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# Marine-target craters on Mars? An assessment study

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**Abstract–**Observations of impact craters on Earth show that a water column at the target strongly influences lithology and morphology of the resultant crater. The degree of influence varies with the target water depth and impactor diameter. Morphological features detectable in satellite imagery include a concentric shape with an inner crater inset within a shallower outer crater, which is cut by gullies excavated by the resurge of water. In this study, we show that if oceans, large seas, and lakes existed on Mars for periods of time, marine-target craters must have formed. We make an assessment of the minimum and maximum amounts of such craters based on published data on water depths, extent, and duration of putative oceans within "contacts 1 and 2," cratering rate during the different oceanic phases, and computer modeling of minimum impactor diameters required to form long-lasting craters in the seafloor of the oceans. We also discuss the influence of erosion and sedimentation on the preservation and exposure of the craters. For an ocean within the smaller "contact 2" with a duration of 100,000 yr and the low present crater formation rate, only ~1–2 detectable marine-target craters would have formed. In a maximum estimate with a duration of 0.8 Gyr, as many as 1400 craters may have formed. An ocean within the larger "contact 1-Meridiani," with a duration of 100,000 yr, would not have received any seafloor craters despite the higher crater formation rate estimated before 3.5 Gyr. On the other hand, with a maximum duration of 0.8 Gyr, about 160 seafloor craters may have formed. However, terrestrial examples show that most marine-target craters may be covered by thick sediments. Ground penetrating radar surveys planned for the ESA Mars Express and NASA 2005 missions may reveal buried craters, though it is uncertain if the resolution will allow the detection of diagnostic features of marine-target craters. The implications regarding the discovery of marine-target craters on Mars is not without significance, as such discoveries would help address the ongoing debate of whether large water bodies occupied the northern plains of Mars and would help constrain future paleoclimatic reconstructions.

## INTRODUCTION

The number of known impact craters on Earth that can be related to an aqueous target environment has increased significantly in recent years, in step with increasing knowledge about their typical associated features and geology (Ormö and Lindström 2000; Tsikalas et al. 1999). Several of these craters have proved useful in estimating the water depth at the time of impact (Lindström et al. 1996a; Ormö et al. 2002). All the marine-target craters found so far were formed in epicontinental seas with shallow to deep shelf water depths. Therefore, the target environment commonly consisted of both water and sediments of differing degrees of consolidation overlying crystalline basement materials. Mars is another planetary body in our solar system where oceans (Parker et al. 1987, 1993; Baker et al. 1991; Clifford and Parker 2001; Fairén et al. 2003) and lesser water bodies such as seas and lakes (Scott et al. 1995) may have persisted over long periods of time (Table 1; discussed in the section Duration and Extent of Putative Water Bodies in the Northern Lowlands of Mars). Since at least the Late Noachian, the martian surface erosion rate has been significantly lower than that of Earth (Scott et al. 1995). Furthermore, subduction related to plate tectonics may have occurred on Mars (Sleep 1994) only during its early phase of evolution (Baker et al. 2002a, b; Dohm et al. 2002; Fairén et al. 2002; Fairén and

 

Table 1. Elevation, area, volume, depth (for present-day topography), and length of the basins analyzed in this study (based on Öner T. and Ruiz J., personal communication). Note that the "contact 1-Meridiani" shoreline is composed from two different proposed shorelines (Parker et al. 1989, 1993; Edgett and Parker 1997), as described in Fairén et al. (2003) and shown in Fig. 1.

| Shoreline enclosing basin | Mean elevation (m) | Basin area ($10^7$ km$^2$) | Basin volume ($10^7$ km$^3$) | Mean depth (km) | Maximum depth[a] (km) | Global equivalent layer (km) | Length (Myr) |
|---|---|---|---|---|---|---|---|
| Contact 1: Meridiani shoreline | –1500[b] | 5.35 | 10.54[c]–10.66[d] | 1.99 | 3.75 | 0.74 | 0.1–800 (likely ~100) |
| Contact 2 | –3792[e] | 2343 | 1.31[c]–1.43[d] | 0.54 | 1.46 | 0.09 | 0.1–800 (likely ~100) |

[a]In the North Polar Basin (minimum topography, –5250 m).
[b]Parker et al. (2000).
[c]North Polar Cap not included.
[d]North Polar Cap included.
[e]Carr and Head (2003).

Dohm 2003). These circumstances should favor the discovery of marine-target impact craters on Mars. The existence of such craters would give further credence to the proposed lacustrine and/or marine phases in the history of the planet and constitute a basis for future studies on water depth and areal coverage of paleolakes and oceans.

The objectives of this study are to: 1) evaluate the possibility for the existence of marine-target impact craters on Mars; and 2) present search criteria for future investigations, including techniques that will be used in future Mars missions to detect buried geological structures. For this assessment study, we combine the present knowledge of marine impact cratering on Earth, published information on the duration and extent of putative martian paleo-oceans, and the size and frequency of impactors large enough to have generated detectable seafloor craters in these oceans. Due to the uncertainties in duration of the paleo-oceans, here we will adopt a simple quantitative approach for the assessment of the number of marine-target craters that may have formed in these oceans, using reasonable maximum and minimum published estimates on the durations. We use the MEGAOUTFLO hypothesis (Baker et al. 1991, 2000) for the estimated minimum estimated duration, while the maximum durations are based on Fairén et al. (2003) (Table 1). We also use numerical modeling to find the smallest impactor capable of cratering the seafloor of these oceans, assuming a different average water depth for each basin. The approximate crater formation rates for each resulting impactor diameter is known (e.g., Hartmann et al. 1981; Neukum and Hiller 1981). Combined with the extent and duration of the oceans, it is possible to make minimum and maximum estimates on the number of marine-target craters formed. Each step in this calculation requires a number of assumptions. Nevertheless, it is the best possible assessment given the current knowledge of martian paleo-oceans. As it is impossible to achieve an exact result, we choose to make minimum and maximum assessments on the number of possible marine-target craters for each of the two dominant putative oceanic phases (Fairén et al. 2003).

## BACKGROUND

### Putative Water Bodies in the Northern Plains of Mars

Geomorphologic evidence for ancient oceans in the Martian northern lowlands (Parker et al. 1987, 1993; Baker et al. 1991) and/or seas and lakes (Scott et al. 1995) has been proposed, though this evidence has been highly debated by the scientific community (e.g., Malin and Edgett 1999). In addition, lacustrine environments (some perhaps long lasting) have been proposed for large impact crater basins such as Hellas Planitia and Argyre based on a collection of geomorphic evidence (including terraces) and stratigraphic information (Moore and Wilhelms 2001; Kargel and Strom 1992). Similar evidence (including deltas) have been documented in smaller craters indicating that standing bodies of water have existed on the Martian surface (Cabrol and Grin 1999; Cabrol et al. 2001a, c) with durations of at least thousands of years (Ori et al. 2000). However, these lakes are not used in our study.

It has been suggested that the topographic boundary that delineates parts of the northern lowlands may represent the extreme reach of lowland pyroclastic or basaltic flows (Tanaka et al. 1992), but the same boundary has alternatively been interpreted as the partial remnant of an ancient shoreline (Parker et al. 1989). Baker et al. (1991) termed the paleocean within the boundary Oceanus Borealis and presented additional geomorphologic evidence that can be explained by episodic ocean formation. The MEGAOUTFLO hypothesis (Baker et al. 2000) first proposed by Baker et al. (1991) genetically links episodic magmatic-driven activity with floods that pond in the northern lowlands to form oceans and associated relatively short-lived climatic perturbations (duration of tens of thousands of years).

If interpreted as shorelines, the boundaries known as "contact 1" and "contact 2" in the northern lowlands (Table 1 and Fig. 1) would indicate at least two distinct sea levels in the northern lowlands (Parker et al. 1989, 1993). Head et al.

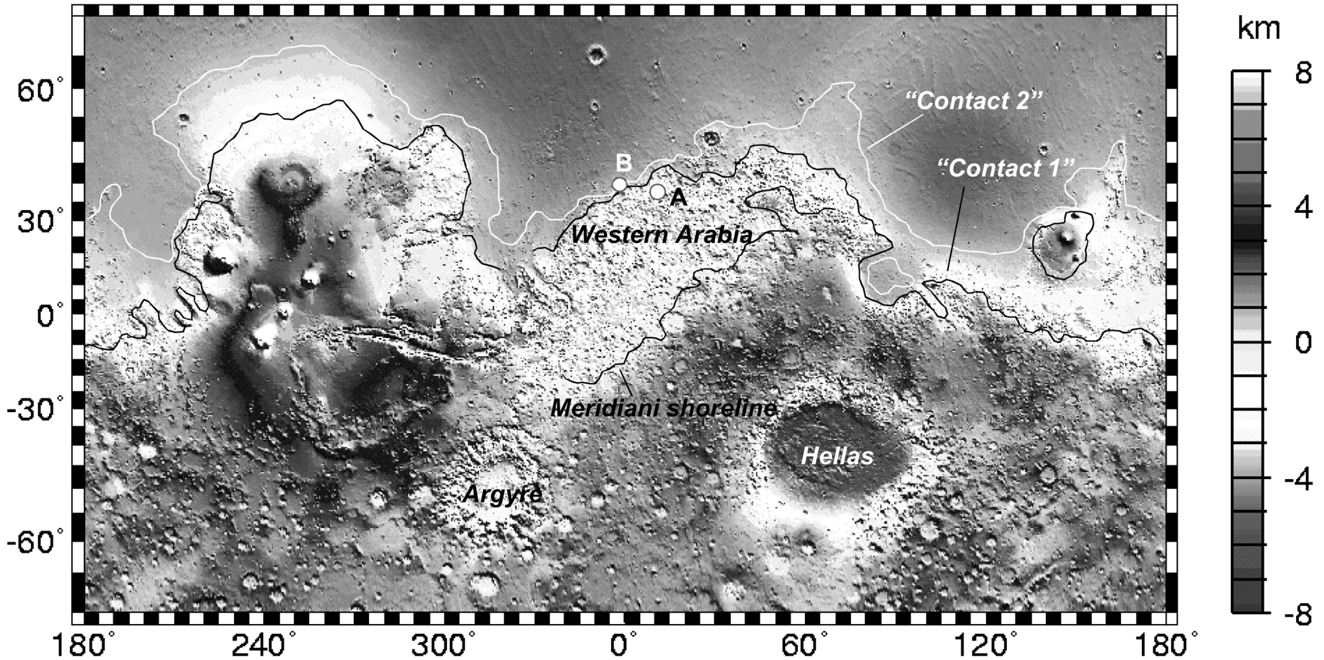


Fig. 1. Global topography of Mars (modified from Smith et al. [1999]). Circles indicate the locations of 2 candidate marine-target craters (A is the location of the crater in Fig. 5a, and B is the crater in Fig. 5b) proposed by Ormö and Muinonen (2000). The white and the black lines show proposed shorelines (contacts 1 and 2; Parker et al. 1989, 1993). The "Meridiani shoreline" indicates the shoreline proposed by Edgett and Parker (1997), which is included in the contact 1-Meridiani ocean used in this study.

(1998) estimate that contact 1 formed during the Hesperian epoch and that contact 2 is younger than contact 1. Volumetric calculations of the lowlands within the two contacts using Mars Orbiter Laser Altimeter (MOLA) data support the northern plains ocean hypothesis, best portrayed by contact 2 (Head et al. 1998, 1999).

During the post-heavy bombardment period in the Late Noachian, even low-lying highland regions may have been covered by water (Edgett and Parker 1997). Western Arabia and Sinus Meridiani are presently below the 2 km USGS datum. These intermediate areas between the highlands and the lowlands are clearly visible in the MOLA topography. The minimal formation of valley networks across this Western Arabia Shelf (Frey and Roark 1997), in contrast to the extensive valley networks in higher areas (Scott et al. 1995), indicates that the region may have been underwater (Edgett and Parker 1997). From a recent synthesis of geologic (stratigraphic and paleotectonic), geomorphic, and topographic information, Fairén et al. (2003) show at least three main stages for water evolution on Mars: 1) first, a great Noachian/Early Hesperian northern plains ocean (Fairén and de Pablo 2002) (related to stages 1–3 of Tharsis development; Dohm et al. 2001a, b), best portrayed by the martian dichotomy boundary (approximately equal to contact 1 and parts of Western Arabia delineated by the Meridiani shoreline in Edgett and Parker [1997]; indicated in Fig. 1); 2) after a transient dry period, a Late Hesperian sea (related to stage 4 of Tharsis [Dohm et al 2001a, b] and Elysium development) that extended over the deeper areas in the lowlands inset within the boundary of the first great ocean, portrayed by contact 2; and 3) a number of widely distributed minor lakes, which may represent a reduced Late Hesperian sea or ponded waters in the deepest reaches of the northern lowlands related to minor Tharsis- (stage 5) (e.g., Anderson et al. 2001) and Elysium- (Skinner and Tanaka 2001) induced Amazonian flooding (Baker et al. 1991; Scott et al. 1995). The hypothesis, which suggests short-lived hydrological cycles, contrasts with the general belief that Mars was warm and wet and later transitioned into persistent cold and dry desert conditions.

Controversy exists concerning whether contacts 1 and 2 (Parker et al. 1989, 1993) are actual shorelines of paleo-oceans (e.g., Head et al. 1998, 1999; Ruiz et al. 2003) and whether standing bodies of water have existed for more than a few hundred years (e.g., Segura et al. 2002; Banfield et al. 2003; Hoefen et al. 2003). Several arguments have been advanced to assess whether water masses ever accumulated on the martian surface, including: insufficient Mars Orbital Camera (MOC)-based evidence to support the existence of coastlines (Malin and Edgett 1999); $CO_2$ as the primary erosional agent to explain the geomorphic characteristics of the martian surface (Hoffman 2001); the rough coincidence between the peak in the valley network formation and the period of heavy bombardment in the Noachian (Segura et al. 2002); the hydrodynamic escape of atmospheric volatiles, including $CO_2$ (e.g., Anders and Owen 1977); and the lack of carbonates observed at the martian surface (e.g., Carr and Head 2003; Bandfield et al. 2003). The MOC controversy on the coastlines has been largely discussed elsewhere (Clifford and Parker 2001; Parker et al. 2001; Fairén et al. 2003); liquid $CO_2$ reservoirs should not persist for tens of thousands and perhaps millions of years, especially during periods of magmatic and tectonic activity or impact cratering events, and it is doubtful they can sustain their erosive capabilities for great lengths of time at martian conditions (Fairén et al. 2003). Segura et al. (2002) favors a cold and dry planet, even during the end of the heavy bombardment (3.5 Gyr), with only very short (<1000 yr) episodic flash floods caused by heat from large impacts. As the peak in the highland valley network formation is roughly coincident with the period of heavy bombardment, Segura et al. (2002) conclude that rains excavating the valley networks were triggered almost exclusively by large body impacts, a process the frequency of which made the persistence of temperate and wet conditions impossible. Although valley network formation may have developed as a result of large impact events, there are many other viable contributors to valley network formation and other Noachian aqueous activity (e.g., Scott et al. 1995; Craddock and Howard 2002; Carr 2002). In addition, up to 40% of the valley networks may be younger than Noachian (Scott and Dohm 1992), and some of them are unambiguously Hesperian or Amazonian (Carr 2002), indicating that fluvial activity has occurred throughout the history of Mars, adding further credence to the ocean hypothesis.

Carbonates have not been observed in great percentages spectrally in outcrops at the surface at current TES- and THEMIS-resolution. However, there may be explanations for this other than their absence, such as sparse spectral data of sufficient resolution to see in detail the diversity of rock materials that may exist on the surface of Mars; depletion of carbonates in the uppermost martian crust due to secondary chemical alteration (Huguenin 1974; Settle 1979; Craddock and Howard 2002); hiding under younger materials (Banin et al. 1997; Clark 1999; Blaney 1999); or, maybe the putative transient ocean environments were simply not adequate for the formation of carbonates.

Though results from the NASA MER and ESA Beagle Landings, as well as future Science-driven missions, undoubtedly will shed further light on this long-standing controversy, here, we assume that water bodies did exist in the northern plains of Mars in the past.

## The Formation and Preservation of Marine-Target Craters on Earth

To discuss the criteria for finding marine-target craters on Mars, it is first necessary to review the geologic features of terrestrial marine-target craters. Ormö and Lindström (2000) listed the most prominent geologic features that can be expected to have resulted from impact cratering events at specific water depths. They base their conclusions on 14 well-documented terrestrial marine-target impact craters, as well as published laboratory and modeling results. Continued studies (Ormö et al. 2002; Lindström et al. Forthcoming) have provided additional information on impact crater formation into water bodies, including associated landforms and morphologic characteristics.

Because water is a Newtonian fluid where the rate of sheer stress is directly proportional to the rate of deformation, it can be considered a mechanically-resistant target (e.g., Melosh 1989; Gault and Sonett 1982). As such, the effect of an impact on the seabed is strongly dependent on the depth of the water column. The laboratory experiments by Gault and Sonett (1982) show that a reduction of the water-depth/expended energy ratio produces a larger diameter of the transient cavity formed in the water mass but not necessarily crater excavation in the seafloor. They referred to this as a "protective effect of water" from cratering of the seafloor. Therefore, even impacts that, according to the standard transient depth/diameter ratio (approximately 1/3; Melosh 1989), should have a transient cavity deeper than the thickness of the water layer may only stir the bottom sediments. In these cases, the cavity has a cylindrical shape (see the section Duration and Extent of Putative Water Bodies in the Northern Lowlands of Mars). A shallow excavation flow develops along the interface between the more competent basement materials and the viscous water column. This is seen both in field geology and numerical simulation (Ormö et al. 2002). For certain relations between impactor diameter and target water depth, the shallow excavation flow generates a concentric shape of the transient crater (terminology adopted from Quaide and Oberbeck 1968). This has been reproduced in numerical simulations (e.g., Roddy et al. 1987; Shuvalov and Artemieva 2001; Artemieva and Shuvalov 2002; Ormö et al. 2002). The shallow excavation flow along the seafloor can generate a permanent concentricity of the crater by stripping not only the water from the surroundings of the nested, central crater depression but the weaker layers of the seafloor as well (e.g., the Lockne crater, Sweden; Lindström et al. 1996b; Ormö and Lindström 2000). The degree of concentricity of the crater depends on relative water depth (Fig. 2). The maximum outer/inner crater diameter ratio (approximately 2.5) occurs for water depths between one and three times greater than the impactor diameter (Shuvalov and Artemieva 2001; Ormö et al. 2002). For very shallow relative water depths, the crater develops an elevated rim similar to a land-target crater (Fig. 2b). In this case, the expansion of the water cavity follows the expanding seafloor crater, and no concentric shape will develop. Furthermore, the rim of the seafloor crater is so high that it prevents the resurge of expelled water from entering the crater. The ejected water can cause instability of the rim, which may result in increased slumping (Fig. 2b). Some marine-target craters may develop a concentric shape (e.g.,

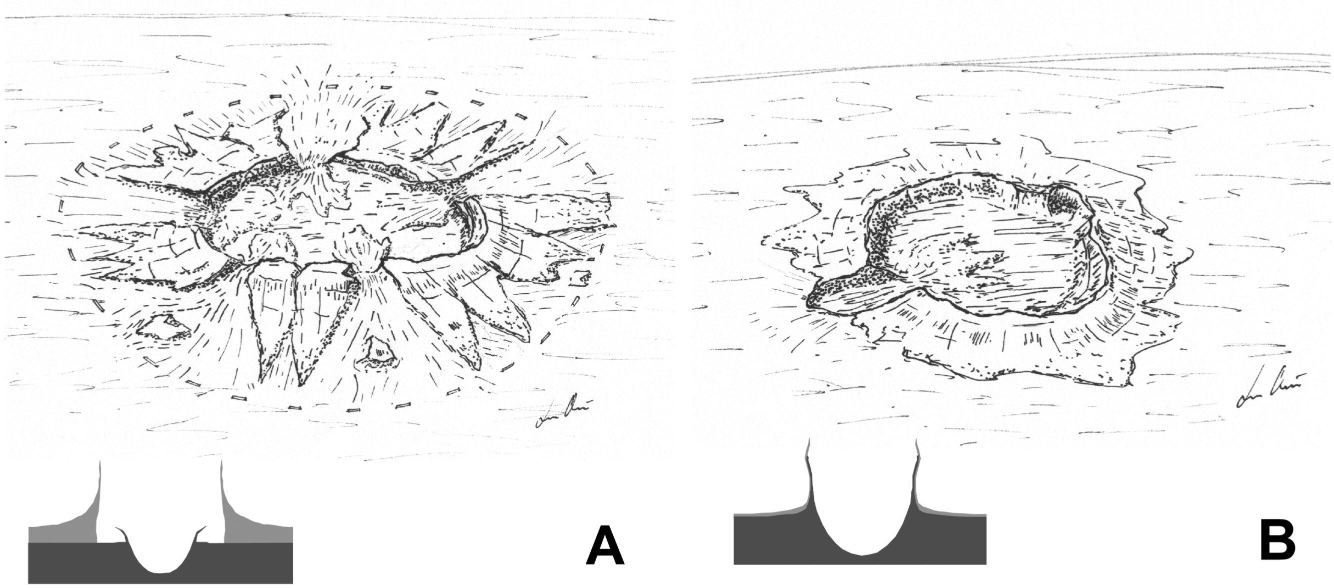


Fig. 2. Schematic illustrations of marine-target craters formed at different relative water depths. The drawings show the seafloor craters after early modification (e.g., the collapse of the water cavity). For simplicity, the water has been removed. Inset profiles show the transient cavity, where water (light shade) and bedrock (dark shade) are shown (modified from Ormö et al. [2002]). The morphology and geology of the crater depends on the water depth in relation to expended energy (roughly the impactor diameter; see Shuvalov and Artemieva 2001). At intermediate target water depth (a), the seafloor crater may develop a concentric shape with a deeper, nested crater surrounded by a less distinct, shallow outer crater (stippled line in drawing). The nested crater has an overturned ejecta flap at the rim (e.g., Lockne crater, Sweden). Resurge has occurred through tears in the flap (Lindström et al. Forthcoming). Sedimentation continues and will eventually cover the crater; b) shallow target water depth. The crater resembles a land-target crater both in shape and geology. Localized rim collapse can cause openings for the surrounding sea to enter, but less forcefully than in (a) (e.g., the Kärdla crater, Estonia). Depending on the strength of the target, extensive collapse may occur along low-angle decollement surfaces, which can cause a concentric crater morphology (e.g., Chesapeake Bay crater, USA and Mjølnir crater, Norway).

Chesapeake Bay and Mjølnir), even though the relative target water depth was too shallow for a low angle excavation flow of the Lockne type. The concentricity may instead be due to low-angle slumping along a decollement surface surrounding the inner, deeper crater depression, possibly as a consequence of great differences in target strength and very little, if at all, related to the water depth. At Chesapeake Bay, whether there was any shallow excavation flow stripping sediments outside the central basement crater is still uncertain. Features such as resurge gullies may exist but also may have been severely altered by the block faulting. Strata are missing in the top parts of several of the slump blocks and in the vicinity of the crater (Powars and Bruce 1999). This may indicate that a shallow excavation flow had created a shallow outer crater in the sediments before the slumping begun. The resurge also may have removed some of the missing sediments. However, it is likely that the block slumping had begun already, before or during the resurge flow, supporting a formation due to the excavation flow. This issue needs to be studied further before it is possible to estimate the influence of the seawater on the morphology. If the intense block slumping is a consequence of the seawater in combination with certain thickness and properties of the target sediments, and if unconsolidated sediments occur on Mars, craters with this morphology could exist there as well. Clearly, Chesapeake Bay and other craters formed in very shallow relative water depth (e.g., Chicxulub) share many geological features, such as resurge deposits, with marine-target craters formed in deeper water. However, these features are not easily recognized in remote sensing.

When the water is deep enough to surpass the basement crater rim, both erosional and depositional features can evolve that give further indications of the target water depth. The flow of debris-loaded water back into the crater is strongly erosive due to sediment-bulking of the flow and increased density (von Dalwigk and Ormö 2001). This resurge can generate radial gullies with deep and narrow apices pointing toward the inner crater (Fig. 2a). Lindström et al. (Forthcoming) suggest that the resurge may be canalized by tears (rip-apart) in the ejecta deposited just outside the rim of the nested crater. At the Lockne crater (7.5 km nested crater diameter, 14 km outer crater diameter, 1 km target water depth), this ejecta has been largely emplaced as a semi-rigid flap, which must have been influenced by tangential stresses when deposited. Such tears would widen outward from the basement crater rim and be partly responsible for the wedge-shape of the resurge gullies. Even at great distances from the crater, there may exist traces of the resurge, such as ripple beds (Sturkell et al. 2000).

Lockne is the best known example of a crater with resurge gullies, but similar features occur also at Mjölnir

(45 km diameter, 500 m water depth; Tsikalas and Faleide Forthcoming), Kamensk (38 km diameter, 200 m water depth; Ormö and Lindström 2000), and at the possibly marine Tookoonooka crater (66 km diameter; Gostin and Therriault 1997). A strong indication that the features at these craters are resurge gullies are the floors of the gullies, which are covered by breccias and turbidites, similar to the established Lockne resurge deposit. Localized rim collapse at very shallow water (Fig. 2b) can cause breaches through the crater rim where water can enter less forcefully (e.g., Kärdla, 4 km; Puura and Suuroja 1992).

At great relative water depths, no impact crater develops in the seafloor. The only known terrestrial example of such an event is the Eltanin impact site off the coast of Chile (Gersonde et al. 1997). The explosion occurred mainly in the water mass, causing only minor excavation of the seafloor.

The characteristics of marine-target craters described above have been obtained from studies on Earth where it has been possible to use extensive drilling, sampling, and traditional geophysical methods (see Ormö and Lindström 2000). Sedimentation subsequent to the impact event has completely covered or subdued the topographic expressions of all the known marine-target craters on Earth. This has preserved their distinctive records from erosion, which can be displayed in seismic or drill core data. However, surface expression is mostly absent. Nevertheless, a buried crater can later be exposed at the surface due to subsequent erosion (e.g., the Lockne crater; Lindström et al. 1996b).

## DURATION AND EXTENT OF PUTATIVE WATER BODIES IN THE NORTHERN LOWLANDS OF MARS

The modeling of marine impact events requires definitive time constraints. We determine these constraints using models consistent with our interpretation of the paleohydrologic history of Mars. Using the conceptual framework described by Fairén et al. (2003), and the absolute age estimates of Hartmann and Neukum (2001), we estimate the duration of the Noachian-Early Hesperian ocean, defined by contact 1-Meridiani shoreline (Fig. 1; Table 1), to be between 100,000 yr and 800 Myr, likely reaching the upper limit. The duration of the Late Hesperian-Early Amazonian ocean, defined by contact 2 (Table 1), is estimated to have been between 100,000 yr and 800 Myr, most likely lasting about 100 Myr.

The lower time constraints are based on the minimum time required for an episodic ocean-forming event, described by Baker et al. (1991), as well as the formation of related geomorphic features, such as sapping channels and glaciers (e.g., Kargel and Strom 1992; Baker et al. 1991). The upper time constraints assume a quasi-steady input of endogenetically-driven flood waters from the Tharsis region (Baker et al. 1991, 2000; Dohm et al. 2001 a, b; Fairén et al. 2003). A northern plains ocean would eventually freeze, allowing steady-state conditions to be reached due to the insulating effects of the cap ice and geothermal heat from the ocean bottom. However, an unknown amount of water over geologic time would still be lost to the subsurface and atmosphere. Periodic floods, in the upper time constraint scenario, are envisioned to recharge the ocean by melting the cap ice and recycling depositional sediments.

The spatial extent of the northern plains oceans is assumed to be approximately consistent with contact 1-Meridiani shoreline and contact 2 (see Fig. 1; Table 1). Furthermore, an average ocean depth was selected based on the probability of a bolide striking any part of the ocean surface being approximately equivalent. Therefore, the surface area of the Noachian-Early Hesperian ocean, defined by contact 1-Meridiani shoreline, is approximately $5.4 \times 10^7$ km$^2$, with an average depth of about 2000 m and a maximum depth of 3.75 km (assuming present-day topography). The surface area of the Late Hesperian-Early Amazonian ocean, defined by contact 2, is $2.4 \times 10^7$ km$^2$, with an average depth of about 560 m (Head et al. 1999) and a maximum depth of 1.46 km (assuming present-day topography).

## POTENTIAL FOR THE DISCOVERY OF MARINE-TARGET CRATERS ON MARS

### Projectile Diameter from Numerical Simulations

Marine-target craters that are recognizable using the present remote-sensing technology must have been powerful enough to make distinct craters in the seafloor. The average water depth of the late, smaller ocean within contact 2 would have been about 560 m (Head et al. 1999). For the greater, earlier ocean that reached up to contact 1 and parts of Western Arabia Shelf (the Meridiani shoreline), the average water depth is estimated to have been about 2 km (Table 1). To generate a distinct crater in the seafloor of a 500 m-deep sea on Earth, the impactor should have a diameter of at least 250 m (based on Gault and Sonett 1982). Due to the differing gravity and average impact velocity on Mars as compared to the Earth, the minimum projectile diameter will be different.

We performed a numerical simulation using the Eulerian mode of SALE-B, a 2D hydrocode modified by Boris Ivanov based on SALES-2 (Gareth and Melosh 2002), to analyze the threshold diameter for cratering the seafloor of a martian sea with the water depth (h). For case 1, h = 500 m, and for case 2, h = 2000 m.

We assume vertical impacts, as they reduce to 2D problems using the radial symmetry, in hypothetical seas covering the northern lowlands. Hence, the gravity constant (g) is selected to the polar value of 3.76 m/s$^2$ (Yoder 1995).

The ANEOS analytical equation-of-state (Thompson and Lauson 1972), with input data for dunite and water, is used to prepare tables and to describe the material pressure and temperature versus density and internal energy of the projectile, water layer, and target materials in a wide range of both parameters. We omit modeling of oblique impact, as the code used does not permit a water layer with planar symmetry assumption.

The code enables two different materials. As we have water as the weaker upper layer, the projectile and the lower target materials have to be the same. Dunite, with density of 3.3 kg/m$^3$, was selected as the projectile and target material because it is described by a reasonably reliable equation-of-state (Benz et al. 1989). Study of a low-density projectile impact, such as a comet impact, would require a low density for the lower part of the target, which is not considered to be a valid option. Furthermore, only a small component of the flux of impactors is comets.

For a fixed water depth, a first set of simulations was done with a low-resolution computational grid. We used 20 cases of projectile diameter (d), depending on the water depth, with 10 nodes resolution to describe the radial projectile in a range of 40 m to 10 km to pinpoint the approximate projectile diameter (d*) capable of cratering the seafloor, as well as to determine the damaged zone where a fine mesh resolution is required. When a first set of simulations was achieved, we performed five additional simulations using a refined mesh and with a projectile diameter close to the diameter (d*). The non-uniform computational grid of the detailed simulations consisted of 401 nodes in the horizontal direction and 501 nodes in the vertical direction; the total nodes describe half of the crater domain because of axial symmetry. The grid size of the total domain is memory limited. We have enough points to describe the damaged zone, but due to wave reflection problems at the boundaries, we increased the mesh size progressively outward from the center with a 1.05 coefficient to have a larger spatial domain. The central cell region around the impact point where damage is greater (i.e. somewhat wider than the crater area) was a regular mesh with 150 nodes resolution in the x direction and 120 nodes resolution in the y direction, describing half of the damaged zone

The simulations used to achieve the minimum required impactor diameters (Fig. 3) were performed with spherical projectiles and a velocity (v) of 10 km/s, which can be considered close to the average impact velocity on Mars (Steel 1998). For reference, we also conducted simulations with the impact velocity of 30 km/s, as well as impacts into targets without a water layer. The increased impact velocity only generated an insignificantly wider water cavity and is, therefore, not presented in Fig. 3. The results from the land-target simulations are also not presented in Fig. 3 but are used for the crater rate calculations listed below.

With set projectile velocity and water depth, we examined the critical diameter of the projectile that can crater the seafloor. Based on the initial simulations, a simulation was performed using the best fit for a 500 m water depth (d = 150 m). Another simulation was performed using the best fit for a 2000 m water depth (d = 600 m). For reference, we have

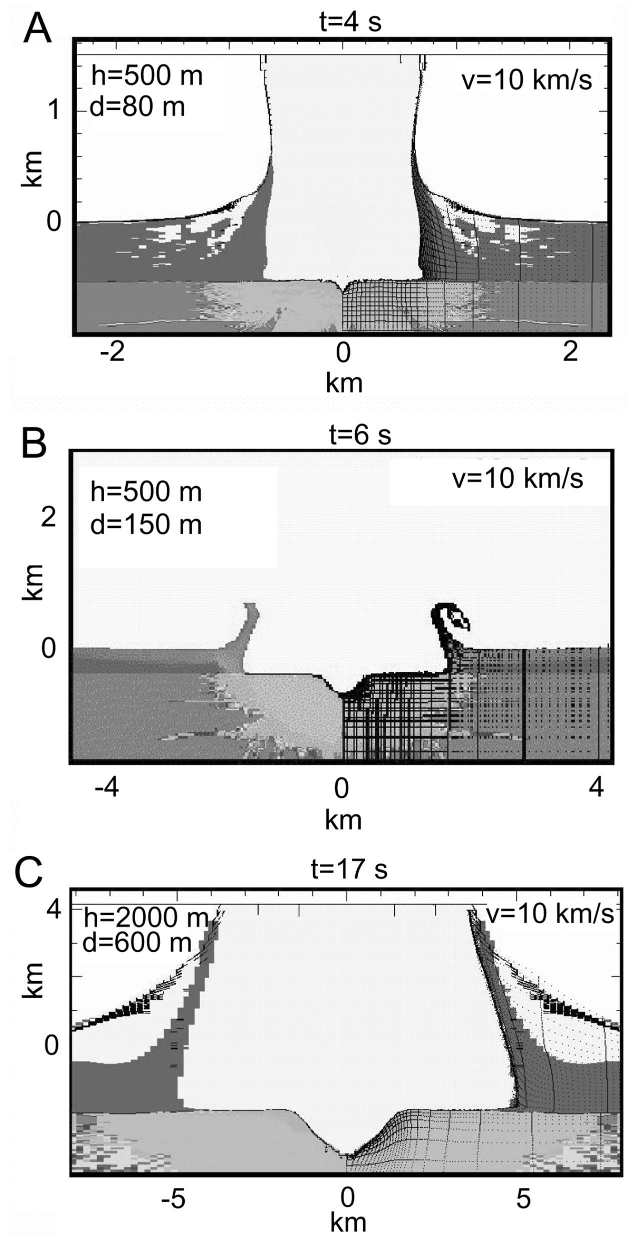


Fig. 3. Computer simulations of marine impacts on Mars. The simulations were performed to search for the minimum impactor diameter capable of producing a preservable (>3 Gyr; see Hartmann and Neukum 2001) crater in the seafloor of a putative sea: a) vertical impact of an 80 m-diameter impactor into a 500 m-deep sea. Note the cylindrical opening cavity without significant cratering of the seafloor; b) the same premises as in (a) but with a 150 m-diameter impactor. In this case, a distinct seafloor crater develops. Note the concentric shape of the opening cavity; c) vertical impact of a 600 m-diameter impactor into a 2000 m-deep sea.

chosen to present the simulation of the impact of an 80 m-diameter projectile in 500 m deep water.

Figure 3 shows the timeframes of the opening cavities for the maximum depth reached in the detailed simulations using the 80 m impactor, the 150 m impactor, and the 600 m impactor (with 2 km water depth). The grid represented in the figure is not the computational grid, but a grid to illustrate the damage. Each cell contains 100 points.

The difference between the d = 80 m case and the d = 150 m case illustrates the "protective effects of a layer of seawater" discussed by Gault and Sonett (1982). The excavation flow along the seafloor causes a flat-floored cavity with a high diameter-to-depth ratio very different from the bowl-shaped cavity typical for homogenous target (Melosh 1989). An 80 m projectile would have been sufficient to crater into the seafloor if the cavity had developed according to the standard model. Due to the water layer, the projectile diameter has to be increased significantly before cratering occurs in the seafloor. Only projectiles larger than ~150 m generate seafloor craters that have a chance to be visible after resurge modification and after long periods of time (based on Hartmann and Neukum [2001]). In a 2000 m-deep ocean, the projectile needs to be at least 600 m in diameter to generate a distinct seafloor crater (Fig. 3). The effect of obliquity on the final crater shape could not be tested in this study due to the software limitations. However, due to the higher energy release in the upper part of the target at oblique impact (Melosh 1989), to some extent, the result is similar to an increased impact velocity (Ormö et al. 2002). Thus, a comparison can be made with our simulation of a 30 km/s impact, which only caused a slightly higher ratio between the water cavity diameter and the inner crater diameter. Shuvalov et al. (Forthcoming) has done a more detailed study of oblique impacts in a marine environment. They noticed how the water cavity becomes offset in the down-range direction. However, the diameter of the final seafloor crater does not change significantly compared to the vertical impact.

After establishing the critical impactor diameters (150 m and 600 m), we also needed to model the final crater diameter in a target without a water layer for each of the impactors. The land-target crater diameter is needed for comparisons with published cratering rates for Mars (e.g., Hartmann and Neukum 2001; Hartmann et al. 1981; Neukum and Hiller 1981; Fig. 2 in Hartmann and Neukum 2001). The 150 m impactor generates a final crater somewhat less than 2 km in diameter, and a 600 m impactor generates a final crater about 8 km in diameter.

**Number of Possible Marine-Target Craters**

Due to the position of Mars relative to the main belt of asteroids located between the orbits of Mars and Jupiter, the impact rate on Mars may be somewhat higher than on the Moon, potentially up to two times higher (Hartmann et al. 1981; Neukum and Hiller 1981). However, the high impact rate may not result in a higher cratering rate (Ivanov 2001); this can be explained by the larger surface gravity and the lower impact velocity, as well as the overall protection that a denser atmosphere and a permanent hydrosphere may have against the mean-sized impactor populations, which results in a smaller crater size compared to the same impact on the Moon. The number of marine impact craters that may have formed on Mars can be estimated based on the cratering rate of projectiles capable of cratering the seafloor of the putative oceans (150 m diameter in the case of a contact 2 ocean and 600 m diameter in the case of a contact 1-Meridiani ocean, see modeling above) and the spatial area and duration of the postulated oceans.

When correlating our obtained impactor diameters with published data on martian cratering rates, it is evident that, at the present impact rate, very few (probably less than 1–2 bolides with the required diameter of 150 m) will strike per 100,000 yr on an area on Mars equal to the area within contact 2. However, for the same area with the exposure time equal to the maximum estimated duration for the contact 2 ocean (0.8 Gyr), as many as 1400 craters may have formed.

Before 3.5 Gyr ago, the cratering rate was higher (Hartmann and Neukum 2001), maybe about 200 times higher (Hartmann 1965, 1966), which means that more impacts may have occurred during the time of the Noachian/Early Hesperian ocean (contact 1-Meridiani). Still, the conservative estimate of 100,000 yr is so short that very few, if any, 600 m-diameter impactors may have struck the contact 1-Meridiani ocean. With the maximum estimated duration of 0.8 Gyr, about 160 impacts of that magnitude may have occurred.

Hence, for the later contact 2, the number of impacts of the minimum required impactor diameter (150 m) ranges from ~1 to ~1400. For the earlier, greater contact 1-Meridiani ocean, the number of impacts of the minimum impactor diameter (600 m) ranges from about 0 to 160. Of course, in addition to impacts from the minimum impactors, a few larger impacts also may have occurred in each of the oceans at their estimated maximum durations (0.8 Gyr).

If we do not assume average water depths for each ocean in our calculations, then we will get different values, as more parameters would lead to greater complexity. Therefore, we have defined the specific parameters used in our calculations for this study. Note that the area between contact 2 and the higher contact 1-Meridiani is of about the same size as the area within contact 2. In addition, this area would also have about the same water depth during the contact 1-Meridiani ocean, which is similar to the average water depth of the later contact 2 ocean. If we use this water depth (500 m) in our calculations for these areas of the contact 1-Meridiani ocean, while assuming a higher cratering rate, then 200 times as many 150 m impactor hits could have occurred, as in the minimum and maximum alternatives for the contact 2 ocean. Consequently, in the deeper parts of the basin (for both oceans), there would be far fewer craters formed by each of the minimum impactors used in our study.

Our assessment is that the number of marine-target craters is low if using shorter oceanic phases. Furthermore, for these craters to be detected, they both must have been preserved during much of the geologic history of Mars and must be exposed at present.

## Exposure and Examples of Possible Marine-Target Craters

Marine-target craters have their preservation hampered by various environmental factors. Close to the shoreline, and especially near the outflow channels, smaller marine-target craters are probably covered by huge amounts of fluvial and colluvial sediments, and features like resurge gullies at bigger craters may be hard to recognize. Large quantities of material have apparently been transported to the northern plains by repeated catastrophic floods (Baker et al. 1991; Dohm et al. 2001a, b; Fairén et al. 2003), and eolian deposition has persisted for much of the suitable areas for potential marine-target craters (Carr et al. 1984). Furthermore, impact structures can act as efficient sediment traps (Lindström et al. 1994; Hartmann and Neukum 2001). Beals and Hitchen (1970) conducted experiments on how much sediment is necessary to completely obliterate a crater. They simulated both mechanical and chemical deposition. Their simulation indicated that the sediment thickness would need to be approximately three times the floor-to-rim depth of the crater (i.e., 730 m-thick sediments at a 1.6 km-wide, fresh bowl-shaped crater). If the crater lacks much of the rim already at the beginning of sedimentation, which would be typical for a concentric marine-target crater, the required sediment thickness can be even less. Hartmann and Nekum (2001) estimated the maximum lifetime of 350 m scale craters (the craters in our modeling are ~1–3 km in diameter) to be about 3.0 Gyr. A crater with a 16 km diameter has a lifetime of only ~3.5 Gyr due to an infill rate one or two orders of magnitude higher before ages ~3 Gyr ago. Also, the erosion rate was two to three orders of magnitude higher in the Noachian to Hesperian than in the Hesperian to present (Golombek and Bridges 2000). If, by chance, a martian marine-target crater is visible at the surface, it is likely that sediment fill has subdued the morphological expression of the crater (e.g., concentricity and gullies).

In regard to marine-target craters, what we can expect to distinguish in orbital images of Mars are rimless concentric craters with long and wedge-shaped resurge gullies. The gullies should show the same state of preservation (indicative of age) as the rest of the crater. Hence, they should not be confused with channels formed by fluvial activity or ground water sapping after the crater was formed. On Earth, where sampling is possible, the resurge-infill of the gullies connects them to the cratering process. Craters with breaches through their rims occur frequently on Mars. The majority appear to have been caused by fluvial activity, slumping, and groundwater-sapping long after the crater formation (e.g., Scott and Tanaka 1986; DeHon 1992; Scott 1993; Scott et al. 1995; Rotto and Tanaka 1995; Nelson and Greeley 1999; Cabrol et al. 2001a–c). This is indicated by anastomosing patterns, adherent incised channels over the crater floor, and/or deltas within the crater (Ori et al. 2000).

As previously mentioned, the concentricity of the marine-target impact crater is thought to be due to the layered target with seawater and sediments on top of a more rigid basement. However, concentric craters may form at any situation where a weaker target material overlays a rigid one. Although it has only been shown in experiments and lunar craters <1 km in diameter (Quaide and Oberbeck 1968), this relation has been reported for Mars (Gilmore 1999). The craters presented by Gilmore (1999) are located in Xanthe Terra and below contact 1-Meridiani. The concentric shape is obvious, but they lack other features that could be indicative of marine-target craters (Fig. 4). They also have a strikingly low amount of sedimentary infill.

Ormö and Muinonen (2000) performed a systematic search for marine-target craters in the northern lowlands and other large basins where seas may have existed. They used Viking images because of the complete coverage and the adequate resolution. They suggested two possible candidates on the Western Arabia Shelf with locations near contact 2 (Figs. 1 and 5). The craters have a concentric shape, flow structures on the crater infill, and breaches through the rims that show some resemblance to gullies and collapse features of terrestrial marine-target craters. These impact craters are the best candidates found so far on Mars. On Earth, continuous sedimentation has occurred at all known marine-target craters, resulting in complete burial of all the craters. However, this is not a strong argument against the interpretation. The craters may have formed during a short transgression of the sea. The sediment rate in martian paleoceanic and paleolacustrine environments is unknown. The craters may also have been exposed by the removal of the sediment cover, similar to the presently exposed 455 Myr Lockne crater, Sweden. There are strong indications from Mars Orbiter Laser Altimeter (MOLA) data that a large number of buried craters exist in the northern lowlands. Frey et al. (2001) report 644 "quasi-circular depressions" with diameters larger than 50 km, which, if craters, supports a Noachian age for the buried northern plains basement surface. Such craters may have originated before the contact 2 ocean formed on the lowlands, and so some of those 644 impacts likely occurred when the contact 1-Meridiani ocean was present, and the rest occurred in the frozen standing residue that may have occupied the lowlands between the two oceanic phases (Fairén et al. 2003; Fairén and Dohm 2003). Kreslavsky and Head (2001) presented what they called "stealth craters" in the northern lowlands. The smallest identified "stealth crater" is 6 km in diameter. They found about 140 "stealth craters" larger than 16 km. They estimated an early Hesperian age for that buried surface.

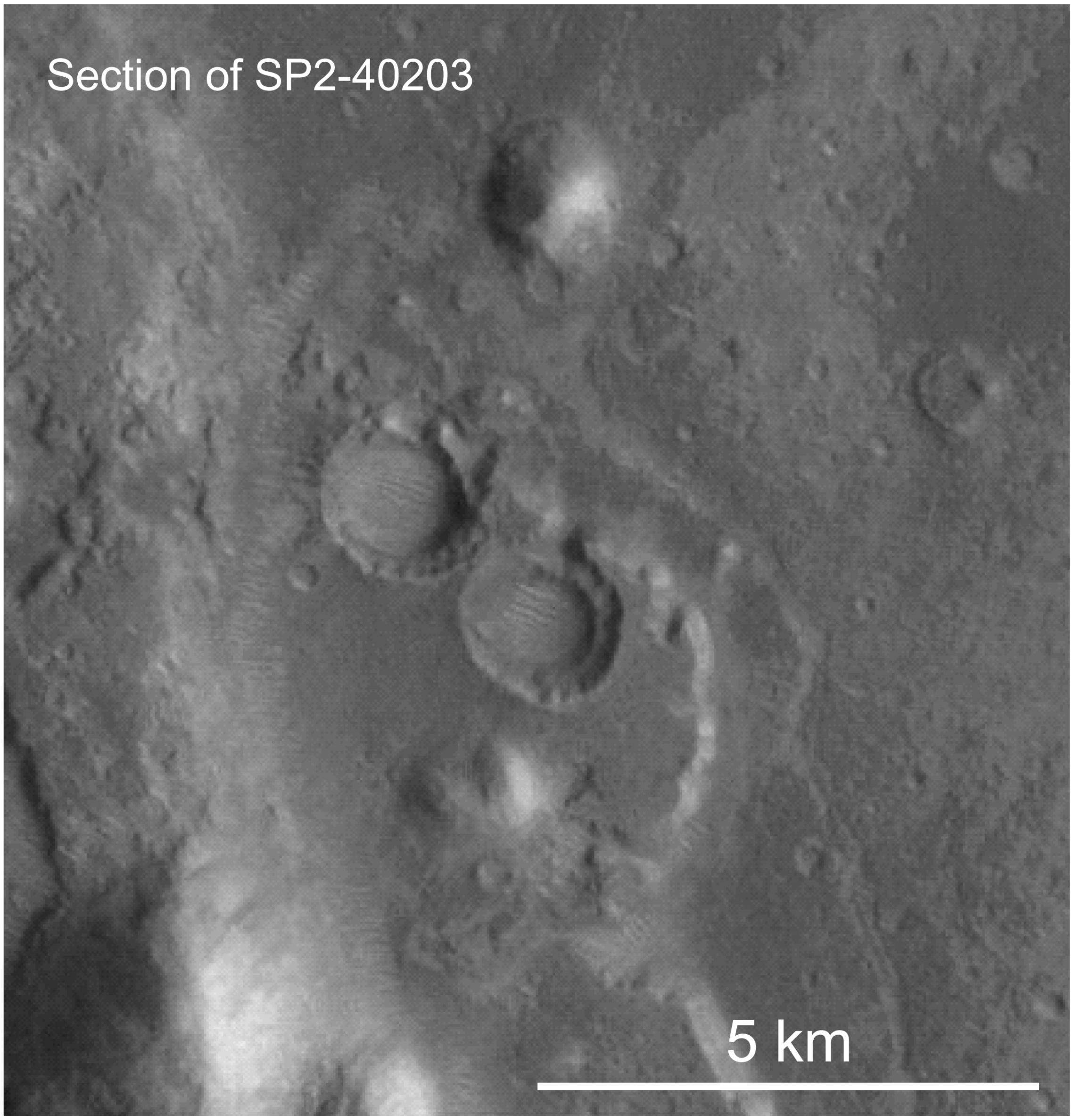


Fig. 4. Section of MOC image SP2–40203 showing concentric craters presented by Gilmore (1999) as examples of craters formed in layered target. The original image resolution is 13.94 m/pixel. The original image is centered at 45.43°W 13.10°N.

## DETECTION IN UPCOMING SURVEYS AND IMPLICATIONS

The upcoming ground penetrating radar experiments on the ESA Mars Express (MARSIS) and NASA 2005 (SHARAD) missions will have the potential to see buried structures in the upper crust. The Mars Advanced Radar for Subsurface and Ionospheric Sounding (MARSIS) will have a lateral resolution of about 5–10 km and vertical resolution of 50–100 m (Marinangeli et al. 2001). In general, the geomorphologic features such as concentricity and resurge gullies are too small for known terrestrial marine-target craters to be conclusively detected using the poor resolution of the MARSIS. One of the largest marine-target craters on Earth with documented resurge gullies is the Kamensk crater, Russia (see Ormö and Lindström [2000] for review). It has a 20 km-wide inner crater surrounded by a 38 km-wide shallow outer crater with up to 20 km-long and up to 4 km-wide gullies. On Mars, the weaker gravitation causes the simple/complex crater transition to occur at greater crater sizes than on Earth. It is possible that tens of km-wide buried craters may be discovered with the resolution of MARSIS. The Shallow Subsurface Sounding Radar (SHARAD) will have a vertical resolution of some tens of m and a spatial resolution of the order of some hundreds of m (Biccari et al. 2001), possibly allowing for the detection of smaller craters and even large gully structures. The detection of buried craters in the northern plains would be valuable for understanding the

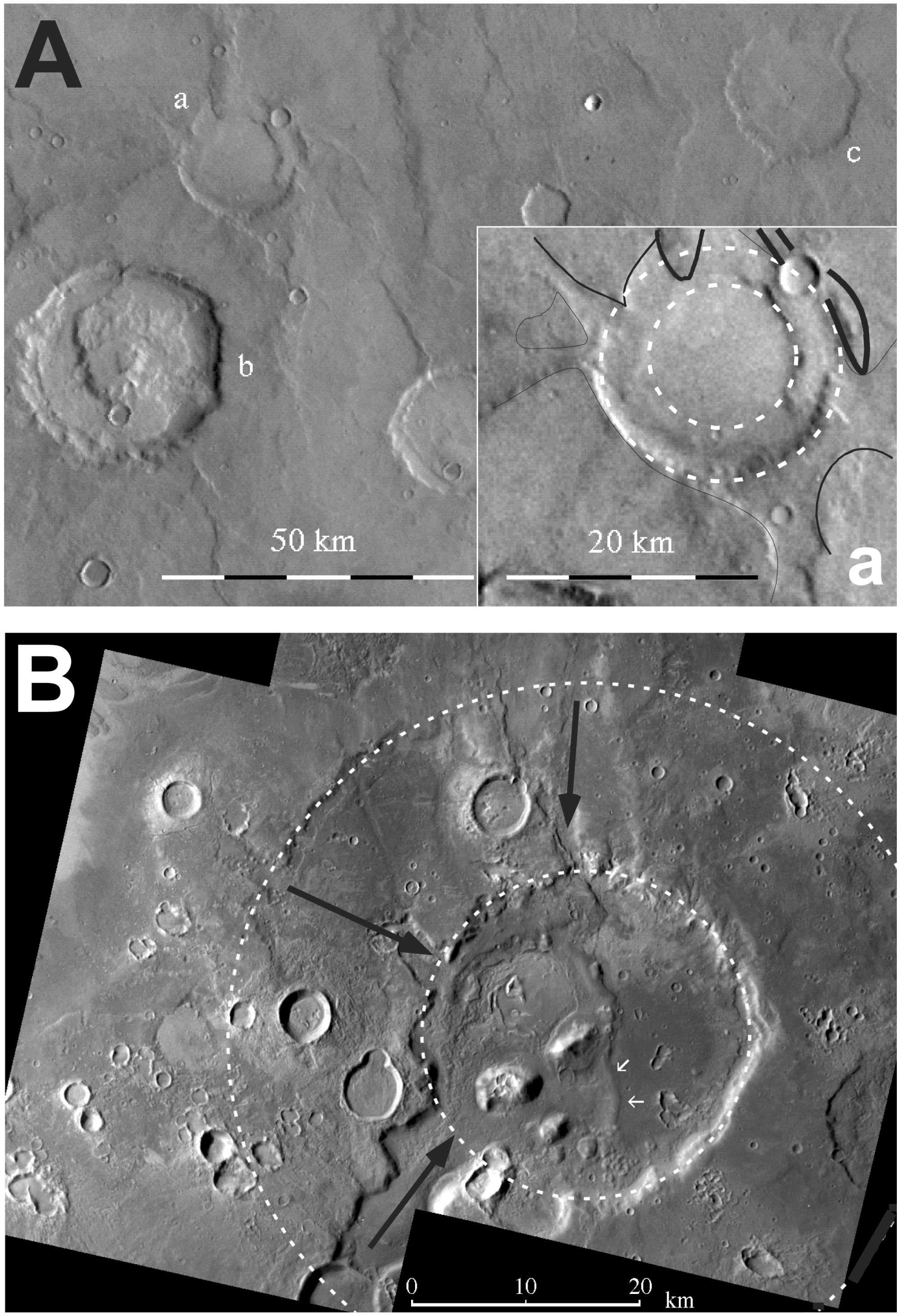


Fig. 5. Possible marine-target craters proposed by Ormö and Muinonen (2000) in the northern lowlands of Mars (for location see Fig. 1): a) concentric crater with gully-like features ("a"). The location is 38°N, 13°E. The inset figure shows an attempt to interpret where concentricity (white stippled circles) and possible breaches ("gullies") through the rim (dark lines) occur. Crater "b" has a terraced central peak that could indicate a standing body of water at some stage. Crater "c" shows a similar morphology to crater "a." Original Viking image F529A09; b) large crater with localized rim collapse and gully-like features. The location is 39°N, 10°W. The original image resolution is ~43 m/pixel. The stippled line indicates a possible, weak outer crater. The black arrows point at breaches in the rim that may resemble resurge gullies. The white arrows indicate a possible debris flow front from a flow that appears to have passed through a row of collapsed smaller craters in the lower left part of the image. The image is a mosaic of the three Viking images F072A30, F072A31, and F072A32.

formation of the dichotomy and aiding in the dating of the fluvial events that have brought sediments to the plains. In addition, if the craters show features typical of terrestrial impacts into an aqueous environment, further credence would be given to the suggested Oceanus Borealis hypothesis and useful information would be provided for future paleoclimatic reconstructions. It is important to remember that only drilling through the complete resurge sequence within a crater can provide definitive evidence for a marine impact, while the geomorphology of the crater can only be considered indicative.

## SUMMARY

Accounting for the martian impact rate, plus the temporal duration and spatial coverage of the hypothesized paleoceans and/or lakes, it should be possible to find impact craters that resemble craters formed in a marine environment on Earth. Our calculations indicate that, in an early, Noachian/Early Hesperian contact 1-Meridiani ocean, 160 marine-target craters may have formed when assuming a maximum duration of 0.8 Gyr and an average water depth of ~2 km. In a later, smaller contact 2 ocean, which has been suggested to have existed during the Late Hesperian to the Early Amazonian, as many as 1400 seafloor craters may have formed when assuming a maximum duration of 0.8 Gyr and an average water depth of ~500 m.

The lowland areas of Vastitas Borealis, the Western Arabia Shelf, and the large impact basins would be the natural places to locate marine-target craters. However, as is the case for most of the marine-target craters on Earth, thick blankets of sediment most likely cover the martian analogues. The upcoming ground penetrating radar experiments MARSIS and SHARAD may detect buried crater structures with features indicative for marine impacts.

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