# Differential subsidence and rebound in response to changes in water loading on Mars: Possible effects on the geometry of ancient shorelines

D. W. Leverington and R. R. Ghent

Center for Earth and Planetary Studies, National Air and Space Museum, Smithsonian Institution, Washington, D. C., USA

Received 22 June 2003; revised 23 October 2003; accepted 30 October 2003; published 8 January 2004.

[1] If large liquid or solid water bodies existed in the past in Martian basins such as Argyre, Hellas, and the northern lowland plains, they would have represented massive surficial loads. Earth analogs suggest that the magnitudes of crustal displacements related to water loading and unloading would have varied spatially within and around affected basins. It is likely that the effects of bathymetric irregularities on lithospheric flexure would have caused shoreline areas proximal to deep regions to be depressed (and, subsequently, to rebound) to a greater extent than shoreline areas located near shallow regions. Crustal and mantle inhomogeneities would have likely further contributed to spatial variation in water-related subsidence and rebound. Simple two-dimensional elastic flexure models suggest that Martian shoreline features of common age and formed in association with a large ancient water body could vary in elevation today by hundreds of meters as a result of these differential effects. Examples of similar shoreline offsets on Earth suggest that sets of Martian strandline features will reflect the changing nature of loading through time. *INDEX TERMS:* 6225 Planetology: Solar System Objects: Mars; 5475 Planetology: Solid Surface Planets: Tectonics (8149); 1824 Hydrology: Geomorphology (1625); *KEYWORDS:* Mars, ocean, lake, loading, isostasy, shoreline

**Citation:** Leverington, D. W., and R. R. Ghent (2004), Differential subsidence and rebound in response to changes in water loading on Mars: Possible effects on the geometry of ancient shorelines, *J. Geophys. Res.*, *109*, E01005, doi:10.1029/2003JE002141.

## 1. Introduction

[2] The valley and channel systems of Mars are indicative of the past presence of running water at the surface [*McCauley et al.*, 1972; *Masursky*, 1973; *Milton*, 1973; *Baker and Milton*, 1974; *Sharp and Malin*, 1975; *Baker*, 1982; *Carr*, 1996], and suggest that the early Martian environment may have been different from the cold and dry conditions of today [*Sagan et al.*, 1973; *Masursky et al.*, 1977; *Pollack et al.*, 1987; *Baker et al.*, 1991; *Clifford*, 1993; *Jakosky and Phillips*, 2001; *Craddock and Howard*, 2002; *Hynek and Phillips*, 2003]. The past occurrence of flowing water on Mars implies that water would likely have been impounded in catchment basins [e.g., *De Hon*, 1992], although strong evidence in support of the largest hypothesized water bodies [e.g., *Parker et al.*, 1993; *Clifford and Parker*, 2001] remains elusive [*Malin and Edgett*, 1999; *Carr*, 2001; *Bandfield et al.*, 2003; *Tanaka et al.*, 2003]. The discovery of clear geomorphological evidence for the past existence of large Martian water bodies would have profound implications for our understanding of the climate and volatile history of Mars. Geomorphological or textural transitions that occur at roughly uniform elevations are typically central to the search for candidate nearshore features formed in association with such hypothetical water bodies. However, potentially significant effects related to water loading have largely been neglected in the identification and evaluation of candidate nearshore features, and in the determination of possible basin forms and likely relations between presumed water levels and the levels of tributaries and outlets. In this paper, we consider the manner in which basins can be differentially influenced through loading and unloading by water in its liquid and solid forms. We also present two-dimensional elastic beam flexure calculations that are useful for demonstrating the manner in which differential subsidence and rebound can occur due to the loading of irregularly-shaped basins with water. Finally, we discuss implications for the locations and geometries of possible preserved Martian shoreline materials.


0148-0227/04/2003JE002141$09.00

## 2. Background

[3] Geomorphic features that have been used to identify the locations of past water bodies on Mars include terraces and elongate slope anomalies, which have been interpreted as possible wave terraces and other shoreline features [e.g., *Parker et al.*, 1989, 1993; *Ori and Mosangini*, 1998; *Head et al.*, 1999; *Cabrol and Grin*, 1999; *Ori et al.*, 2000; *McGill*, 2001]. Accumulations of material at the mouths of inlet channels have been interpreted as possible lacustrine deltas and fans [*Grin and Cabrol*, 1997; *Ori et al.*, 2000; *Cabrol and Grin*, 2001], and regions of low relief and relative smoothness in the interiors of drainage basins may have been subjected to past aqueous sedimentation [*Goldspiel and Squyres*, 1991; *Ori and Mosangini*, 1998].

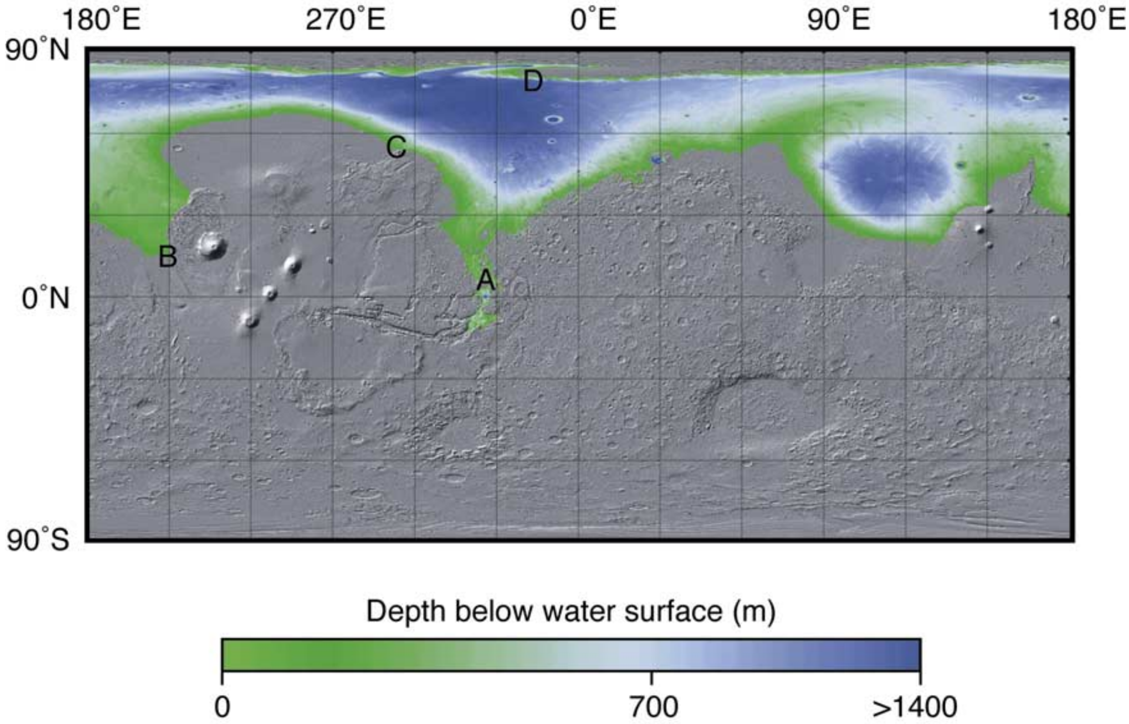


**Figure 1.** Bathymetric map of a hypothetical northern ocean on Mars, based on the −3700 m level of *Parker et al.* [1993] and on the simplifying assumption that modern topography can be used as a gross approximation for ancient topography at the time of a northern ocean. If such an ocean actually existed, it would have extended irregularly over much of northern Mars with water depths well in excess of 1 km over large areas. Letters denote regions discussed in section 4. Topographic data after *Smith* [2003].

The association of outlets with otherwise enclosed basins is suggestive of the past presence on Mars of water bodies of a wide range of sizes [e.g., *De Hon*, 1992; *Cabrol and Grin*, 2001; *Irwin et al.*, 2002; *Milkovich et al.*, 2002]. Also cited as potential evidence for the locations of past lakes on Mars is the presence, at the surface of basin interiors, of hematite deposits [*Christensen et al.*, 2000, 2001] and possible deposits of evaporite minerals [*Forsythe and Zimbelman*, 1995; *Cabrol and Grin*, 1999].

[4] Water bodies on Mars may have taken the form of lakes located in a wide variety of basin types, including craters [*Newsom et al.*, 1996; *Grin and Cabrol*, 1997; *Forsythe and Blackwelder*, 1998; *Cabrol and Grin*, 1999; *Cabrol et al.*, 1999; *Rathbun and Squyres*, 2002; *Milkovich et al.*, 2002], canyons [e.g., *Lucchitta*, 1982; *Nedell et al.*, 1987; *De Hon*, 1992], and regions of chaos [e.g., *Ori and Mosangini*, 1998]. Surface water bodies may also have been associated with more subtle topographic depressions in open areas [e.g., *Goldspiel and Squyres*, 1991; *De Hon*, 1992].

[5] Large Martian basins in both the lowlands and highlands are proposed to have contained extensive liquid or solid water bodies in the past [e.g., *Chapman*, 1994; *Dohm et al.*, 2001; *Ivanov and Head*, 2001; *McGill*, 2001; *Thomson and Head*, 2001]. A number of horizontal or near-horizontal terrace-like features have been identified within Utopia Basin, and have been interpreted as possible shorelines formed in association with a large water body [*Head et al.*, 1999; *Thomson and Head*, 2001; *McGill*, 2001]; on the basis of Mars Orbiter Laser Altimeter data [*Smith*, 2003], such a water body would have had a maximum diameter of almost 2000 km and maximum depths over large areas of about 650 m (to maximum local depths of about 2 km within the deepest craters of the basin). *Moore and Wilhelms* [2001] found that scarps and contacts of sedimentary units within the Hellas basin trace contours of constant elevation over distances of thousands of kilometers, and proposed that they formed in association with a large body of liquid and solid water with an implied maximum diameter of about 2200 km and maximum depths over large areas of about 3.5 km (to maximum local depths of about 5 km). A large drainage system has been proposed to have connected the south polar region to Chryse Planitia through the Argyre basin, with water level in the Argyre basin constrained by the level of a northern outlet [*Parker et al.*, 2000, 2003; *Clifford and Parker*, 2001]; the Argyre water body would have had an implied diameter of about 1400 km and maximum water depths over large areas of about 3 km (and of 5 km within the deepest craters of the basin). On the basis of the presence of discontinuously extensive margins that separate textural, morphological, or drainage units in the northern hemisphere of Mars, an ancient ocean has been hypothesized to have extended over much or all of the northern lowland plains [*Lucchitta et al.*, 1986; *Parker et al.*, 1989, 1993; *Baker et al.*, 1991, 2000; *Head et al.*, 1998, 1999; *Clifford and Parker*, 2001] with maximum depths up to (or possibly exceeding) about 3 km (e.g., Figure 1); as with other large Martian basins where water bodies have been proposed, however, analyses of MOC high-resolution imagery of the northern plains do not provide clear support for such interpretations [*Malin and*

*Edgett*, 1999; *Carr*, 2001], and plausible alternative interpretations to shorelines have been suggested [e.g., *Carr and Head*, 2003]. A smaller northern ocean has been proposed with an extent defined in part by the distribution of exposed Upper Hesperian materials of the Vastitas Borealis Formation [*Kreslavsky and Head*, 2002; *Carr and Head*, 2003].

## 3. Loading by Large Water Bodies on Earth

[6] Both solid and liquid water bodies on Earth act as surface loads, and changes in their size and extent can cause significant lithospheric deflection. For example, during continental glaciation, the formation and expansion of an ice sheet causes elastic terrain subsidence below the ice sheet and in broad regions adjacent to its margins [*Walcott*, 1970; *Peltier*, 1974, 1985; *Peltier and Andrews*, 1976]. In addition, this subsidence can induce lateral displacement of material within the viscoelastic mantle, resulting in the growth of peripheral flexural bulges as well as other more distant changes in the geoid [e.g., *Walcott*, 1972a; *Chappell*, 1974; *Peltier*, 1994]. The reduction in the mass of an ice sheet during and following deglaciation causes terrain emergence in the region of loading through the combined processes of glacio-isostatic rebound and lateral redistribution of mantle materials [e.g., *Walcott*, 1972b; *Farrell and Clark*, 1976; *Clark et al.*, 1978, *Andrews*, 1989]. Within a region influenced by glacial loading, the nature and magnitude of crustal displacement in response to isostatic rebound and redistribution of mantle material will vary from location to location because of spatial variation in factors such as 1) the position and form of the loading ice sheet, and 2) the physical characteristics of the underlying crust and mantle [e.g., *Andrews*, 1970, 1982; *Walcott*, 1972a; *Peltier*, 1974].

[7] As with ice sheets, the presence and growth of large liquid water bodies at the surface of the Earth cause crustal loading and subsidence. Separate sub-regions of both the continents and oceans respond differently to changes in oceanic loads, and as a result the addition or removal of a particular volume of water to the Earth's oceans does not elevate or reduce global relative sea levels by a geographically uniform amount. When a total water volume of more than 50,000,000 km$^3$ [*Lambeck et al.*, 2002] was transferred from the continents to the oceans during the last deglaciation, the oceans did not rise uniformly with respect to land units. Instead, water level rose with respect to land by different amounts in different areas, due to the combined effects of differential glacio-isostatic rebound, spatial variation in the degree of water loading, and far-reaching lateral movements of mantle materials; even locations far from the glacial melting associated with deglaciation were affected in this manner [*Walcott*, 1972a; *Farrell and Clark*, 1976; *Clark et al.*, 1978]. Because of the combined effects of both glacial unloading of the continents and meltwater loading of the oceans, the vertical movement of land during deglaciation may have varied by as much as 50% to 200% of the late-glacial "eustatic" rise of $\sim$120 m. When the estimate of variation is restricted only to consideration of the changing water load on ocean basins (with no elastic rebound from glacial unloading, and with vertical basin sides and thus no change in the area of the oceans), the variation in sea level in the vicinity of the shorelines over the entire planet is expected to be 6% of sea level rise [*Farrell and Clark*, 1976]. Global models that take into account flooding and loading of new areas have not been constructed, but regional models for flooding of the Persian Gulf region during deglaciation suggest possible local variations of about 12% of sea level rise [*Lambeck*, 1996], implying the potential for global hydroisostatic variations in excess of this value.

## 4. Water Loading on Mars

[8] Hypothesized large ancient water bodies on the surface of Mars would have represented massive loads that could have tectonically deformed the Martian lithosphere [e.g., *Hiesinger and Head*, 2000; *Parker et al.*, 2000; see also *Johnson et al.*, 2000]. Terrestrial experience suggests that the growth and subsequent removal of large liquid and solid water bodies on Mars should produce differential loading and rebound. Models of the extent or depth of ancient water bodies must therefore account for deflection of the lithosphere due to loading by water (Figure 2). Understanding the magnitude and geometry of this deflection may aid in the search for and evaluation of possible present-day remnants of nearshore features, and in the interpretation of the relationships between large basins that may have held water and the positions and elevations of related features such as proximal drainage divides.

[9] We can gain valuable insight into this problem from a simple calculation of the two-dimensional elastic flexure resulting from an asymmetric water load using a Cartesian geometry. The results presented below serve to demonstrate the manner in which simple load asymmetry can lead to potentially significant differential rebound, and yields insight into other geometric aspects of flexure resulting from an asymmetric water load. Our goal is to compare the relative positions of shoreline features following unloading of a hypothetical asymmetric basin filled with water (see Figure 2) with the relative positions of the same features when they originally formed. We calculate the amount of deflection experienced by a basin of known loaded (full) shape that has undergone flexure due to water loading, and from this we determine the unloaded (empty) shape. We then compare the locations and elevations of shorelines between the loaded and unloaded cases. We treat the lithosphere as an elastic layer of uniform thickness and mechanical properties overlying an inviscid substrate. We investigate three cases: 1) a 100 km wide basin with a maximum depth of 1 km; 2) a 500 km wide basin with a maximum depth of 3 km; and 3) a 2500 km wide basin with a maximum depth of 5 km. For each case, we consider values of 25 km, 50 km, and 100 km for elastic thickness ($T_e$); these values span a range consistent with current estimates of elastic thickness [e.g., *Zuber et al.*, 2000; *McGovern et al.*, 2002; *McKenzie et al.*, 2002] (note that Martian crustal heat flow in the past may have been higher than at present [*Zuber et al.*, 2000], suggesting a correspondingly thinner $T_e$). Values for the other parameters used in our calculation are given in Table 1. Because the horizontal dimensions of these loads are small relative to the radius of Mars, a flat plate approximation that neglects the geoid displacement associated with depression of the lithosphere into the mantle is appropriate; though this approximation starts to break down for the 2500 km wide

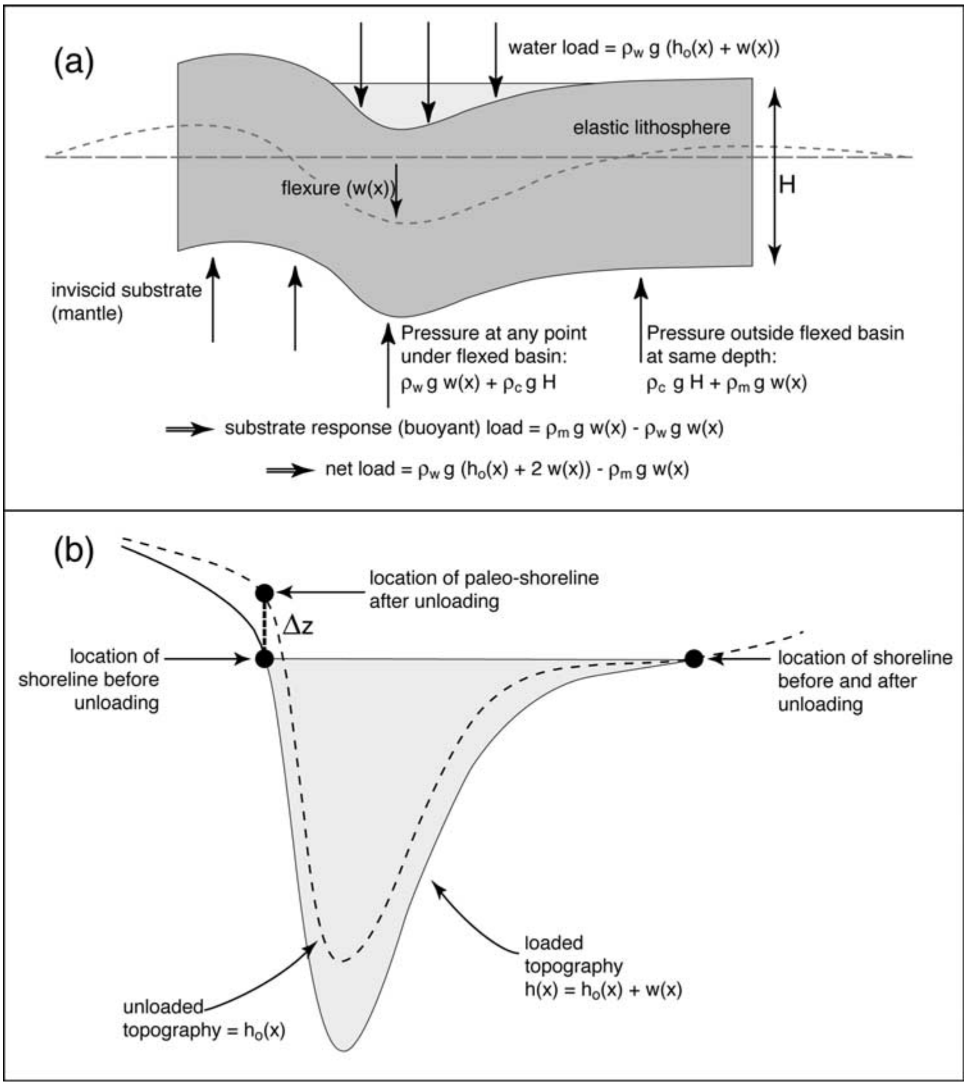


**Figure 2.** (a) Schematic diagram of flexural response of the elastic lithosphere to loading an asymmetric basin with water, showing the loads relevant to this problem. (b) Enlarged schematic diagram of basin shape showing loaded topography (solid curve) and unloaded topography (dashed curve). Filled circles denote positions of shorelines before and after unloading. Vertical offset in shoreline position is denoted by $\Delta z$.

basin, we retain that case because it helps to illustrate the elastic behavior of the lithosphere for loads much wider than the elastic thickness. The flat plate approximation also neglects membrane stresses, which should represent the dominant support mechanism on Mars only for topographic loads wider than about 2700 km [*Turcotte et al.*, 1981].

[10] The flexure is given by

$$D\frac{d^4w(x)}{dx^4} = q(x) \tag{1}$$

[e.g., *Turcotte and Schubert*, 2002]; $w(x)$ is the flexure (positive downward), $q(x)$ is the net load acting on the elastic layer, and $D$ is the flexural rigidity, given by

$$D = \frac{ET_e^3}{12(1-\nu^2)} \tag{2}$$

where $E$ is Young's modulus, $\nu$ is Poisson's ratio, and $T_e$ is elastic layer thickness. The loads in this linearized problem are separable; because we are interested only in the effect of the water load on the elastic layer, we can ignore the topographic load from the slightly non-uniform elastic thickness in the neighborhood of the basin. The net load $q(x)$ for our problem is

$$q(x) = \rho_w g[h_o(x) + 2w(x)] - \rho_m g w(x) \tag{3}$$

where $\rho_m$ and $\rho_w$ are mantle and water densities, respectively, $g$ is gravitational acceleration, and $h_o(x)$ is

**Table 1.** Parameter Values for Flexure Calculation

| Parameter | Symbol | Value | Units |
|---|---|---|---|
| Load (water) density | $\rho_w$ | 1000 | kg/m$^3$ |
| Mantle density[a] | $\rho_m$ | 3500 | kg/m$^3$ |
| Young's modulus[a] | E | $1 \times 10^{11}$ | Pa |
| Poisson's ratio | $\nu$ | 0.25 | – |
| Gravitational acceleration | g | 3.72 | m/s$^2$ |

[a]For example, *Johnson et al.* [2000] and *McGovern et al.* [2002].

the unloaded basin shape (see Figure 2a for derivation). Substituting (3) into (1) yields

$$D\frac{d^4w(x)}{dx^4} + (\rho_m - 2\rho_w)gw(x) = \rho_w g h_o(x) \tag{4}$$

Because we wish to assign the form of the loaded basin, we express the unloaded basin shape, $h_o(x)$, as

$$h_o(x) = h(x) - w(x) \tag{5}$$

Finally, this yields

$$D\frac{d^4w(x)}{dx^4} + (\rho_m - \rho_w)gw(x) = \rho_w g h(x) \tag{6}$$

Because we are interested in the differential flexure associated with asymmetric loading, we use a function of the form $x_L^2 e^{-px_L}$ over an interval $0 < x_L \leq$ L for the loaded topography (Figure 2). We solve equation (6) by approximating the basin topography $h(x_L)$ as the sum of narrow rectangular loads of amplitude $h_n$ [see *Watts*, 2001]. The flexure equation for the nth rectangular component is subject to the following boundary conditions: $w\ (x \rightarrow \pm\ \infty) = 0$; $dw/dx\ (x = x_{Ln}) = 0$; and $\rho_w \int_{-\infty}^{\infty} h(x)\ dx = (\rho_m - \rho_w) \int_{-\infty}^{\infty} w(x)dx$. The flexure due to each narrow rectangular load component is given by

$$w_n(x) = \frac{\rho_w h_n}{2(\rho_m - \rho_w)} \int_a^b e^{-\lambda x}(\cos\lambda x + \sin\lambda x)dx \tag{7}$$

where $h_n$ is the amplitude of the nth rectangular load component, and $\lambda$ is the flexural wavelength, given in this case by

$$\lambda = \left(\frac{(\rho_m - \rho_w)g}{4D}\right)^{(1/4)} \tag{8}$$

The limits of integration $a$ and $b$ in (7) depend on the distances to the left and right edges, respectively, of the nth rectangular load component from the point at which the flexure is being calculated. The total flexure is the sum of the contributions from each narrow rectangular load component. The empty basin topography $h_o(x)$ is obtained by subtracting the flexure $w(x)$ from the loaded topography $h(x)$. For each scenario, we compare the positions of the shorelines before and after unloading in each case, and measure the resulting vertical shoreline offset ($\Delta z$ in Figure 2b).

[11] Figure 3 shows the loaded basin topography and resulting flexure for each of the three basin scenarios. There are several salient points to note. Firstly, because the load is asymmetric, the flexural bulges to the left and right of the load are not the same height or width, but represent the cumulative response to the total load on each side of the basin axis. The largest value of $\lambda$, associated with the largest elastic thickness, corresponds to the widest flexural bulge in each case. Secondly, for the smallest basin, the water load is largely supported by the rigidity of the lithosphere, particularly for the largest value of elastic thickness. The maximum flexural deflection for this basin is small relative to the maximum basin depth. By contrast, the load for the largest basin is largely compensated by the mantle; in this case, the load is so large relative to the elastic thickness that the lithosphere cannot support the load through rigidity alone.

[12] Figure 4 shows the loaded and unloaded basin topography for each of the three basin scenarios. Each plot highlights the past (loaded) and present (unloaded) locations of shoreline features, and shows the vertical offsets of the shorelines on the left side relative to those on the right side of each basin for each value of $T_e$. The vertical shoreline offset ($\Delta z$) ranges from 2 m for the smallest basin with $T_e$ = 100 km to >500 m for the largest basin with $T_e$ = 100 km. Interestingly, the relationships between basin size and $T_e$ and $\Delta z$ are not monotonic; for instance, $\Delta z$ for a $T_e$ of 50 km is larger for the 500 km basin than for the 2500 km basin. This is because shoreline elevation following unloading depends on the shape of the basin flexure. The location, magnitude, and width of the flexural bulges determine how much any given point in the basin rebounds following unloading; the height and shape of the flexural bulges, in turn, are functions of the relationship between the lithospheric rigidity (controlled by $T_e$) and the shape and size of the load. Because our hypothetical basins are highly asymmetric, the position of the left-hand shoreline with respect to the corresponding flexural bulge is much different from that of the right hand shoreline. Similarly asymmetric basin profiles would have existed on Mars in association with, e.g., hypothetical large water bodies in the northern plains. For example, regions *A* and *B* in Figure 1 are adjacent to broad shallow shelves, whereas regions *C* and *D* are adjacent to much steeper transitions in depth, and are proximal to the regions of greatest water loading. Regions *A* and *B* are analogous to the shallow right-hand areas of our profiles (Figures 3 and 4), whereas regions *C* and *D* are analogous to the left-hand steep-slope areas. Therefore we would expect regions *C* and *D* to experience greater rebound following unloading than regions *A* and *B*. Though the simplified two-dimensional calculations given above do not strictly apply to a basin of this size and complexity, they nevertheless suggest that the lithosphere in such a region could experience an asymmetric flexural response that is qualitatively similar to the results shown in Figures 3 and 4.

[13] An equally dramatic effect to the vertical shoreline offset is the change in calculated shoreline position relative to proximal basin divides or other near-basin topography. For instance, for the 500 km basin, rebound on the shallow (right) side of the basin results in a subtle topographic divide between the two shorelines for most values of $T_e$ (Figure 4). It is likely that a search for shoreline features associated with the empty basins in these cases would focus on regions to the left of these divides, when in fact, if shoreline features were preserved, they would reside to the right of the divides, which did not exist when the basins were loaded. This result highlights the importance of considering the effects of water loading on both the form of the main basin and that of surrounding regions in determining the likely geographic positions of preserved nearshore features.

[14] To calculate the actual shoreline offsets for particular Martian basins, a suitable approach would be to treat Mars' lithosphere as an elastic shell surrounding a fluid interior [e.g., *Brotchie and Silvester*, 1969; *Turcotte et al.*, 1981;

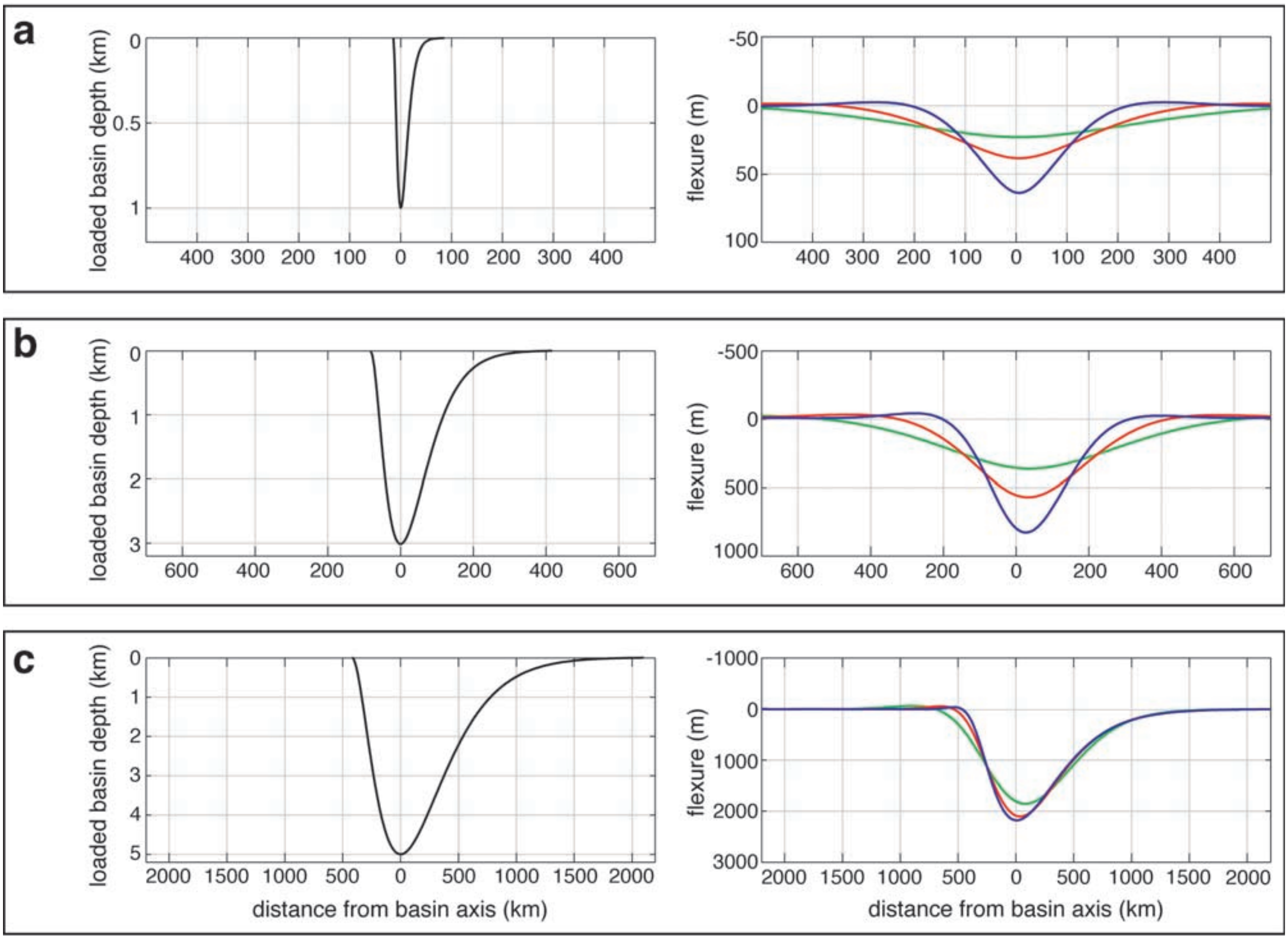


**Figure 3.** Plots of loaded basin depth (left) and associated flexure (right) vs. distance from basin axis for (a) 100 km wide basin, maximum depth 1 km; (b) 500 km wide basin, maximum depth 3 km; and (c) 2500 km wide basin, maximum depth 5 km. Blue, red, and green curves correspond to $T_e$ = 25, 50, and 100 km, respectively.

*Banerdt et al.*, 1982; *Banerdt*, 1986; *Johnson et al.*, 2000]. Such a model would need to consider support of loads via bending stresses arising from the rigidity of the shell, membrane stresses in the shell, and buoyant stresses in the mantle arising in response to the applied load. At long wavelengths, viscous relaxation of topography becomes important [e.g., *Zhong and Zuber*, 2000] due to flow of mantle or lower crustal material (for cases in which the base of the elastic layer resides within the crust); thus elastic shell models can be augmented by accompanying estimates of the characteristic timescales over which mechanical equilibrium is achieved. Construction of such a model for actual Martian basins is beyond the scope of this paper, but represents an avenue for further investigation.

## 5. Discussion

[15] Identification and evaluation of possible water-marginal features associated with large liquid or solid water bodies on Mars are typically made for candidate basins on the basis of geomorphological or textural transitions that occur at roughly uniform elevations. On the basis of the considerations examined above, however, it would appear that for large basins associated with irregular bathymetries or distinct crustal inhomogeneities, it may be necessary to, in addition to the determination of the effects of other influences such as tectonism and changes in the nature of geological loading and the thermal state of the lithosphere [e.g., *Clifford and Parker*, 2001; *Dohm et al.*, 2001; *Tanaka et al.*, 2001; *Ruiz*, 2003], consider the possible influence of the proposed water bodies themselves on the past form of the basins. For such basins, the terrestrial record suggests that modern horizontality of ancient Martian water-marginal features of common age should not necessarily be assumed.

[16] If large water bodies once existed at the surface of Mars, most or all individual water-marginal features such as shoreline terraces and beach ridges, having little topographic relief and in many cases being composed of unconsolidated materials, might be expected to have succumbed over geological timescales to the effects of erosion, burial, and in situ reworking of materials, and thus may not be preserved or distinguishable from other features at the surface of the planet today [e.g., *Carr*, 2001]. Even large morainal or other ice-marginal features that could have been associated with the growth or flow of large frozen water bodies might not be expected to have remained to the present day as obvious features exposed at the surface, particularly if such features were formed early in the

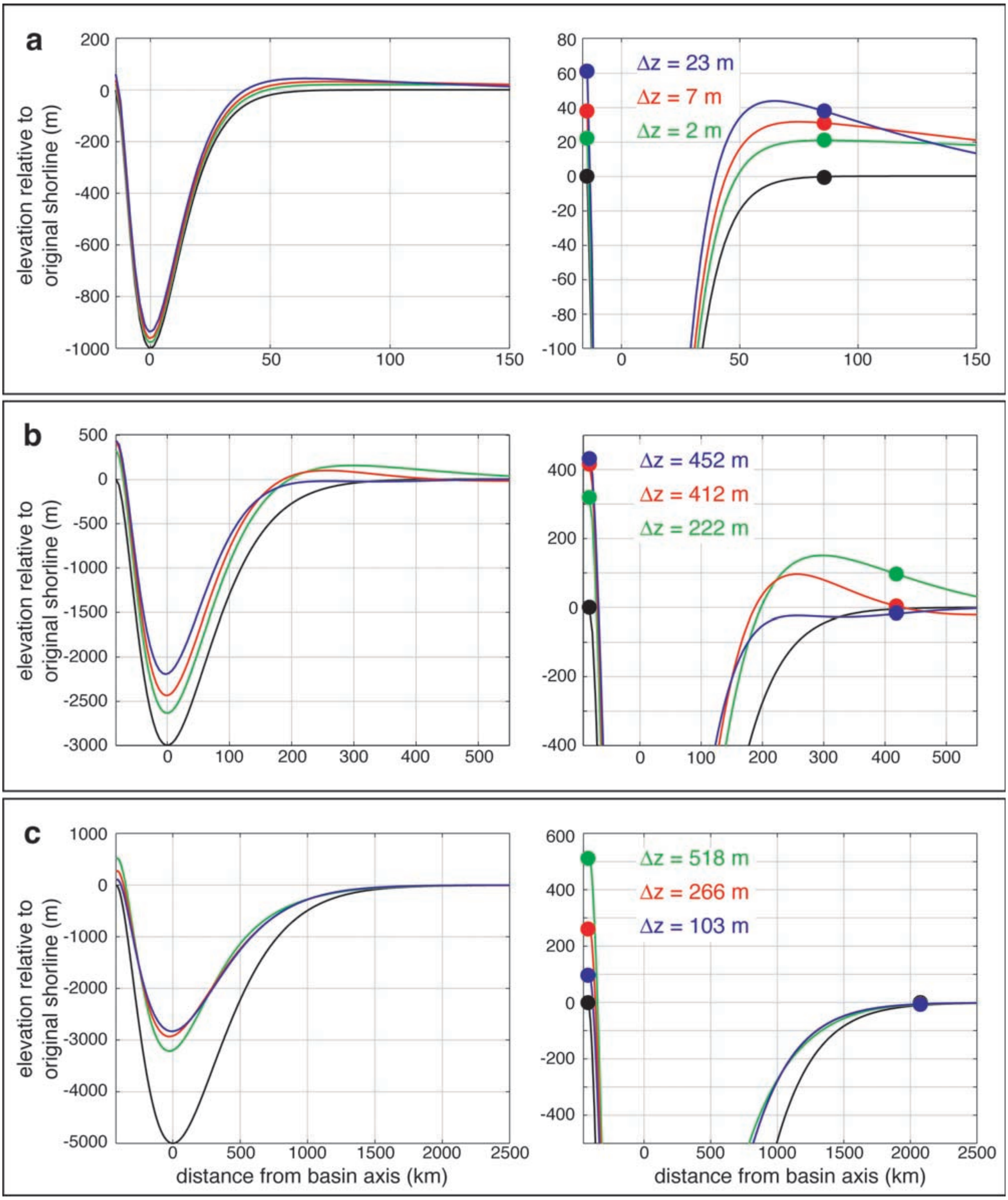


**Figure 4.** Plots of loaded and unloaded topography for (a) 100 km wide basin; (b) 500 km wide basin; and (c) 2500 km wide basin. Black lines represent loaded topography; blue, red, and green curves are calculated unloaded topography for $T_e$ = 25, 50, and 100 km, respectively. Plots on the right show enlargements of the shoreline regions; filled circles denote locations of shoreline features before and after unloading, and the vertical offset $\Delta z$ associated with each shoreline is labeled.

history of Mars. It is nevertheless possible that analysis of the positions and elevations of features associated with tributaries and outlets, in combination with analysis of the extent of well-preserved sedimentary units thought to have been deposited in large water bodies, could form the basis for the coarse definition of deformed ancient water planes. The future identification of strong candidates for segments of nearshore constructional and mass wasting features could contribute greatly to the refinement of such reconstructions.

[17] In cases on Mars where widespread preservation of ancient water-marginal features might have taken place, what should the expected nearshore signature be for basins whose geometries have been differentially influenced by water loading and unloading? Although the effects of late-Quaternary differential water loading noted in the Persian Gulf [*Lambeck*, 1996] are compelling, their magnitudes and significance are not easily extrapolated to larger basins of the Earth's oceans. The Earth has not lost its oceans, and thus a terrestrial analog directly comparable to the Martian basins is not available. Nevertheless, substantial clues are present in the terrestrial geological record. Glacial Lake Agassiz, the largest lake in North America during the last deglaciation, with a maximum volume that may have exceeded 160,000 km$^3$ [*Leverington et al.*, 2002; *Clarke et al.*, 2003], formed and evolved in a region of continuing differential glacio-isostatic rebound, intimately tying the form and dynamics of the lake to this rebound. The strandline record of Lake Agassiz reflects the positions of ancient shorelines, and corresponding deformed paleo-water-planes today rise toward the northeast, toward zones of maximum past loading [*Upham*, 1895; *Leverett*, 1932; *Johnston*, 1946; *Elson*, 1967; *Teller and Thorleifson*, 1983]. Preserved shoreline segments of different ages are nested inside one another, and having been influenced by different periods and magnitudes of rebound, each set of strandline features of a given age has a unique geometry. Although the root processes, lengths of time, and geographical extents associated with Lake Agassiz and hypothetical Martian oceans are distinctly different (with Lake Agassiz having existed over a period of only 5000 years, with a volume less than 1% of that of the largest hypothesized Martian oceans, and with its differential rebound dominated by glacio-isostatic rather than hydro-isostatic processes), it is nevertheless evident that the differential loading and unloading effects of a large Martian water body may similarly influence the spatial properties with which associated common-aged water-marginal features such as tributaries, outlets, and beaches are formed and preserved.

[18] A wide variety of hypotheses have been proposed to explain how large water bodies might have developed and disappeared on the surface of Mars. For example, it has been suggested that large liquid water bodies at the surface of the planet could have existed throughout the first billion years of Mars' history, maintained under warm conditions by a global groundwater system recharged by precipitation, or under cold conditions by ice cover and a groundwater system recharged by basal melting of ice-rich polar deposits; according to this model, these water bodies would have frozen at the end of the Noachian, and most of the water would have been transferred by sublimation and condensation to the poles or by vapor diffusion to the extensive subsurface cryosphere [*Clifford and Parker*, 2001; see also, e.g., *Carr*, 2002]. Another model involves periodic growth and diminution of large oceans and glaciers over much of the history of Mars, with multiple past transfers between surface water bodies and subsurface repositories driven by factors such as changes in the mantle heat flux, and taking place in concert with major climatic cycles [e.g., *Baker et al.*, 1991, 2000; *Baker*, 2001; *Fairén et al.*, 2003]. Other models propose that large lakes and oceans did not form at all; for example, it has been suggested that the largest surface water bodies on Mars may have taken the form of transient ice sheets of relatively limited extent, formed by the repeated extrusion of water from confined aquifers under exclusively cold conditions [*Gaidos and Marion*, 2003]. It remains uncertain if any presently hypothesized hydrological cycles actually functioned on Mars, but on the basis of the discussion in the preceding section, it is apparent that the manner in which large contemporaneous transfers of water mass may have taken place could have influenced the nature of the effects of water loading. For example, mass transfer between adjacent regions (such as from a large northern ocean to high-latitude regions) could have complicated the effects of water loading on Mars by combining in some locations the loading signal of one nearby region with the unloading signal of another region. It is furthermore apparent that the rate at which mass transfers may have occurred would also have been an important factor in determining the magnitude and nature of the effects of water loading. It has been suggested that large surface water bodies might have formed at geologically rapid rates through, e.g., catastrophic release of groundwater [e.g., *Baker et al.*, 1991; *Zuber et al.*, 2000], while much longer rates of water-body formation tied to long-term climate change are also possible. The rate at which water mass might move from a frozen lake or glacier to the poles and subsurface reservoirs via sublimation would depend on such factors as water-body size, latitude and climate conditions, ground thermal flux, sediment cover, and the surface temperature of the ice [e.g., *Farmer and Doms*, 1979; *Carr*, 1990, 1996; *Clifford*, 1993]; a large ancient water body located at latitudes poleward of 40° and protected by a cover of material several meters thick could theoretically persist until the present time, whereas a fully-exposed 1 km-thick ice mass could sublimate under present Martian conditions in roughly $10^6$ years [e.g., *Carr*, 1990, 1996; *Gaidos and Marion*, 2003].

## 6. Conclusions

[19] With considerable depths over large areas, hypothesized ancient Martian water bodies would have represented massive surficial loads. If these water bodies indeed existed, the irregular form of some Martian basins, as well as inhomogeneities in crustal and mantle properties, would have caused the magnitudes of lithospheric deflections related to loading and unloading to vary spatially. Terrestrial studies have found that spatial variation in the displacement effects of oceanic loading and unloading can exceed 10% of the magnitude of vertical changes in water level (and when occurring in concert with other mass-transfer processes can exceed 200% of this magnitude). Simple models point to significant lithospheric deflection due to loading by proposed liquid and solid water bodies. It is likely that shoreline areas proximal to deep regions will be depressed (and subsequently rebound) to a significantly greater extent than shoreline areas located near shallow regions. If so, this suggests that Martian water-marginal features, of common age and formed in association with a large ancient water body, could conceivably vary in elevation today by up to hundreds of meters as a result of differential rebound. The

modern horizontality of sets of possible strandline features should therefore not be considered a necessary precondition for their interpretation as having formed in association with a large water body. The terrestrial record suggests that the geographic pattern of strandline features should reflect the changing nature of loading through time.

[20] **Acknowledgments.** We thank two anonymous referees for their helpful reviews. We also thank Bruce Campbell and Andrew Dombard for numerous detailed discussions regarding earlier versions of the manuscript. This work was supported by the National Air and Space Museum, Smithsonian Institution.

## References

Andrews, J. T. (1970), A geomorphological study of post-glacial uplift with particular reference to Arctic Canada, *Spec. Publ. 2*, Inst. of Br. Geogr., London.
Andrews, J. T. (1982), On the reconstruction of Pleistocene ice sheets: A review, *Quat. Sci. Rev.*, *1*, 1–30.
Andrews, J. T. (1989), Postglacial emergence and submergence, in *Quaternary Geology of Canada and Greenland*, *Geol. Can. 1*, edited by R. J. Fulton, pp. 546–562, Geol. Surv. of Can.
Baker, V. R. (1982), *The Channels of Mars*, Univ. of Tex. Press, Austin.
Baker, V. R. (2001), Water and the Martian landscape, *Nature*, *412*, 228–236.
Baker, V. R., and D. J. Milton (1974), Erosion by catastrophic floods on Mars and Earth, *Icarus*, *23*, 27–41.
Baker, V. R., R. G. Strom, V. C. Gulick, J. S. Kargel, G. Komatsu, and V. S. Kale (1991), Ancient oceans, ice sheets and the hydrological cycle on Mars, *Nature*, *352*, 589–594.
Baker, V. R., R. G. Strom, J. M. Dohm, V. C. Gulick, J. S. Kargel, G. Komatsu, G. G. Ori, and J. W. Rice Jr. (2000), Mars: Oceanus Borealis, ancient glaciers, and the MEGAOUTFLO hypothesis, *Lunar Planet. Sci.*, *XXXI*, abstract 1863.
Bandfield, J. L., T. D. Glotch, and P. R. Christensen (2003), Spectroscopic identification of carbonate minerals in Martian dust, *Science*, *301*, 1084–1087.
Banerdt, W. B. (1986), Support of long-wavelength loads on Venus and implications for internal structure, *J. Geophys. Res.*, *91*(B1), 403–419.
Banerdt, W. B., R. J. Phillips, N. H. Sleep, and R. S. Saunders (1982), Thick shell tectonics on one-plate planets: Applications to Mars, *J. Geophys. Res.*, *87*(B12), 9723–9733.
Brotchie, J. F., and R. Silvester (1969), On crustal flexure, *J. Geophys. Res.*, *74*, 5240–5252.
Cabrol, N. A., and E. A. Grin (1999), Distribution, classification, and ages of Martian impact crater lakes, *Icarus*, *142*, 160–172.
Cabrol, N. A., and E. A. Grin (2001), The evolution of lacustrine environments on Mars: Is Mars only hydrologically dormant?, *Icarus*, *149*, 291–328.
Cabrol, N. A., E. A. Grin, H. E. Newsom, R. Landheim, and C. P. McKay (1999), Hydrogeologic evolution of Gale crater and its relevance to the exobiological exploration of Mars, *Icarus*, *139*, 235–245.
Carr, M. H. (1990), D/H on Mars: Effects of floods, volcanism, impacts and polar processes, *Icarus*, *87*, 210–227.
Carr, M. H. (1996), *Water on Mars*, Oxford Univ. Press, New York.
Carr, M. H. (2001), Mars Global Surveyor observations of Martian fretted terrain, *J. Geophys. Res.*, *106*(E10), 23,571–23,593.
Carr, M. H. (2002), Elevations of water-worn features on Mars: Implications for circulation of groundwater, *J. Geophys. Res.*, *107*(E12), 5131, doi:10.1029/2002JE001845.
Carr, M. H., and J. W. Head III (2003), Oceans on Mars: An assessment of the observational evidence and possible fate, *J. Geophys. Res.*, *108*(E5), 5042, doi:10.1029/2002JE001963.
Chapman, M. G. (1994), Evidence, age, and thickness of a frozen paleolake in Utopia Planitia, Mars, *Icarus*, *109*, 393–406.
Chappell, J. (1974), Late Quaternary glacio- and hydro-isostasy, on a layered Earth, *Quat. Res.*, *4*, 429–440.
Christensen, P. R., et al. (2000), Detection of crystalline hematitemineralization on Mars by the Thermal Emission Spectrometer: Evidence for near-surface water, *J. Geophys. Res.*, *105*(E4), 9623–9642.
Christensen, P. R., R. V. Morris, M. D. Lane, J. L. Bandfield, and M. C. Malin (2001), Global mapping of Martian hematite mineral deposits: Remnants of water-driven processes on early Mars, *J. Geophys. Res.*, *106*(E10), 23,873–23,885.
Clark, J. A., W. E. Farrell, and W. R. Peltier (1978), Global changes in postglacial sea level: A numerical calculation, *Quat. Res.*, *9*, 265–287.
Clarke, G., D. Leverington, J. Teller, and A. Dyke (2003), Superlakes, megafloods, and abrupt climate change, *Science*, *301*, 922–923.
Clifford, S. M. (1993), A model for the hydrologic and climatic behavior of water on Mars, *J. Geophys. Res.*, *98*(E6), 10,973–11,016.
Clifford, S. M., and T. J. Parker (2001), The evolution of the Martian hydrosphere: Implications for the fate of a primordial ocean and the current state of the northern plains, *Icarus*, *154*, 40–79.
Craddock, R. A., and A. D. Howard (2002), The case for rainfall on a warm, wet early Mars, *J. Geophys. Res.*, *107*(E11), 5111, doi:10.1029/2001JE001505.
De Hon, R. A. (1992), Martian lake basins and lacustrine plains, *Earth Moon Planets*, *56*, 95–122.
Dohm, J. M., J. C. Ferris, V. R. Baker, R. C. Anderson, T. M. Hare, R. G. Strom, N. G. Barlow, K. L. Tanaka, J. E. Klemaszewski, and D. H. Scott (2001), Ancient drainage basin of the Tharsis region, Mars: Potential source for outflow channel systems and putative oceans or paleolakes, *J. Geophys. Res.*, *106*(E12), 32,943–32,958.
Elson, J. A. (1967), Geology of glacial Lake Agassiz, in *Life, Land, and Water*, edited by W. J. Mayer-Oakes, pp. 36–95, Univ. of Manitoba Press, Winnipeg, Manitoba, Canada.
Fairén, A. G., J. M. Dohm, V. R. Baker, M. A. de Pablo, J. Ruiz, J. C. Ferris, and R. C. Anderson (2003), Episodic flood inundations of the northern plains of Mars, *Icarus*, *165*, 53–67.
Farmer, C. B., and P. E. Doms (1979), Global seasonal variation of water vapor on Mars and the implications for permafrost, *J. Geophys. Res.*, *84*(B6), 2881–2888.
Farrell, W. E., and J. A. Clark (1976), On postglacial sea level, *Geophys. J. R. Astron. Soc.*, *46*, 647–667.
Forsythe, R. D., and J. R. Blackwelder (1998), Closed drainage crater basins of the Martian Highlands: Constraints on the early Martian hydrologic cycle, *J. Geophys. Res.*, *103*(E13), 31,421–31,431.
Forsythe, R. D., and J. R. Zimbelman (1995), A case for ancient evaporite basins on Mars, *J. Geophys. Res.*, *100*(E3), 5553–5563.
Gaidos, E., and G. Marion (2003), Geological and geochemical legacy of a cold early Mars, *J. Geophys. Res.*, *108*(E6), 5055, doi:10.1029/2002JE002000.
Goldspiel, J. M., and S. W. Squyres (1991), Ancient aqueous sedimentation on Mars, *Icarus*, *89*, 392–410.
Grin, E. A., and N. A. Cabrol (1997), Limnologic analysis of Gusev crater paleolake, Mars, *Icarus*, *130*, 461–474.
Head, J. W., III, M. Kreslavsky, H. Hiesinger, M. Ivanov, S. Pratt, and N. Seibert (1998), Oceans in the past history of Mars: Tests for their presence using Mars Orbiter Laser Altimeter (MOLA) data, *Geophys. Res. Lett.*, *25*, 4401–4404.
Head, J. W., III, H. Hiesinger, M. A. Ivanov, M. A. Kreslavsky, S. Pratt, and B. J. Thomson (1999), Possible ancient oceans on Mars: Evidence from Mars Orbiter Laser Altimeter data, *Science*, *286*, 2134–2137.
Hiesinger, H., and J. W. Head III (2000), Characteristics and origin of polygonal terrain in southern Utopia Planitia, Mars: Results from Mars Orbiter Laser Altimeter and Mars Orbiter Camera data, *J. Geophys. Res.*, *105*(E5), 11,999–12,022.
Hynek, B. M., and R. J. Phillips (2003), New data reveal mature, integrated drainage systems on Mars indicative of past precipitation, *Geology*, *31*, 757–760.
Irwin, R. P., III, T. A. Maxwell, A. D. Howard, R. A. Craddock, and D. W. Leverington (2002), A large paleolake basin at the head of Ma'adim Vallis, Mars, *Science*, *296*, 2209–2212.
Ivanov, M. A., and J. W. Head (2001), Chryse Planitia, Mars: Topographic configuration, outflow channel continuity and sequence, and tests for hypothesized ancient bodies of water using MOLA, *J. Geophys. Res.*, *106*(E2), 3275–3295.
Jakosky, B. M., and R. J. Phillips (2001), Mars' volatile and climate history, *Nature*, *412*, 237–244.
Johnson, C. L., S. C. Solomon, J. W. Head, R. J. Phillips, D. E. Smith, and M. T. Zuber (2000), Lithospheric loading by the northern polar cap on Mars, *Icarus*, *144*, 313–328.
Johnston, W. A. (1946), *Glacial Lake Agassiz, With Special Reference to the Mode of Deformation of the Beaches*, *Bull. Geol. Surv. Can.*, *7*, 20 pp.
Kreslavsky, M. A., and J. W. Head (2002), Fate of outflow channel effluents in the northern lowlands of Mars: The Vastitas Borealis Formation as a sublimation residue from frozen ponded bodies of water, *J. Geophys. Res.*, *107*(E12), 5121, doi:10.1029/2001JE001831.
Lambeck, K. (1996), Shoreline reconstructions for the Persian Gulf since the last glacial maximum, *Earth Planet. Sci. Lett.*, *142*, 43–57.
Lambeck, K., T. M. Esat, and E.-K. Potter (2002), Links between climate and sea levels for the past three million years, *Nature*, *419*, 199–206.
Leverett, F. (1932), Quaternary geology of Minnesota and parts of adjacent states, *U.S. Geol. Surv. Prof. Pap.*, *161*, 148 pp.

Leverington, D. W., J. D. Mann, and J. T. Teller (2002), Changes in the bathymetry and volume of glacial Lake Agassiz between 9200 and 7700 $^{14}C$ yr B. P., *Quat. Res.*, *57*, 244–252.
Lucchitta, B. K. (1982), Lakes or playas in Valles Marineris, in *Reports of Planetary Geology Program-1982*, *NASA Tech. Memo. 85127*, pp. 233–234.
Lucchitta, B. K., H. M. Ferguson, and C. Summers (1986), Sedimentary deposits in the northern lowland plains, Mars, *J. Geophys. Res.*, *91*, 166–174.
Malin, M. C., and K. S. Edgett (1999), Oceans or seas in the Martian northern lowlands: High resolution imaging tests of proposed coastlines, *Geophys. Res. Lett.*, *26*, 3049–3052.
Masursky, H. (1973), An overview of geologic results from Mariner 9, *J. Geophys. Res.*, *78*, 4037–4047.
Masursky, H., J. V. Boyce, A. L. Dial, G. G. Schaber, and M. E. Strobell (1977), Classification and time of formation of Martian channels based on Viking data, *J. Geophys. Res.*, *82*, 4016–4037.
McCauley, J. F., M. H. Carr, J. A. Cutts, W. K. Hartmann, H. Masursky, D. J. Milton, R. P. Sharp, and D. E. Wilhelms (1972), Preliminary Mariner 9 report on the geology of Mars, *Icarus*, *17*, 289–327.
McGill, G. E. (2001), The Utopia Basin revisited: Regional slope and shorelines from MOLA profiles, *Geophys. Res. Lett.*, *28*, 411–414.
McGovern, P. J., S. C. Solomon, D. E. Smith, M. T. Zuber, M. Simons, M. A. Wieczorek, R. J. Phillips, G. A. Neumann, O. Aharonson, and J. W. Head (2002), Localized gravity/topography admittance and correlation spectra on Mars: Implications for regional and global evolution, *J. Geophys. Res.*, *107*(E12), 5136, doi:10.1029/2002JE001854.
McKenzie, D., D. N. Barnett, and D. N. Yuan (2002), The relationship between Martian gravity and topography, *Earth Planet. Sci. Lett.*, *195*, 1–16.
Milkovich, S. M., J. W. Head III, and S. Pratt (2002), Meltback of Hesperian-aged ice-rich deposits near the south pole of Mars: Evidence for drainage channels and lakes, *J. Geophys. Res.*, *107*(E6), 5043, doi:10.1029/2001JE001802.
Milton, D. J. (1973), Water and processes of degradation in the Martian landscape, *J. Geophys. Res.*, *78*, 4037–4047.
Moore, J. M., and D. E. Wilhelms (2001), Hellas as a possible site of ancient ice-covered lakes on Mars, *Icarus*, *154*, 258–276.
Nedell, S. S., S. W. Squyres, and D. W. Andersen (1987), Origin and evolution of the layered deposits in the Valles Marineris, Mars, *Icarus*, *70*, 409–441.
Newsom, H. E., G. E. Brittelle, C. A. Hibbitts, L. J. Crossey, and A. M. Kudo (1996), Impact crater lakes on Mars, *J. Geophys. Res.*, *101*(E6), 14,951–14,955.
Ori, G. G., and C. Mosangini (1998), Complex depositional systems in Hydraotes Chaos, Mars: An example of sedimentary process interactions in the Martian hydrological cycle, *J. Geophys. Res.*, *103*(E10), 22,713–22,723.
Ori, G. G., L. Marinangeli, and A. Baliva (2000), Terraces and Gilbert-type deltas in crater lakes in Ismenius Lacus and Memnonia (Mars), *J. Geophys. Res.*, *105*(E7), 17,629–17,641.
Parker, T. J., R. S. Saunders, and D. M. Schneeberger (1989), Transitional morphology in the west Deuteronilus Mensae region of Mars: Implications for modification of the lowland/upland boundary, *Icarus*, *82*, 111–145.
Parker, T. J., D. S. Gorcine, R. S. Saunders, D. C. Pieri, and D. M. Schneeberger (1993), Coastal geomorphology of the Martian northern plains, *J. Geophys. Res.*, *98*(E6), 11,061–11,078.
Parker, T. J., S. M. Clifford, and W. B. Banerdt (2000), Argyre Planitia and the Mars global hydrologic cycle, *Lunar Planet. Sci.*, [CD-ROM], *XXXI*, abstract 2033.
Parker, T. J., J. A. Grant, F. S. Anderson, and W. B. Banerdt (2003), From the south pole to the northern plains: The Argyre Planitia story, in *Sixth International Conference on Mars* [CD-ROM], abstract 3274, Lunar and Planet. Inst., Houston, Tex.
Peltier, W. R. (1974), The impulse response of a Maxwell Earth, *Rev. Geophys.*, *12*, 649–669.
Peltier, W. R. (1985), The LAGEOS constraint on deep mantle viscosity: Results from a new normal mode method for the inversion of viscoelastic relaxation spectra, *J. Geophys. Res.*, *90*, 9411–9421.
Peltier, W. R. (1994), Ice age paleotopography, *Science*, *265*, 195–201.
Peltier, W. R., and J. T. Andrews (1976), Glacial-isostatic adjustment: I. The forward problem, *Geophys. J. R. Astron. Soc.*, *46*, 605–646.
Pollack, J. B., J. F. Kasting, S. M. Richardson, and K. Poliakoff (1987), The case for a warm, wet climate on early Mars, *Icarus*, *71*, 203–224.
Rathbun, J. A., and S. W. Squyres (2002), Hydrothermal systems associated with Martian impact craters, *Icarus*, *157*, 362–372.
Ruiz, J. (2003), Amplitude of heat flow variations on Mars from possible shoreline topography, *J. Geophys. Res.*, *108*(E11), 5122, doi:10.1029/2003JE002084.
Sagan, C., O. B. Toon, and P. J. Gierasch (1973), Climate change on Mars, *Science*, *181*, 1045–1049.
Sharp, R. P., and M. C. Malin (1975), Channels on Mars, *Geol. Soc. Am. Bull.*, *86*, 593–609.
Smith, D. E. (2003), Mars Orbiter Laser Altimeter (MOLA) Experiment Gridded Data Record (MEGDR), *MEG064*, NASA Planet. Data Syst., Washington Univ., St. Louis, Mo.
Tanaka, K. L., W. B. Banerdt, J. S. Kargel, and N. Hoffman (2001), Huge, $CO_2$-charged debris-flow deposit and tectonic sagging in the northern plains of Mars, *Geology*, *29*, 427–430.
Tanaka, K. L., J. A. Skinner Jr., T. M. Hare, T. Joyal, and A. Wenker (2003), Resurfacing history of the northern plains of Mars based on geologic mapping of Mars Global Surveyor data, *J. Geophys. Res.*, *108*(E4), 8043, doi:10.1029/2002JE001908.
Teller, J. T., and L. H. Thorleifson (1983), The Lake Agassiz-Lake Superior connection, in *Glacial Lake Agassiz*, edited by J. T. Teller and L. Clayton, *Geol. Assoc. Can. Spec. Pap.*, *26*, 261–290.
Thomson, B. J., and J. W. Head III (2001), Utopia basin, Mars: Characterization of topography and morphology and assessment of the origin and evolution of basin internal structure, *J. Geophys. Res.*, *106*(E10), 23,209–23,230.
Turcotte, D. L., and G. Schubert (2002), *Geodynamics*, Cambridge Univ. Press, New York.
Turcotte, D. L., R. J. Willemann, W. F. Haxby, and J. Norberry (1981), Role of membrane stresses in the support of planetary topography, *J. Geophys. Res.*, *86*(B5), 3951–3959.
Upham, W. (1895), The Glacial Lake Agassiz, *U.S. Geol. Surv. Monogr.*, *25*.
Walcott, R. I. (1970), Isostatic response to loading of the crust in Canada, *Can. J. Earth Sci.*, *7*, 716–727.
Walcott, R. I. (1972a), Past sea levels, eustasy and deformation of the Earth, *Quat. Res.*, *2*, 1–14.
Walcott, R. I. (1972b), Late Quaternary vertical movements in eastern North America: Quantitative evidence of glacio-isostatic rebound, *Rev. Geophys.*, *10*, 849–884.
Watts, A. B. (2001), *Isostasy and Flexure of the Lithosphere*, Cambridge Univ. Press, New York.
Zhong, S., and M. T. Zuber (2000), Long-wavelength topographic relaxation for self-gravitating planets and implications for the time-dependent compensation of surface topography, *J. Geophys. Res.*, *105*(E2), 4153–4164.
Zuber, M. T., et al. (2000), Internal structure and early thermal evolution of Mars from Mars Global Surveyor topography and gravity, *Science*, *287*, 1788–1793.

R. R. Ghent and D. W. Leverington, Center for Earth and Planetary Studies, National Air and Space Museum, Smithsonian Institution, P.O. Box 37012, Room 3774, MRC 0315, Washington, DC 20560-0315, USA. (leveringtond@nasm.si.edu)