplitude of temperature change of the balloon gases, thereby shortening flight duration even more. (3) We do not know how the polar hood affects visibility in the lowest kilometer of the martian atmosphere.

The above potential problems warrant detailed calculations to be made of the duration of daylight for the range of latitudes from 45°N to 75°N for a plausible season for MBEV exploration following each launch window to Mars from 1999 to 2014. Estimates of atmospheric pressure also need to be made for each of these seasons.

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GEOMORPHIC EVIDENCE FOR AN EOLIAN CONTRIBUTION TO THE FORMATION OF THE MARTIAN NORTHERN PLAINS. J. R. Zimbelman, Center for Earth and Planetary Studies, National Air and Space Museum, Smithsonian Institution, Washington DC 20560, USA.

The northern plains of Mars have many morphologic characteristics that are uncommon or absent on the rest of the planet. Mariner 9 [1,2] and Viking [3,4] images obtained north of latitude 30°N revealed "smooth" and "mottled" plains of an uncertain origin. Some or all of the northern plains have been interpreted to consist of lava plains intermixed with eolian and volcanic materials [5–7], thick eolian mantles that buried portions of the mid latitudes [1,8], periglacial deposits resulting from the presence of ground ice [3,9–11], and as water-transported sediments derived from fluvial runoff [12,13], lacustrine deposition in standing bodies of water [14–16], or glacial runoff [17]. The highest-resolution Viking images show many intriguing details that may provide clues to the origin of this complex and distinctive terrain. This abstract reviews some of the informative features present in the best Viking images, comparing the observations to what may be expected from various hypotheses of formation. While the results are not conclusive for any single hypothesis, eolian processes have played a major role in the erosion (and possibly deposition) of the materials that make up the surface exposures in the martian northern plains.

The high inclination of the Viking 2 Orbiter provided the best coverage of the northern plains, particularly after the periapsis was lowered to 300 km on orbit 431 [18]. Table 1 lists the highest-resolution Viking 2 images that show portions of the northern plains, where image resolution ranges from 8 to 17 m/pixel. Two global dust storms that occurred during the Viking Extended Mission severely affected the image quality [19]; Table 1 lists only images that show identifiable morphology.

The dominant characteristic of the high-resolution images of the northern plains is that of a strongly scoured or eroded surface. This is consistent with the prevalence of eolian erosion in high-resolution images from many areas on Mars [20,21]. Evidence of eolian erosion does not necessarily imply that the eroded materials themselves are eolian in origin [11], but the implied scale dependence of geologic interpretation of images must not result in a bias against high-resolution images as lacking in useful geologic information. Along with the eroded nature of the visible surfaces, the abundance of multiple layers exposed within the eroded materials must be considered [8]. Various exposures indicate that from three to more than eight separate layers are present within scoured exposures of northern plains material that is most likely on the order of 50 m in thickness (see 461B and 466B images in particular). Such layering could result from either water-transported or eolian deposition, but it is highly unlikely that the layering would be preserved in deposits that have been subjected to extensive mobilization by periglacial processes (such motion would certainly disrupt any internal organization of the material that existed prior to the mobilization).

TABLE 1. Instructive high-resolution Viking images of the martian northern plains.

Frame Number	Location	Region Name	Range
425B01-16	42°N,272°W	Utopia Planitia	>650 km
430B01-28	32°N,247°W	Utopia/Elysium Planitia	>660 km
432B01-28	30°N,235°W	Utopia/Elysium Planitia	>350 km
434B01-14	34°N,216°W	Utopia Planitia	>350 km
454B01-11,21 32	44°N,24°W	Acidalia Planitia	>480 km
455B01-14,21 36	35°N,16°W	Acidalia Planitia	>360 km
456B01-16	41°N,5°W	Acidalia Planitia	>380 km
458B41-51,61 72	44°N,346°W	Deuteronilus Mensae	>490 km
459B01-30	48°N,334°W	Lyot Crater	>430 km
461B01-20	46°N,318°W	Protonilus Mensae	>510 km
466B81-98	44°N,270°W	Utopia Planitia	>490 km
470B01-10	45°N,277°W	Utopia Planitia, near VL2	>620 km
472B81-87	34°N,219°W	Utopia Planitia	>620 km
476B21-28	35°N,180°W	Arcadia Planitia	>380 km

Stratigraphic relationships visible in the high-resolution images also provide useful constraints on the processes that may have been active in the area. For example, the scoured appearance of some valley-fill material (461B09 and 14) is essentially identical to the scoured appearance of the surrounding plains (461B06), with both areas showing the disruption of the surface with the characteristics of a single layer of material. At other locations a single layer is superposed on top of ejecta scour from a nearby impact crater (425B11), clearly postdating the impact events of the underlying bedrock. These examples indicate that the upper surface materials in the northern plains are probably separable from the underlying plains, and probably resulted from processes distinct from those that formed the plains themselves. Is this surface material simply a "coating" that blankets the "real geology," or is it a material unit with its own geologic history? The observable thickness of the material, along with the layering evident at some locations, indicates that the material must not be disregarded as a mere obstruction to the "true geologic history" of this part of the planet. It is also clear that this material was emplaced relatively late in the history of the area, so viable processes must be capable of emplacement in recent martian history.

The visible features on the surface of the northern plains do not include something that is definitive of a single process of formation. The best Viking images are consistent with any relatively recent process that can superpose one (or more) layers of material on the local terrain. The emplacement process is sufficiently quiescent that fairly delicate preexisting morphology (ejecta scour) is not destroyed by either the emplacement or the removal of the material. These conditions may be met by either an eolian or a water-transported (fluvial/lacustrine) depositional process, as long as martian climatic conditions favorable to the preservation of standing bodies of water can be achieved late in martian history.

What might be the best way to test which hypothesis is the most likely explanation for the northern plains surface materials? Both types of deposits would probably consist of relatively small particles (capable of suspension or transport through either atmosphere or water), with the larger (greater than pebble-sized) particles left close to the respective source regions. Also, composition is not likely to vary greatly for small particles transported by either medium. Thermal measurements obtained by the Thermal Emission Spectrometer on Mars Observer would therefore not be well suited to distinguish between the two alternatives. Morphology visible in the Mars Observer Camera high-resolution images may resolve depositional landforms such as cross-bedding, but it is not clear that even these characteristics would be diagnostic of atmospheric vs. water-transported deposition. Perhaps the most distinguishable measurement for identifying the process responsible for the northern plains material will be accurate elevations obtained by the Mars Observer Laser Altimeter; sediments emplaced within standing bodies of water will of necessity be constrained below whatever elevation represented the shoreline, while sediments derived from either atmospheric suspension or saltation should not be constrained to occur within any specific elevation range. Such measurements should be made beginning later in 1993, providing the all-too-rare opportunity to test between working hypotheses for processes inferred to be active on another planet.

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