Extraterrestrial coastal geomorphology

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Abstract

Earth is the only planet in the solar system where large amounts of liquid water have been stable at the surface throughout geologic time. This unique trait has resulted in the production of characteristic landforms and massive accumulations of aqueous sediments, as well as enabled the evolution of advanced and diverse forms of life. But while Earth is the only planet with large bodies of water on its surface today, Venus and Mars may have once had lakes or oceans as well. More exotic fluids may be stable in the outer solar system. Prior to the Voyager flybys of the outer planets during the 1970s and 1980s, the moon of Neptune, Triton, was thought to be much larger than the Voyager cameras revealed it to be, and predictions that liquid nitrogen lakes or oceans might be found were made. The moon of Saturn, Titan, however, was found to have a massive atmosphere, so the possibility remains that it may have, or may once have had, lakes or oceans of liquid hydrocarbons. The recent, high-resolution synthetic aperture radar imaging of Venus has failed to reveal any evidence of any putative clement period, but the results for Mars are much more intriguing. Herein, we briefly review work on this subject by a number of investigators, and discuss problems of identifying and recognizing martian landforms as lacustrine or marine. In addition, we present additional examples of possible martian coastal landforms. The former presence of lakes or oceans on Mars has profound implications with regard to the climate history of that planet. © 2001 Published by Elsevier Science B.V.

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1. Introduction

Beginning with the first deep space missions of the early 1960s, our geomorphic horizons have been expanded beyond Earth to include hundreds of millions of square kilometers of previously unknown, sometimes quite strange surfaces on other solid bodies in our solar system from Mercury to the moon of Neptune, Triton. Pluto remains the only major planet not visited by spacecraft to date. We have seen that impact cratering has been one of the most important processes operating throughout the history of the solar system. Thirty years ago, however, debate raged as to whether lunar craters were of impact or volcanic origin (e.g., Green, 1971). Evidence for other, more familiar processes such as volcanism, was quickly recognized on planetary surfaces, though the
tremendous size of many volcanic features on Venus, Mars, and the moon of Jupiter, Io, has been surprising. Structural features such as fault scarps and grabens had been recognized on the moon soon after telescopes were able to resolve them, so they were quickly recognized on other planetary surfaces as well, once spacecraft images became available.

Differences in the sizes of familiar features on other planets can usually be attributed to the astronomical, geophysical, or climatological conditions specific to a particular planetary body. For example, the moon of Jupiter, Io, appears to experience global volcanic resurfacing because its interior is kept hot through tidal friction caused by gravitational tugging between Jupiter and the other Galilean satellites (Gailitis, 1982), and yet it is comparable in size to our own moon, on which regional-scale volcanism appears to have ceased over 3 billion years ago (e.g., Wilhelms, 1987). The largest shield volcanoes in the solar system have developed on Mars because that planet lacks terrestrial-style plate tectonism, such that the lithosphere cannot move relative to mantle hot spots, and because its thin atmosphere and cold, dry surface environment preclude terrestrial rates of erosion. At the same time, Mars is scarred by gigantic flood channels that require up to three orders of magnitude greater discharges than the largest known similar floods on Earth, and by abundant valley networks that suggest long-term groundwater sapping and surface runoff (e.g., Mars Channel Working Group, 1983). These hark back to an earlier, wetter climate, and also attest to the extremely slow destruction of landforms on Mars over much of geologic time when compared to terrestrial rates of erosion.

For many terrestrial surface processes, analogous processes have been identified, or at least inferred, on other planetary surfaces. Earth is unique in the solar system, however, in one important aspect, in that two thirds of its entire surface is covered by water. Because of the atmospheric pressure and distance from the sun, water can exist in all three physical states somewhere on the surface or in the atmosphere of Earth; at least, enough water to insure that not all can be contained in the subsurface. The presence of so much surface water in the oceans throughout geologic time, and in continental basins for geologically significant periods of time, has led to the accumulation of massive marine and lacustrine sedimentary deposits and characteristic recent and modern coastal landforms. But has Earth always been the only planet with standing bodies of water? What about other, exotic fluids that might be stable in liquid form elsewhere in the solar system? Could these mimic the behavior of water on Earth and collect into lakes, seas, or even oceans?

In this paper, we take a look at recent discussions about and geomorphic evidence for extraterrestrial lacustrine or marine environments in the current literature and comment on the potential for further discoveries. We are interested in lakes/oceans in the “traditional” sense (to the extent that there is one), and so we will exclude lava lakes from this discussion.

2. Narrowing the field

We shall limit the scope of this study to solid-surface bodies that have or may have had atmospheres in which some abundant liquid might have been stable long enough to collect into topographic depressions and leave characteristic geomorphic signatures. Further limiting this study to objects whose surfaces have been imaged leaves Venus, Earth, Mars, the moon of Saturn, Titan, the moon of Neptune, Triton, and Pluto.

3. The outer solar system

On the icy satellites of the outer planets and on Pluto, water ice is an important lithic component. Indeed, many landforms seen during the Voyager flybys of the satellites of outer planets have been interpreted as produced by eruption of cryogenic solutions containing water as a volcanic melt, to which the term “cryovolcanism” has been applied (e.g., Helfenstein et al., 1992). These cryogenic solutions might temporarily collect in a depression or caldera on one of these satellites, as appears to have happened on Triton, but these are lava lakes in the strictest sense. Extant lakes composed of liquids stable in the outer solar system have yet to be seen directly, though there may be good reason to suspect
a presence on Titan. In the absence of extant lakes, it is difficult to predict what paleolakeshores might look like on one of these satellites, because we have no experience with interactions between large bodies of liquids and icy terrain that would be stable at these very cold temperatures; constructing physical models, therefore, would also be difficult. Nevertheless, predictions of lakes or paleolakes have been made for Titan and Triton. Pluto is thought to be very similar in size and composition to Triton, and its distance from the sun is comparable, so we will accept the current hypotheses that predict a similar surface environment, at least until spacecraft data become available.

3.1. Titan

At 5150 km in diameter, Titan is the second largest moon in the solar system (the moon of Jupiter, Ganymede, is 5262 km in diameter). The existence of an atmosphere on Titan had been known since Kuiper was able to detect gaseous methane in its spectrum in 1944 (Kuiper, 1944). Prior to the Voyager flybys, the density of this atmosphere was generally thought to be only on the order of a few times as high as the atmosphere of Mars, or a few tens of millibars (e.g., Kaufmann, 1978), though Huntten (1978) correctly predicted that nitrogen might be a primary constituent of a massive atmosphere. One of the hopes of the Voyager science team was to image the surface of Titan during the flybys, though at least a partial, and possibly a global, cloud cover had been anticipated based on the very small variations in the brightness of Titan viewed from Earth as it rotates. Titan turned out to have an impenetrable, global photochemical haze of organic compounds, including ethane and acetylene, so its surface could not be imaged at optical wavelengths from orbit. More important gaseous nitrogen, which was not detected spectroscopically from Earth, was found to be a major component of the Titan atmosphere, with the result that the surface pressures are actually on the order of one and a half to three times as high as those on Earth (e.g., Lunine, 1993).

Several investigators have proposed, based on this discovery and models of the evolution of the surface and atmosphere of Titan through the history of the solar system, that large bodies of liquid hydrocarbons, perhaps even a global ocean, might be found on the surface today, or that evidence for paleolakes might be found (e.g., Caldwell, 1978; Hunten, 1992; Flasar, 1983; Lunine et al., 1983; Lunine, 1993). Lunine’s (1993) summary of this work points to the fascinating probability that one, or parts of all these models will be verified during the course of the Cassini Mission to Saturn. Recent earth-based observations at radar wavelengths (Muhleman et al., 1990) and with the Hubble Space Telescope in the near infrared (Smith et al., 1996), however, have revealed large regions with topographic or albedo variations, thus ruling out the hypothesis of a global ocean. Tidal models of the evolution of the orbit of Titan would also seem to rule out the global ocean, but disconnected seas or lakes remain a possibility (Dermott and Sagan, 1995; Sears, 1995; Sohl et al., 1995). Currently enroute (Lebretson and Matson, 1992), Cassini includes the Saturn Orbiter, which will make numerous flybys of Titan employing synthetic aperture radar (SAR), and the Huygens atmospheric probe equipped with a spectral radiometer, which will be deployed into the atmosphere of Titan.

3.2. Triton

In the case of Triton, speculations that lakes of liquid nitrogen might be found were made prior to the Voyager flyby in 1989 (e.g., Cruikshank et al., 1984, 1988). These were based on the assumptions made, before Voyager, that Triton was much larger than it turned out to be, perhaps even nearly as large as Mars (Kaufmann, 1978), and that a greenhouse effect kept surface temperatures warm enough to allow liquid nitrogen to be stable. These early estimates that Triton might be the largest satellite in the solar system were made based on its brightness and an assumed range of “reasonable” albedos. Triton turned out, however, to be one of the more reflective moons in the solar system such that, at only 2706 km in diameter, it is less than half of the diameter of Mars and is actually smaller than our own moon (3476 km). For these reasons, instead of a substantial atmosphere and greenhouse effect, the atmosphere of Triton is actually on the order of a few microbars (e.g., Brown et al., 1991). As a result, nitrogen is stable only as an ice or gas at the surface, though a number of active nitrogen geysers were seen by
Voyager (e.g., Lee et al., 1991). This evidence that Triton is geologically active today, such that volatiles are being injected into the atmosphere, suggests the possibility of a denser ancient atmosphere with an associated greenhouse effect.

3.3. Pluto

The similarity of Pluto to Triton was inferred based on the information obtained about size and surface temperatures of Triton during the Voyager flyby. The dimensions of Pluto had been fairly accurately known for several years longer than those for Triton, however. The discovery of the satellite of Pluto, Charon, in the late 1970s, and recent observations of mutual occultations, has enabled accurate determinations of the masses and dimensions of the two bodies. Pluto is approximately 2300 km in diameter, whereas Charon is about 1200 km in diameter, making it the largest moon in the solar system in proportion to its primary. The occultation studies have even allowed crude global albedo maps to be constructed of Pluto and Charon (e.g., Buie et al., 1992). Like that of Triton, the atmosphere of Pluto is quite tenuous, and the principal volatiles probably partly condense seasonally onto the surface as solids. (Pluto is currently closer to the sun in its 248.5-year orbit than Neptune.) At only 41–42 K, the surface of Pluto is far too cold for liquid N$_2$, and pure He and H$_2$ are not likely to be present in significant quantities to form lakes. It would seem unlikely, therefore, that we can expect to find bodies of liquids on Pluto once spacecraft data become available sometime next century (Staehle et al., 1993). Like Triton, however, ancient greenhouse atmospheres remain a possibility for Pluto, so evidence of ancient lakes may eventually be discovered.

4. The inner solar system

Though only the atmosphere of Earth is the right temperature and density today to permit liquid water, the ancient atmospheres of Venus and Mars may have allowed liquid water as well. But these planets lack surface water today, so verification of the former existence of lakes or oceans requires geomorphic or stratigraphic evidence. Searches for geomorphic evidence can be made with images acquired from orbiting spacecraft, provided the pixel scale (and the landform) is large enough to enable identification. Stratigraphic evidence will probably require confirmation by surface landers, though it might be recognizable in erosional scarps and hollows from orbit or in multi-spectral orbiter images that provide unequivocal compositional information.

4.1. Venus

Venus presents us with the opposite extreme in surface conditions from the outer planetary satellites. Whereas aqueous melts appear to be volcanic fluids in the outer solar system, silicate melts on Venus appear to have carved long, meandering channels, including the longest known channel in the solar system, Baltis Vallis (formerly Hildr Fossa; e.g., Baker et al., 1992). Venus has a massive, dominantly CO$_2$ atmosphere, producing a substantial greenhouse. Surface pressures are as high as 90 bars or more and temperatures are on the order of 750 K, hotter even than daytime equatorial temperatures on Mercury. Metals like lead and zinc would be molten at the surface of Venus, as would sulfur and many sulfur compounds (Baker et al., 1992).

The Magellan Orbiter mission used synthetic aperture radar (SAR) to map more than 98% of the surface of Venus at resolutions between 120 and 350 m/pixel (such that more of the surface of Venus is covered at this scale compared to the surface Earth). These data reveal a surface dominated by volcanism and tectonism, with evidence of limited impact cratering, eolian sediment transport, and chemical weathering. Though there had been hopes that ancient surfaces might retain a record of a past clement epoch on Venus, when the atmospheric pressure may have been lower (in the present surface pressure environment, water boils around 570 K) and liquid water may have been available to carve channels and coastal landforms, none have been found. The number and apparently uniform global distribution of impact craters seems to suggest a surface age on the order of several hundred millions of years or less (Phillips et al., 1992; Schaber et al., 1992).

Planetary evolution models have placed a water-rich Venus at billions of years ago (e.g., Matsui and Abe, 1986; Kasting, 1989; Grinspoon, 1993). High deuterium/hydrogen ratios in the upper atmosphere
of Venus imply either early abundant surface water (e.g., Donahue and Hodges, 1992; Hartle et al., 1996), on the order of lakes or oceans worth, or recent replenishment from the mantle or cometary impacts (Grinspoon, 1993). If Venus ever did possess an Earth-like climate, we will probably have to search for the evidence in the stratigraphic record. Even coarse stratigraphic resolution will probably require complex extended missions by a surface rover, or perhaps an intermediate-stage, multi-spectral aerial reconnaissance imaging mission. Still, the implication that Venus may have lost its oceans because of a runaway greenhouse may have a profound impact on our understanding of how the climate of Earth has evolved and might evolve in the future. The development of search strategies for indicators of former aqueous processes on Venus should at least be considered for follow-on surface science missions to that planet.

4.2. Mars

In terms of environmental properties, the present-day Mars is the most Earth-like planet in the solar system. Because the atmosphere of Mars cannot support liquid water today, several mechanisms for producing observed channels were suggested that do not necessarily require liquid water. Proposed fluids ranged from wind (Cutts and Blasius, 1981) to lava (Carr, 1974). Today, however, most investigators conclude that liquid water, in an earlier wetter environment, was responsible for the outflow channels and valley networks.

The presence of the channels evokes three important questions. What was the source of the water that carved the channels? What was the fate of the water? What were the climatic conditions that enabled channel formation? The largest and most widely known channels are those along the western and southern rim of Chryse Planitia (Figs. 1 and 7). These channels are typically on the order of several tens of kilometers wide and up to thousands of kilometers long (Mars Channel Working Group, 1983). All the circum-Chryse channels head in large collapse depressions, or chaotic terrain, implying a subterranean source (e.g., Carr, 1979). Often-cited examples of valley networks also head in incised, theater terminations, suggesting groundwater sapping (Mars Channel Working Group, 1983). However, several other outflow channels and many valley networks appear to head at the surface with no distinct chaotic terrain or theater-headed source region (Parker and Gorsline, 1991a,b; Parker, 1994), requiring spillover of surface lakes or atmospheric precipitation to initiate channel flow.

The question of the fate of the water discharged by the channels is dependent on how much water was discharged during a single flood event (in the case of the outflow channels) and/or what the prevailing climatic and substrate conditions were at the time. In essence, whether fluvial sedimentation into martian basins occurred in a subaerial or lacustrine setting would have depended on the volume of transported material, the sediment/water ratio, the rate and duration of fluvial discharge, the number of simultaneously active channels draining into a given basin, and the stability of liquid water and ice at the martian surface at the latitude of the basin at the time. For the purpose of searching for evidence of paleolakes, therefore, it is reasonable to focus on basins that have inward-draining channel systems that are large or have large catchment areas relative to the volume or area of the basin. Better still, basins with both inward- and outward-draining systems can be sought, as the outward-draining channel would indicate overflow from a former lake within the basin. Several large basins, and hundreds of small to intermediate-size craters, have channels terminating within them. These include most of the largest impact basins on the planet, such as Argyre, Chryse, Hellas, and Isidis. Chryse and Isidis are “embayments” in Mars’ vast northern lowland plains, which comprise approximately one third of the surface of the planet (Fig. 1).

5. Martian paleolakes and oceans

5.1. Origins of martian lacustrine basins

Because the diameter of Mars (about 6800 km) is just over half that of Earth, it does not exhibit global tectonism on a scale comparable to Earth and Venus. But because it is still a large body compared to Mercury and the moon, it has had an atmosphere and climate over solar system history. This is why Mars has been able to retain surfaces produced both through volcanic and climatic processes that are
largely intermediate in age between volcanic surfaces on the moon and Mercury and both types of surfaces on Venus and Earth. For the purposes of this discussion, this has important implications about the origins and evolution of topographic depressions that potentially may have contained lakes.

Tectonism is probably the most important process on Earth for producing closed depressions on the continents, and is clearly responsible for maintenance of the ocean basins through geologic time. In addition, tectonism is probably the most important process for forming depressions in the highland terrains and lowland plains of Venus. On Mars, however, tectonism appears limited to relatively small amounts of regional extension, compression, and vertical motion largely because of crustal loading of the two major volcanic provinces—Tharsis and Elysium (Banerdt et al., 1982; Plescia, 1991; Tanaka et al., 1991; Watters, 1993). The global crustal dichotomy separating the heavily cratered southern highlands from the sparsely cratered northern lowland plains is an ancient feature, resulting from tectonism or the formation of a single, or several overlapping giant impact basins in the northern plains (e.g., McGill and Squyres, 1991; McGill and Dimitriou, 1990; Breuer et al., 1993).

Impact craters and large impact basins (including all or parts of the northern plains) are clearly more important sites for potential lake basins on Mars, though they were likely more important on Earth and Venus as well during the period of heavy meteorite bombardment prior to 3.5 Ga. Comparisons of the relative importance of other formative processes on Mars with those on Earth are less obvious, and some may be quite speculative, because our understanding of the early martian environment is still rather limited. Nevertheless, a preliminary assessment of the origins of lacustrine (and oceanic) basins on Mars, based on those of terrestrial lakes by Hutchinson (1957) and Currey’s (1994a,b) discussion of hemiarid lake basins, is listed in Table 1.

5.2. The martian water budget, and the availability of surface water through time

It is perhaps fair to state, even considering our inexperience with other, truly Earth-like planets, that the geomorphic uniqueness of Earth can be attributed to its sheer abundance of surface water. Only a small percentage of the $1.37 \times 10^9$ km$^3$ of water in the oceans can be stored in continental and polar reservoirs at any given time; the nonmarine reservoirs contain $\sim 2.3 \times 10^8$ km$^3$ of water (Hutchinson, 1957).

On the other hand, the volatile reservoirs on Mars, the megaregolith (the impact-brecciated upper few kilometers of the crust), the polar caps, and potential near-surface massive ice deposits (e.g., Costard and Kargel, 1995) that may be protected from sublimating by thin insulating mantles, are of sufficient volume to contain its entire remaining water inventory. If we assume that the pore volume of $\sim 10^8$ km$^3$ for the megaregolith estimated by Clifford (1981) is reasonable and that very little of the original water has been lost to space, and we allocate his 600 m global average layer to the northern plains, this yields an average ocean depth of over 1.5 km.

The actual value of the total martian water inventory is still poorly known, but is likely to be (have been) between one and two orders of magnitude smaller than the inventory on Earth. Recently, Donahue (1995) has suggested that deuterium/hydrogen ratios from SNC meteorites, thought to have originated on Mars, imply a modern water inventory equivalent to a global layer several meters deep. Extrapolating to the early Hesperian (see Fig. 2 for Martian time scale), Donahue believes this inventory would have amounted to several hundred meters spread globally. How much water could have been present at the martian surface during Noachian and Hesperian time depends on the rate at which juvenile or recycled water was produced through volcanic outgassing and heavy bombardment relative to the rate at which it would be lost to the megaregolith, the cryosphere, and to space. Both processes were likely important prior to 3.5 Ga, but volcanism would have been the only significant contributor after that time. The permeability of the megaregolith is also poorly known (MacKinnon and Tanaka, 1989), but it likely decreased as the meteorite impact flux declined over geologic time, allowing chemical weathering and cementation of surface rocks to occur in the presence of liquid water (or ice) and giving an opportunity to form permafrost seams. Percolation would have been further retarded in a...
Table 1

<table>
<thead>
<tr>
<th>Proposed origins of martian lacustrine and oceanic basins</th>
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<tr>
<td><strong>Structural/geomorphic cause of containment</strong></td>
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<td><strong>Cosmogenic basins</strong></td>
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<td>Large impact basins</td>
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<td><strong>External (crater-dammed valleys)</strong></td>
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<td><strong>Tectonic basins</strong></td>
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<td>Regional downwarp</td>
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<td>Graben</td>
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<td>Fault/wrinkle-ridge dammed valleys</td>
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<td><strong>Fluvial</strong></td>
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<td>Natural levee-dammed</td>
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<td>Alluvial fan-dammed</td>
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<td>Delta-dammed</td>
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<tr>
<td>Abandoned channels</td>
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<tr>
<td>Flood-scoured pools</td>
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<td><strong>Eolian</strong></td>
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<td>Deflation/blowouts</td>
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<td>Glacier-impounded</td>
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<td><strong>Thermokarst</strong></td>
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<td>Permafrost thaw</td>
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<td>Kettle</td>
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Cold Martian paleoclimate because outgassed water would simply snow out onto the surface and remain there until removal through basal melting or sublimation.

5.3. Recognition of coastal landforms on Mars

Coastal morphology is probably the most diagnostic evidence of former lakes that we can expect to find on Mars, at least until unequivocal surface compositions can be identified. Free liquid water is confined by gravity to an equipotential upper surface that intersects topography at an essentially constant elevation over large regions. This makes it possible to distinguish shorelines from fault scarps where both feature types occur, such as in the Great Basin of the southwestern United States. Fault scarps cross...
undulations in topography with little or no lateral deflection, whereas even minor topographic undulations profoundly affect the lateral position of a shoreline. Shoreline elevation varies somewhat locally, because of tides (in very large basins), wind stresses (seiches), and a number of other, though less important, factors. Martian tides would have been much smaller than terrestrial tides because of the lack of a large natural satellite and its greater distance from the sun. Winds blowing across open water generate ripples and waves that transfer their energy to the land, resulting in erosion and sediment transport and deposition that is focused within a few meters of the water level. Longshore transport moves sediment eroded from headlands into embayments, producing beaches that transit from erosional to depositional, with wave-cut cliffs grading along shore into constructional barriers and spits (Fig. 3). Spits and coastal barriers remain as large, often arcuate or cuspatel sets of beach ridges after recession of the lake. Predictable longshore trends of geomorphic development distinguish wave-cut and wave-built
terrace from structure-controlled benches in sequences of layered rocks.

Shore platforms and sedimentary bedforms associated with dissipative shorelines (e.g., Wright et al., 1979) can be extensive, so many terrestrial paleolakeshores can be seen in moderate- to high-resolution orbiting spacecraft image data. Steeply shelving shore zones dominated by wave reflection are much narrower in plan view, so that many steep mountain-front paleolakeshores may be undetectable in orbiter images. Narrow shore platforms require high spatial resolutions to be recognizable, typically moderate- to high-resolution aerial photographs.

How would surface waves on Mars differ from those on Earth, and would they tend to increase or decrease the chances of recognizing coastal landforms from orbiter images of Mars? Beginning with scaling for martian gravity, we have identified some interesting differences between water waves on the two planets (Figs. 4–6). To keep the comparison as simple and straightforward as possible, we restricted this exercise to linear or Airy waves in deep water (e.g., Komar, 1976) and assumed a denser early martian atmosphere and above-freezing temperatures, at least seasonally, to allow a free air–water surface and enable wind-driven waves. Coastal ero-
The phase velocity of an Airy wave in deep water is expressed by:

\[ C_w = \frac{gT}{2\pi} \]  

in which \( C_w \) is the phase velocity in deep water, \( g \) is the gravitational acceleration (m/s\(^2\)), and \( T \) is the period (s), from one wave crest to the next. The ratio of martian to terrestrial wave phase velocity is about one-third (Fig. 4). This is accommodated by a similar relationship between martian and terrestrial wavelengths, defined by:

\[ L_w = \frac{gT^2}{2\pi} \]  

such that martian waves must travel about one third as fast or have about one third the wavelength as terrestrial waves with the same period (Fig. 5).

An important factor to consider with regard to the erosive ability of martian waves is the diameter of particle orbits beneath the waves and the depth at which this motion essentially ceases (“wave base”). This, in turn, will determine the slope and width of the shorezone. In deep water the particle orbits be-

![Wave phase velocity for deep water Airy wave](image1)

Fig. 4. Comparison of terrestrial (black line) and Martian (gray line) wave phase velocities (celerity) for Airy waves in deep water.

![Wavelength versus Period for a deep water Airy wave](image2)

Fig. 5. Comparison of terrestrial (black line) and Martian (gray line) wavelengths for Airy waves in deep water.
neath waves are essentially circular, so the horizontal and vertical components of the orbit are equal. The orbit diameter is then determined by:

\[ d = H \exp \left( \frac{4\pi^2 Z_0}{g T^2} \right) \]  

where \( d \) is the particle orbit diameter, \( H \) is the wave height, and \( Z_0 \) is the depth below the still water level to the center of the particle orbit. The results of substituting martian and terrestrial gravity into the equation for the specific case of 1-m high waves of 10-s period are shown in Fig. 6. On Mars, \( d \) decreases much more rapidly with depth than it does on Earth. This suggests that terrestrial waves are much more effective at moving sediment at depth than “equivalent” martian waves. At the same time, it also indicates that martian waves entering shallow water “feel bottom” later than terrestrial waves and that martian wave energy is focused within a narrower depth range than terrestrial wave energy. On Mars, this should lead to the production of dissipative shorezones that are broader and have gentler slopes than those on Earth. This might permit the development of coastal landforms on Mars that are large enough to be visible from orbit, while reducing the “Earth-like” climate requirements somewhat.

An additional or complementary mechanism for potentially producing broad paleoshore platforms on Mars may be found in the strandflats of Earth. On Earth, and particularly along numerous segments of the North Atlantic coast, strandflats occur within a few meters or tens of meters of present (eustatically and isostatically evolved) sea level and range in width (normal to the coast) from less than 1 km to at least 60 km. Martian analogs of the strandflats would almost certainly be detectable as broad, abandoned shore platforms from orbit. Although the origin (or origins) of strandflats on Earth is still a matter of debate (e.g., Trenhaile, 1987), the most likely formative processes—(1) coastal zone frost weathering coupled with debris removal by wave action; (2) rock scour by moving shelf ice, both glacial and non-glacial (the latter exemplified by Ward Hunt Ice Shelf on Ellesmere Island and ice islands in the Arctic Ocean); (3) rock plucking and removal by shore-fast sea ice; and/or (4) rock scour by drifting sea ice or circulating pack ice—are not inconsistent with plausible martian paleoclimatic conditions.

5.4. Martian paleolake studies

Martian paleolake studies have taken two approaches. One group of investigators has focused on the heavily cratered, relatively ancient southern highlands as the most likely place to preserve evidence of a former wet climate. The valley networks, while they may not always indicate atmospheric precipitation, nevertheless require significant discharges for sustained periods of time. The other group of investigators has focused on the basin-forming northern lowlands as a likely site for paleolakes or even a paleo-ocean. The primary sources of influx into this

![Fig. 6. Comparison of terrestrial (black line) and Martian (gray line) particle orbit diameters beneath a 1-m high Airy wave of period \( T = 10 \) s.](image-url)
basin are the circum-Chryse outflow channels and several smaller outflow channels and valley networks west and south of the Elysium volcanic complex (Fig. 7). The Chryse channels alone are large enough (Carr, 1986; Rotto and Tanaka, 1991; DeHon and Pani, 1992; DeHon, 1993) to have produced several large lakes, even a sea or ocean, within the northern plains in a time period as short as a few months or years.

In the southern highlands, several basins on the order of several tens to a few hundred kilometers across and parts of Valles Marineris have been found to contain remnants of former alluvial or lacustrine sedimentary deposits. Many of these basins were simply in the paths of short-term catastrophic floods, and thus may not have contained lakes for very long (DeHon, 1992). The thick Valles Marineris layered deposits occupy deep graben and collapse depres-

Fig. 7. Polar projection of northern plains of Mars relative to major bounding provinces. Moderate gray indicates southern cratered uplands and volcanic terrains. Pale gray stipple indicates northern plains. White indicates residual north polar ice cap. The dichotomy between the northern plains and southern highlands is indicated as either “gradational” (Rossbacher, 1985) or “fretted” (Sharp, 1973), except where regional volcanic surfaces appear to obscure it. Major outflow channels are indicated by light gray stipple. Mappable distributions of two major northern plains boundaries, which may be ancient shorelines, to which the terms “gradational boundary” and “interior plains boundary” were applied in Parker et al. (1993), are indicated by thick, dark gray line (coincident with lowland/upland boundary for the most part) and thick, black line, respectively. Map adapted from Parker et al. (1989).
sions in the martian crust, and appear to have been
fed by groundwater rather than overland flow (Nedell
et al., 1987; Komatsu et al., 1993). The deposits are
on the order of kilometers thick and appear finely
laminated (in high resolution Viking Orbiter images),
and so probably indicate the presence of lakes for
long periods of time. These layered deposits and two
notable examples of highland basins have been the
subjects of proposals for future lander missions to
Mars that might search for evidence of fossil organic
materials summarized in Landheim et al., 1993.
The first of these, the 135 km crater Gusev (15°S
lat., 184° lon., Fig. 8), lies at the mouth of the large,
tributary-fed channel Ma’adim Vallis and contains

Fig. 8. Terminal deposit at mouth of Ma’adim Vallis, in the 135-km-diameter impact crater, Gusev (15°S latitude, 184° longitude). This
feature, expressed as partially eroded, flat-topped mesas along the south-central margin of the impact basin, may be the remnants of an
alluvial fan or delta at the mouth of the channel. Portions of Mars Digital Image Mosaics MI15S182, MI15S187.
the remnants of deltaic or alluvial fan deposits and possible lake sediments (Schneebberger, 1989). The other site is an unnamed, 200 km diameter depression along the Parana–Loire valley system in Margaritifer Sinus (23°S lat., 13° lon., Fig. 9). This basin contains a peculiar, hummocky deposit that has been interpreted as the eroded remnants of lake sediments by Goldspiel and Squyres (1991). Similar deposits can be seen in other large, shallow basins in the Phaethontis region of Mars at 35°S latitude, 177° longitude (Atlantis Chaos) and nearby at 30°S latitude, 170° longitude (Gorgonum Chaos, Fig. 10). On Earth, eroded lake deposits are among the more common host terranes for badlands topography, which bears some resemblance to the martian deposits, though fluvial dissection is more pronounced.

Fig. 9. “Etched” deposit in an intercrater depression in Margaritifer Sinus (23°S lat., 13° lon.). This depression lies at the termini of several valley networks (Parana Valles, eastern part of image mosaic), and is the source for Loire Vallis (large valley, northwest part of image mosaic). Such depressions, with channels that drained into them and others that flowed out from them, are not uncommon in the martian highlands, and are perhaps the most convincing evidence of paleolakes in the ancient cratered terrain of Mars. Viking Orbiter images 615a44-48,65-68.
Fig. 10. Three large, degraded basins, which are interpreted to be old impact structures, contain “etched” deposits similar to those found in Gusev and Margaritifer Sinus, but are more extensive. Though several valley networks terminate within these depressions, none appear to flow from them. Basin deposits, Atlantis Chaos (35°S lat., 177° lon.), are located in the northwest part of the mosaic. Basin deposits, Gorgonum Chaos (38°S, 170° lon.), occupy the basin located in the southwest part of mosaic. Portions of Mars Digital Image Mosaics MI30S167, MI30S172, MI30S177, MI45S167, MI45S172, MI45S177.

The largest highland basins, Argyre (900 km interior diameter) and Hellas (1500 km interior diameter) appear to contain massive accumulations of layered sediments that are now being exposed through eolian deflation. Those in Argyre may be on the order of hundreds of meters or more thick (Parker and Gorsline, 1993), whereas those in Hellas appear to have been as much as a few kilometers thick (Moore and Edgett, 1993). Two outflow channels, Dao Vallis and Harmakhis/Reull Valles, terminate within the eastern rim of Hellas Basin with no outlet. Three large valley networks, Surius Vallis, Dzigai Vallis, and Palacopas Vallis, which head near paleopolar deposits at high southern latitudes, drained into Argyre basin and Uzboi Vallis, an outflow channel, appears to have spilled northward from a lake within the basin (e.g., Parker, 1989, 1994; Parker and Gorsline, 1991a,b, 1993).

The origin and evolution of the northern plains of Mars, once largely ignored in favor of focused studies of highland and volcanic landforms that are often much more dramatic in appearance, has recently become a subject of high scientific interest. The reasons for this relate largely to the need to under-
stand the evolution of the water inventory on the planet, particularly its surface water inventory, through geologic time. This, in turn, has important implications regarding the question of exobiology on Mars, particularly if long periods of geologic time with available surface water are indicated.

Over the past several years, a number of investigators have described geomorphic evidence, which indicates that large standing bodies of water or ice sheets once occupied the northern lowland plains of Mars (including paleoclimatic significance; e.g., Jöns, 1985, 1986, 1990; Lucchitta et al., 1986; McGill, 1985, 1986, McGill and Hills, 1992; Parker et al., 1987a,b, 1989, 1993; Baker et al., 1991; Scott et al., 1991a,b; Rotto and Tanaka, 1991; Chapman, 1994; Kargel et al., 1995). The details of the timing, emplacement mechanisms, and sizes of these water bodies differ markedly, however, from one group of investigators to another. For example, Jöns (1985, 1986) envisioned a “mud ocean” covering much of the northern plains, with sediment slurries derived from a variety of peripheral sources, including the fretted terrains and outflow channels. Lucchitta et al. (1986) pictured an ice-covered ocean, fed by large circum-Chryse ice streams, analogous to those in Antarctica. Parker et al. (1989, 1993) indicated two or more highstands of a sea or ocean that, during the early Amazonian, would have been charged by catastrophic floods, but may have existed more or less permanently during Noachian and Hesperian time (Fig. 2). Baker et al. (1991) pictured a plains-wide ocean emplaced by the major outflow channels relatively late in martian history during the early and middle Amazonian, and coined the term “Oceanus Borealis” for this ocean. Similarly, the ice-sheets described by Kargel et al. (1995) and Chapman (1994) are attributed to the early Amazonian. Interestingly, the shorelines of Jöns’ mud ocean, Lucchitta et al.’s ice-covered ocean, and Parker et al.’s (1993) most recent sea, or “interior plains”, coincide almost precisely around the northern lowlands, though the details of the mechanisms by which boundary morphologies are thought to have been produced differ.

Taking a more conservative approach to the question of standing water in the northern plains, Rotto and Tanaka (1991) relied on volume estimates of maximum discharge from the circum-Chryse outflow channels, which they feel limits any standing water to one or a few large, ephemeral lakes. They based the locations of these lakes on the identification of broad, shallow topographic basins on the present Martian topographic maps (U.S. Geological Survey, 1989). Similarly, Scott et al. (1991a) have indicated evidence for several large lakes across the northern plains, some exhibiting connecting spillways, that were fed by a variety of channel sources peripheral to the plains. Delineation of these lakes was based on a similar assessment of the topography but also included the local identification of shore morphology.

All the above studies are now being evaluated in light of recent acquisition of high-resolution topography afforded by the reflight of the Mars Observer Laser Altimeter (Zuber et al., 1992) in late 1996. For example, estimates of basin volume in Parker et al. (1993) are loosely based on the available topography with its very large vertical errors. These estimates, when compared to estimates by others of the water discharged by the Chryse outflow channels, suggest the possibility that the volumes required to fill the basin may be at or beyond the high end of the estimated volumes available from the channels. High resolution topography is needed to sort out the common modifiers of shoreline elevation, such as tectonism and isostatic rebound (with lower amplitudes than on Earth, perhaps) and sediment desiccation and compaction; these modifiers likely altered the topography of the northern plains after the putative surface water was lost, so that the original topography can be reconstructed.

Local elevations can be derived using the currently available high-resolution topographic tools. The precision of topographic measurements taken from stereo image pairs can be maximized by using high-resolution stereo pairs and medium-resolution pairs with large parallax angles. Photoclinometry provides the highest topographic measurement precision currently available, but contains a number of potential sources of errors (e.g., see discussion by Jankowski and Squyres, 1991). Photoclinometric profiles should be corroborated, whenever possible, with other measurements. The vertical precision afforded by photogrammetry is less than that provided by photoclinometry. Photogrammetrically derived elevations, however, can be correlated regionally...
within a stereo scene, whereas photoclinometric elevations cannot. The last, and simplest, topography tool, shadow measurements, is most suited to steep slopes with sharp breaks in slope and low sun elevations. High-resolution images with low sun elevations are relatively uncommon, particularly from late in the Viking mission, but even medium- to low-resolution images can be quite suitable with a low sun and modest relief.

Until the current Mars Global Surveyor mission, elevations derived using the available high-resolution topographic tools could not be accurately tied to the global datum (e.g., Zuber et al., 1992). Nevertheless, these tools are very useful for measuring the relief of local features that are distributed regionally. For example, massifs within the northern lowlands commonly exhibit a surrounding apron with a horizontal upper surface that may indicate the remnants of a

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Fig. 11. Portion of Viking Orbiter image 070a02, showing massif with relatively steep slopes immediately above a surrounding flat-topped apron. Asymmetric photoclinometric profiles (bottom) are indicated on left image. These profiles, based on a “flat field” brightness value, approximate the height and overall shape of the hill and surrounding apron. The apparent difference in height of the apron to either side is due to changes in albedo of the surface from one side to the other. The enlargement of the hill (at right) exhibits a faint dark layer (arrowed) below the top of the escarpment on its southwestern and northwestern (sunlit) sides. On the northwestern side, the layer dips toward the plains, paralleling the hill’s topographic surface, suggesting that the hill has been tilted toward the northeast. The horizontal surface of the apron surrounding the massif is cut at an angle to this layer. North is toward top of scene.
formerly continuous surface above the modern plains (Parker et al., 1987b, 1993). These massifs appear similar to a wave-eroded headland massif in Lake Bonneville, depicted in Fig. 3, and may indicate coastal erosion of inselbergs within the plains.

If a coastal interpretation for these features is correct, photoclinometric and photogrammetric measurements of the heights of the flat upper surfaces of the aprons can be used to provide local estimates of the depth of the paleolake or paleo-ocean. We present examples of measurements of this type for the eastern Acidalia/Cydonia Mensae region, for which uniform high-resolution coverage exists from the Viking Orbiters over a region of several tens of

![Figure 12](image-url)
Fig. 13. To insure that the measurements derived from the photoclinometric profiles are as accurate as possible, it is advisable to cross check those measurements with other techniques. These checks can include shadow measurements of the massifs (when possible), and stereo parallax measurements (severely limited due to small number of stereo pairs at high-resolution). The most reliable check of asymmetric photoclinometric profiles are measurements using a symmetric profile. This pair of profiles compares the height determined via an asymmetric profile (left) and a symmetric profile (right). Both profiles returned a height of approximately 130 m. North is toward top of scene.
thousands of square kilometers, and from northern Acheron Fossae, where high-resolution stereo images are available. Preliminary photoclinometric measurements place the upper surfaces of the massif aprons in northern Cydonia Mensae between 150 and 200 m above the surrounding plains (Figs. 11–13). Additional benches are also evident at the margins of many of the massif aprons (Fig. 12). These appear to lie between 20 and 40 m below the upper surface of the apron around the associated massif.

In northern Acheron Fossae, stereo photogrammetry was used to measure the height of a horizontal bench in the rim of a large, degraded crater (Fig. 14). The crater lies on the north-sloping margin of Acheron Fossae, and is partially buried by northern plains materials on its lower, northwest side (Fig.

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Fig. 14. 35-km-diameter degraded crater at northern edge of Acheron Fossae. Northwest part of crater rim is partially buried by plains materials. Much of the rest of the crater appears subdued, except for a sharply defined portion of the east rim. This area is covered in stereo in the following figure. Viking Orbiter image 035b17.
Fig. 15. Stereo pair of east rim of crater in Fig. 14 with stereo parallax measurement points plotted. The prominent bench at east rim appears to have been cut horizontally into the northward tilted rim of the crater. “Softened” crater rim is below this bench, sharply defined rim is above. Profiles of the crater floor (points labeled (p) for plains), the crater rim (points labeled (c)), and the bench (points labeled (b)) and a plateau of “perched” chaotic terrain material, possibly a remnant of a former deposit within the crater (points labeled (ch)) for Fig. 16 are indicated. Portions of Viking Orbiter images 112a32 (left) and 130a21 (right).

The bench lies between 140 and 180 m above the plains surface (Fig. 16).

A lacustrine/marine origin of these features would have far-reaching implications with regard to martian
surface volatile history and exobiology studies, particularly if long periods with surface water are indicated.

6. Conclusions

We have examined the current planetary datasets for evidence of extraterrestrial coastal landforms. The search was limited to solid surface planetary bodies that have, or may once have had, atmospheres. This includes Venus, Mars, the moon of Saturn, Titan, the moon of Neptune, Triton, and Pluto.

In the outer solar system, compounds other than water would be necessary to collect into standing bodies at the surface, because of the very low temperatures far from the sun. Liquid nitrogen, which was once thought to be stable at the surface of Triton, is instead stable only in solid or gaseous form. Pluto is comparable in size and bulk composition to Triton, and is at a similar distance from the sun, so environmental conditions are thought to be similar. Titan, however, is now known to have a massive atmosphere. Several investigators are proposing that Titan may have lakes or oceans of liquid hydrocarbons. The Cassini Mission to Saturn, due to reach the planet early next century, will image portions of Titan with a Synthetic Aperture Radar on board the Saturn Orbiter, and will deploy the Huygens probe into the atmosphere of Titan. This probe includes a descent imager that will operate from when the parachute opens to the surface. The potential verification of abundant stable liquids at the surface of Titan, either in the past or at present, is an exciting prospect for future planetary geomorphic studies.

Considering the planetary neighbors in the inner solar system, we have seen that Venus and Mars may have had abundant liquid water on the surfaces at some time in the past. The Magellan Mission has revealed Venus to have experienced an episode of global resurfacing, such that surfaces older than several hundred million years have not been preserved. Since a “clement” period on Venus, if it ever had one, is thought to date back billions of years, further exploratory missions, such as aerial reconnaissance or surface rovers capable of identifying ancient sedimentary deposits, will be required.

Mars is much less geologically active than Venus, so very ancient landforms that point to a warmer and/or wetter paleoclimate are preserved to this day. Several investigators have suggested that Mars may have had lakes, ice sheets, and even oceans prior to about 2.5 Ga, and perhaps until as recently as several hundred million years ago.

In our review of the evidence for lakes or oceans on Mars, we conclude the following: Mars lacks large, stable 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 declined over time until it could no longer keep pace with loss to subsurface reservoirs through percolation, chemical weathering, permafrost/polar cap formation, and photodissociation and loss to space; (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 lack of plate tectonism means that Mars has no mechanism to “rejuvenate” large topographic depressions, so basins are likely to have become broader and shallower with each lake occupation, eventually converting the lake basins to playa basins.

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