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Uniformitarian plume tectonics: The post-Archean Earth and Mars

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ABSTRACT

Tectonism and magmatism related to two well-documented mantle plumes, the 1267 Ma Mackenzie and the 17 Ma Yellowstone, are reviewed and discussed. While the Precambrian plume effects inform us of deep-crustal processes, the Tertiary plume effects inform us of surface and subsurface processes. Each of them emphasizes different aspects of plume tectonics, the combination of which is helpful to elucidate extraterrestrial plume tectonics. I show that, despite probable differences in planetary thermal evolution, the tectonic activity at the Tharsis volcanic province on Mars—the largest volcanic-tectonic center of the Solar System—is consistent with the plume-tectonics inferences from the two terrestrial examples dealt with herein. The combination of all three examples helps refine comprehensive multiplanetary plume-tectonics models.

INTRODUCTION

Ancient hotspot outcrops allow scientists to retrieve the influence of mantle plumes in the deep brittle crust (e.g., 7 km in the Mackenzie case), and present hotspots provide information about surface and near-surface volcanic and tectonic events. A consequence is that some of the key advances in plume tectonics (defined as "the entire process of thermal and structural reworking above mantle plumes," Hill et al., 1992, p. 187) during the 1990s has come from assembling deep and surficial observations at old and recent hotspots in a geodynamic perspective, which has led to coherent plume-tectonics models (e.g., Fahrig, 1987; LeCheminant and Heaman, 1989; Richards et al., 1989; White and McKenzie, 1989, 1995; Griffiths and Campbell, 1991; Hill, 1991; Saunders et al., 1992; Courtillot et al., 1999).

The tectonics of Mars and Venus appears to have been much more influenced by mantle plumes than has the tectonics of Earth. On Mars, most tectonic activity observed is geographically correlated with the gigantic Tharsis volcanic province, which covers one-fourth of the planetary surface (Frey, 1979; Plescia and Saunders, 1982; Scott and Tanaka, 1986; Mège and Masson, 1996a). A number of the remaining tectonic structures are associated with the smaller Elysium magma center (Comer et al., 1985; Breuer et al., 1998). Evolution of these two regions has probably been controlled by mantle upwellings (e.g., Spohn et al., 1998). On Venus, mantle plumes are involved in the formation of volcanic rises 1000–2500 km in diameter and associated tectonic structures (e.g., Stofan et al., 1995), intensely deformed volcanic plateaus of similar dimensions (e.g., Ghent and Hansen, 1999), and coronae—ovoid structures a few hundreds to 2000 km in diameter surrounded by intensely deformed annuli (e.g., Stofan et al., 1992). Some of these volcanic provinces are associated with failed rifts (e.g., Magee-Roberts et al., 1992).

All these extraterrestrial examples must be accounted for when dealing with comprehensive plume-tectonics models. In the present paper, I focus on comparing plume tectonics associated with one Proterozoic mantle plume, one currently active plume, and one extraterrestrial plume, each one selected for being especially well documented (Table 1). The ancient mantle plume is the 1267 Ma Mackenzie plume (Fig. 1). The example chosen of a currently active mantle plume is the 17 Ma Yellowstone plume (Fig. 2); emphasis in this paper is on the Columbia Plateau basalts, where the effects of tectonic activity dating

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TABLE 1. PLUME-TECTONICS FEATURES AT THE THREE IGNEOUS PROVINCES DISCUSSED IN THE TEXT

	Mackenzie	Yellowstone*	Tharsis
Geodynamic setting			
Extensional setting	Yes	Yes	Yes
Orientation of max regional principal stress	Northeast-southwest (10)	ENE-WSW (3)	NNE-SSW to NNW-SSE (1)
Plume extent			
Gravity anomaly	Bouguer: 80 to 100 mGal (2)	CP: Bouguer: -100 to -50 mGal (15) WSRP: Bouguer: -150 to -75 mGal (15)	Free air: ≥ 0 to 500 mGal (9)
Gravity high compared to surroundings	Bouguer: 50 to 100 mGal (15)	~200 mGal (15)	Overprinted by later plume events
Length of mafic dike swarm	$\geq$2600 km (8)	CP: $\geq$400 km (3) NNR: $\geq$500 km (3)	2 $\times$ ~2500 km (1)
Flood-basalt surface area	?	CP: 164000 km^2 (12)	10^6–10^7 km^2 (16)
Volume of magma erupted and intruded	230000 km^2 (2)	CP: 174000 km^2 (12)	?
Inferred plume diameter	$\geq$2000 km (8)	1000 km (6)	$\geq$2000 km (19)
Age and recurrence			
Age of first dike swarm and earliest eruptions	1267 $\pm$ 2 Ma (2)	17–14 Ma (3, 6, 14)	Within 3.8–3.1 Ga (1)
Recurring dike-swarm emplacement	Franklin ? 723 $+4/-2$ Ma (11)	Too young to state	Yes (cf. Fig. 4) (1)
Dike length/thickness ratio			
Dike thickness	$\geq$200 m, average = 30 m (4)	NNR: 3–250 m, mean $<$10 m (3) CP: 30 m (5)	Predicted to be $\geq$40–45 m (1)
Length/thickness ratio	5×10^4–6.7×10^4 (19)	1.3×10^4–5×10^4 (19)	5.5×10^4–6.3×10^4, theoretical (17)
Uplift and subsidence history			
Topographic uplift (max)	evidenced (2)	WSRP: initially $>$650 m (6) Y: currently 600 m (6)	~3 km? (20)
Thermal subsidence	Yes (2)	Yes (13)	Yes (1, 18)
Surface extension and compression			
Rift stage attained	Poseidon: oceanization (4)	NNR: continental (failed) (3)	Valles Marineris: failed (21,7)
Vertical throws of rift border faults	?	1.3–3.6 km (3)	2–10 km (20)
Wrinkle ridges in lava flows, normal to dikes	?	Yakima folds (22, 18)	yes (1)
Peripheral contractional belt	Perhaps (18)	Not reported	South Syria Planum ridge belt (18)

*CP—Columbia Plateau; NNR—North Nevada rift; WSRP—West Snake River Plain; Y—Yellowstone caldera.

Note: References are as follows: (1) Mège and Masson, 1996a; (2) LeCheminant and Heaman, 1989; (3) Zoback et al., 1994; (4) Fahrig, 1987; (5) S.P. Reidel, 2000, written communication; (6) Smith and Braile, 1994; (7) Mège and Masson, 1996b; (8) Ernst and Baragar, 1992; (9) Zuber et al., 2000; (10) Ernst et al., 1995; (11) Heaman et al., 1992; (12) Tolan et al., 1989; simultaneous eruptions at the Oregon Plateau with similar petrochemical characteristics produced an additional 65000 km^3 at most (Carlson and Hart, 1988); (13) Brott et al., 1981; (14) Hooper, 1997; (15) Committee for the Gravity Anomaly Map of North America, 1987; (16) after Watters, 1993; (17) Wilson and Head, 2000; (18) Mège and Ernst, this volume; (19) this study; (20) Smith et al., 1999; (21) Schultz, 1995; (22) Hooper and Conrey, 1989.

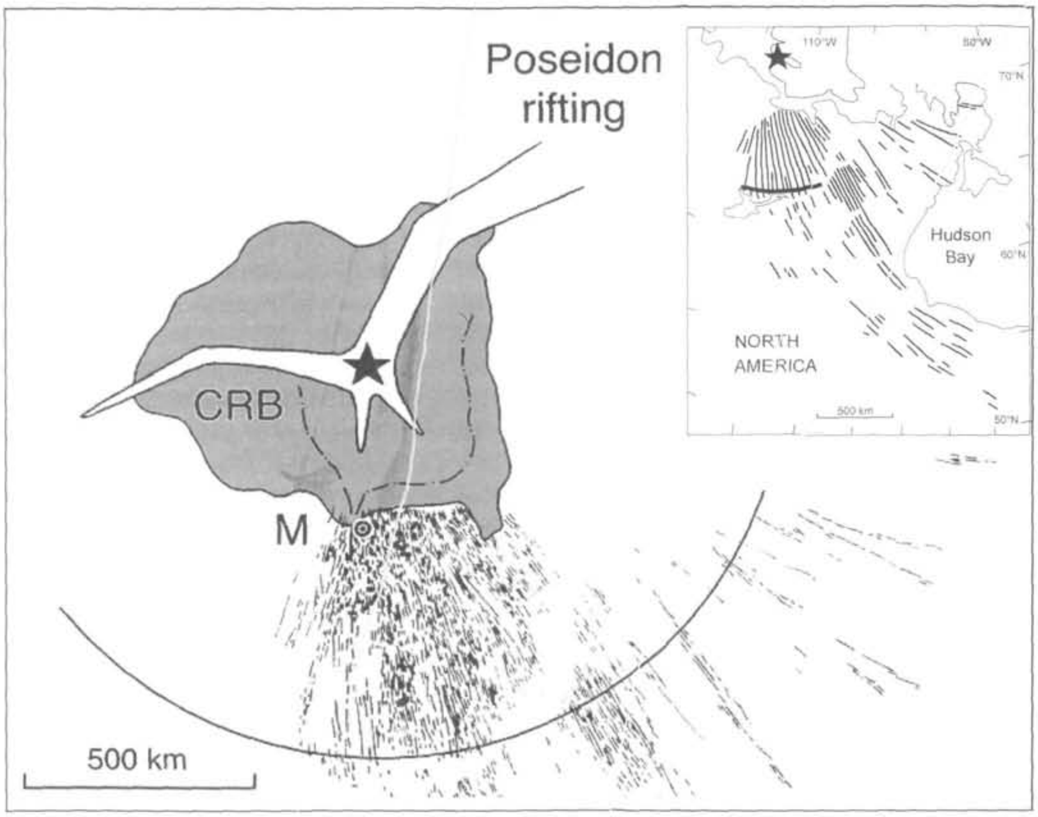

Figure 1. Tectonic and magmatic features associated with the Mackenzie igneous province. Star—the initial plume center relative to the lithosphere. Partial solid circle—estimate of the zone of plume influence on stress-field orientation. Dark lines—dikes; CRB—Coppermine River basalts; M—Muskox intrusion. Modified after LeCheminant and Heaman (1989) and Ernst et al. (1995).

back to the earliest plume events are preserved. The extraterrestrial plume is the Tharsis plume on Mars (Fig. 3). I show that despite the major differences that should exist between the thermal evolution of Earth and Mars, it is striking that plume-tectonics models established from terrestrial mantle plumes and illustrated by the Mackenzie and Yellowstone examples successfully account for a large part of the tectonic and volcanic history of the Tharsis volcanic province. In a companion paper (Mège and Ernst, this volume), we show how reciprocally, the extraterrestrial examples must be taken into account to come up with a comprehensive, multiplanetary plume-tectonics model, especially in regard to contractional tectonics.

PLUME TECTONICS AT REPRESENTATIVE TERRESTRIAL HOTSPOTS: THE MACKENZIE AND YELLOWSTONE PLUMES

In this section, I review and discuss early hotspot and plume characters in the Mackenzie igneous province and the Yellowstone hotspot, including evidence for plume origin, extent of the associated igneous province, and plume diameter. Then I go into more details and discuss timing of magmatic activity, dike length and thickness, topographic uplift and subsidence history, extensional tectonic structures, and contractional structures.

Evidence for plume origin

The Mackenzie igneous province (Fig. 1) includes a mafic dike swarm, a central magma chamber (the Muskox intrusion), and overlying flood-basalts (the Coppermine River basalts). Fahrig and Jones (1969) identified most of the swarm composing the Mackenzie mafic dike swarm. On the basis of the analogy with the North Atlantic rifting event (May, 1971), Halls (1982) suggested that some of the giant mafic dike swarms could have been associated with hotspots, including the Mackenzie dike swarm because of its fanning pattern converging toward the Muskox intrusion. Fahrig (1987) correlated the dike-swarm geometry to opening of a putative Poseidon Ocean in a mantle-plume context. In a key paper, LeCheminant and Heaman (1989) proposed a plume-tectonics model including uplift, subsidence, emplacement of the central Muskox intrusion and three giant radiating dike swarms, eruption of the Coppermine River flood-basalts, and rifting. Ernst and Baragar (1992) studied the magma-flow patterns in the dikes and showed that sub-

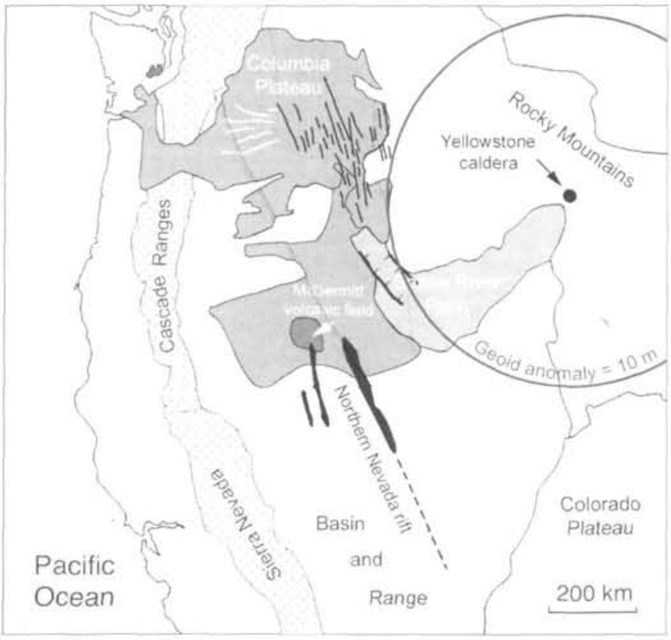

Figure 2. Tectonic and magmatic features associated with the early Yellowstone plume. The McDermitt volcanic field marks the location of plume impact. The circle indicates the possible present-day plume diameter following Smith and Braile (1994). Dark gray—McDermitt volcanic field; middle gray—Columbia River Basalt Group (north) and other tholeiitic basalts of similar age (south); light gray—Snake River Plain volcanism; dotted pattern—coastal ranges; dark lines—dike trends; white lines—Yakima ridge (fold-and-thrust belt) trends; lines with hachures—western Snake River Plain graben. Modified after Zoback et al. (1994) and Cummings et al. (2000).

vertical flow was dominant close to the Muskox intrusion whereas subhorizontal flow dominates at a greater distance, which allowed, with the help of an increasingly growing literature on large igneous provinces and mantle plumes, the proposal of a new plume-tectonics model for this hotspot (Baragar et al., 1996; LeCheminant et al., 1996). Attribution of the Mackenzie igneous and tectonic events to an ancient hotspot appears not to have been a controversial issue to date.

Some of the earliest Yellowstone plume events have been preserved in the well-studied Columbia Plateau (Fig. 2). Only recently, however, have geologic data converged toward this interpretation. In contrast to the Mackenzie plume, although the Yellowstone hotspot was one of the earliest hotspots identified (Morgan, 1971), during a long-lasting controversy the advocates of a hotspot origin for the flood-basalts from the Columbia River Basalt Group (e.g., Duncan, 1982; Pierce and Morgan, 1992; Zoback et al., 1994) have been opposed by those favoring an origin related to backarc spreading (Carlson and Hart, 1987, 1988). Subsequent studies have, however, provided evidence that a mantle plume must have played a primary role in generation of the Columbia River Basalt Group magmas, and temporal relationships make the Yellowstone mantle plume the only appropriate source. Hooper and Hawkesworth (1993) presented isotopic and geochemical evidence that most (97%) Columbia River basalts have a mantle-plume source. Further, they presented evidence for physical, chemical, and isotopic differences between eruptions prior to the onset of Yellowstone plume volcanic activity at 17 Ma and earlier volcanism in the Pacific Northwest. They suggested that after a long period of alkalic to calc-alkalic magmatism (e.g., McKee et al., 1970), probably in part a consequence of subduction of the Farallon and Juan de Fuca plates, a new geodynamic mechanism of magma generation began in Miocene time at ca. 17 Ma (e.g., Carlson and Hart, 1988). Pre-17 Ma magmatism is similar to the volcanism subsequently generated after 15 Ma (Hooper, 1997), i.e., after most of the Columbia River Basalt Group was erupted. In addition to a deep-mantle reservoir (Chamberlain and Lambert, 1994; Carlson, 1997; Chesley and Ruiz, 1998), possibly heterogeneous (Takahashi et al., 1998), the other Columbia River basalt reservoirs may have included the subducting Juan de Fuca plate (Chamberlain and Lambert, 1994) and mafic crustal materials (Chamberlain and Lambert, 1994; Carlson, 1997; Chesley and Ruiz, 1998).

In analogy with numerous flood-basalt provinces associated with mantle plumes, in which flood-basalt eruption is thought to result from the magma generated during spreading of the hot blob at the base of the lithosphere soon after the initiation of the deep thermal instability (White and McKenzie, 1989), eruption of the Columbia River basalts may have directly followed the plume birth (e.g., Pierce and Morgan, 1992). It is possible, however, that the plume is older (Duncan, 1982; Johnston et al., 1996) and incubated for some time prior to basalt eruption below the subducting plate of the Pacific Northwest active margin (Geist and Richards, 1993; Oppliger et al., 1997).

Dike-swarm length and flood-basalt province extent

At both the Mackenzie igneous province and the Yellowstone hotspot, erosion has removed part of the flood basalts. Nevertheless, the extent of the feeder dikes may give a clue to their initial extent. The Mackenzie flood-basalt event was contemporaneous with emplacement of a >2500-km-long mafic dike swarm. If these dikes were feeding lava flows, the minimum radius of the flood-basalt province may have been on the same order as the dike length, maybe more if the influence of topographic slopes is taken into account (Cox, 1989; White and McKenzie, 1995).

During the early development of the Yellowstone plume, eruption of the Columbia River basalts was coeval with emplacement of basaltic lava flows, dikes, and other types of intrusions of similar composition north of the plume center, at Steens Mountain, Oregon (Hart et al., 1989); close to the plume center, at the Nevada, Oregon, and Idaho border; and south of the plume center, along the northern Nevada rift (Zoback and Thompson, 1978; Pierce and Morgan, 1992; Zoback et al., 1994; Ressel, 1996). The same mantle-plume geochemical source has been found west of the northern Nevada rift from 17 Ma until 14.1–14.2 Ma (Ressel, 1996). Differences in chemical compositions of the basalts erupted in these areas have been explained by differences in crustal-reservoir composition (Pierce and Morgan, 1992). Coeval Tertiary basaltic magmatism of similar origin probably exists elsewhere in the western United States, but more work is required before conclusions can be drawn.

The extent of the flood-basalt province associated with the early Yellowstone mantle plume is thus by far not limited to the Columbia Plateau, but has been blurred by subsequent Basin and Range extension southward. If its initial extent was similar north and south of the plume center, it may have been on the order 600 km in radius. Variations in crustal thicknesses may explain why the Columbia Plateau displays the thickest Yellowstone plume-related lava pile. Although the plume impinged on the bottom of a continental lithosphere, the Columbia Plateau formed on an accreted oceanic terrane having intrinsically thin crust (Hooper, 1988, 1997) that was additionally thinned by an Eocene rifting event (e.g., Catchings and Mooney, 1988; Campbell, 1989). Thompson and Gibson (1991) have argued that regions adjacent to plumes having a crust thinner than the crust immediately above the plume center, such as the Columbia Plateau, should trap part of the spreading blob and locally enhance magma production by decompression melting. A large part of flood-basalt magmatism would then occur in the area of thin crust instead of at the plume center. Ebinger and Sleep (1998) suggested that a similar mechanism would explain why the African magmatism and topographic effects due to the Ethiopian plume are spread over a distance one order of magnitude larger than the lateral magma flow required for explaining the Columbia Plateau volcanism from the early Yellowstone hotspot.

Plume diameter

The diameter of the uplifted region above a mantle plume is a good approximation of the size of the diameter of the thermal anomaly (here called the *plume diameter*) at the time that it is spreading out at the bottom of the lithosphere (Griffiths and Campbell, 1991; Olson, 1994). The minimum plume diameter may be estimated several ways. Analysis of magnetic fabric in the Mackenzie dike swarm suggests that magma flow was mainly vertical close to the center of the plume, and mainly subhorizontal farther away (Ernst and Baragar, 1992). If the vertical pattern of magma flow is indicative of the size of the plume head, then its diameter should have been on the order of 1000 km. However, if subhorizontal flow is a consequence of dike ascent to a level of neutral buoyancy in the crust (Lister, 1991), it would not be related to the plume size. Alternatively, the analysis of dike-swarm geometry (Muller and Pollard, 1977; McKenzie et al., 1992; Ernst et al., 1995) may provide an indication of the minimum plume diameter. The boundary between the area of fanning dike geometry and the area of subparallel dikes may be indicative of the minimum plume diameter because it is unlikely that the stress related to a magma center has influence over an area that is smaller than the plume that generated it. Following this analysis, the minimum plume diameter would have been on the order of 2000 km. Analysis of gravity data may provide additional constraints. Dikes converge toward an area that has been correlated with five positive areas of steep maximum horizontal gravity gradients (LeCheminant and Heaman, 1989) corresponding to gravity anomalies within the range of 20–110 mGal (Committee for the Gravity Anomaly Map of North America, 1987) or 80–100 mGal (Goodacre et al., 1987) in magnitude thought to be associated with the source area of the injected magma (Baragar et al., 1996). The difference in gravity attraction between these anomalies and the surrounding areas is ~50–100 mGal (Committee for the Gravity Anomaly Map of North America, 1987). The anomalies define an area of 400 km × 500 km that may reflect the horizontal extent of the main zone of magma melting beneath the lithosphere, which mainly depends on both the plume diameter and the thickness of the lithosphere's mechanical boundary layer during the period of plume activity (Watson and McKenzie, 1991).

The geometry of the early Yellowstone dike swarms from the outcrops mapped to date cannot place any strong constraint on initial plume diameter because the dike distribution is too sparse. Interpretation of the Bouguer gravity anomaly at the Columbia Plateau is made difficult by the subsequent extensional and thermal influence of the Basin and Range. Nevertheless, the anomaly in the Basin and Range is within the range −100 to −300 mGal, whereas the western Snake River Plain is strongly correlated with a −150 to −75 mGal peak and the Columbia Plateau is correlated with a −100 to −50 mGal anomaly, which is the highest Bouguer anomaly east of the Coastal Ranges area (Committee for the Gravity Anomaly Map of North America, 1987). The difference in gravity anomaly between the Basin and Range and the Columbia Plateau probably represents in great part the gravity signature of the early Yellowstone plume. Owing to its location on the border of the Basin and Range, at the first order of correlation, the gravity anomaly is influenced by the Basin and Range, whereas at the second order of correlation, it would reflect loading by the flood basalts and underplated mafic rocks. The voluminous eruptions at the Columbia Plateau and its strong gravity signature suggest that the plume's initial thermal anomaly should have spread beneath the lithosphere over a minimum of 500 km northward but was trapped beneath the Columbia Plateau area because of its thin crust (Thompson and Gibson, 1991). It is interesting that Eocene basins exist also north of the Columbia Plateau (e.g., Heller et al., 1987), but the fact that they were not affected by hotspot magmatism suggests that the current northern end of the Columbia Plateau may also mark the northern end of the spreading blob, which would then have had an initial theoretical diameter on the order of 1000 km. Its shape should have been rapidly modified because of the influence of the moving lithosphere above the plume (Ribe and Christensen, 1994; Smith

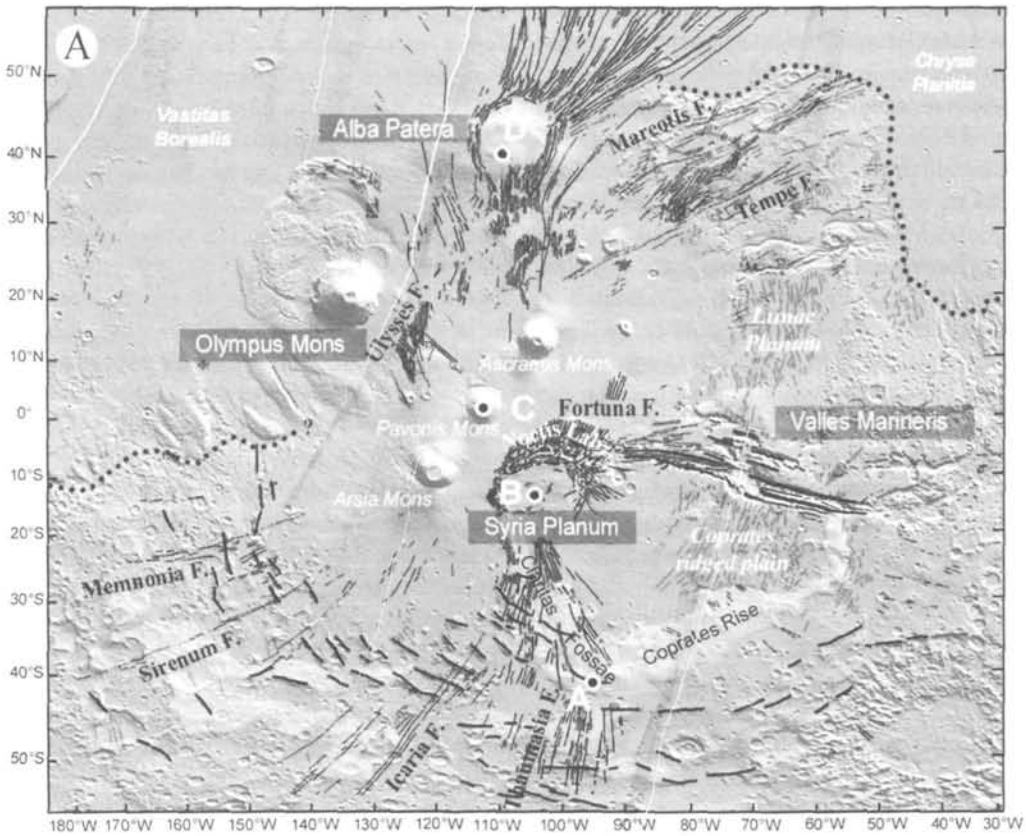

Figure 3. A, Structural map of the Tharsis volcanic province superposed on shaded relief map (MOLA Science Team, 2000). Thin black lines—grabens; thin gray lines—wrinkle ridges; thick black lines—ridges of the South Syria Planum ridge belt (elevation ≈3 km); dotted line—martian dichotomy boundary. A–D—Convergence centers of dike swarms identified by Mège and Masson (1996a). A—Hypothetical early center at Thaumasia Planum. B—Syria Planum magma center. C—Subsequent center in the Tharsis Montes area. D—Alba Patera center. Arrows at Valles Marineris locate Figure 5A (left) and Figure 5B (right). 1° latitude = 55 km. B (facing page), Structural patterns and volcanic flows associated with the Syria Planum events. Larger star—Syria Planum magma center (B in part A). Smaller star—the possible plume recurrence center (Tharsis magma center, C in part A). Circle—estimate of the zone of plume influence on stress-field orientation. From darkest to lightest gray, the lava flows attributed to the Syria Planum magma center are (from Scott and Tanaka, 1986) as follows: 1—late Noachian flood-lavas attributed to the Syria Planum magma center (part of the Npl_2 plateau sequence); 2—lower Hesperian ridged plain material (Hr); 3—lower Hesperian plateau sequence (Hpl_3); 4—upper Hesperian Syria Planum formation (Hsl, Hsu). Structural patterns: Thin black lines—grabens; thin gray lines—wrinkle ridges; thick black lines—ridges from the South Syria Planum ridge belt; lines with hachures—boundaries of the Valles Marineris graben system; dotted line—martian dichotomy boundary; heavy dashed line—southern boundary of lava flows associated with the Tharsis magma center. 1° latitude = 55 km.

and Braile, 1994). This diameter is, however, consistent with the current 1000 km diameter of the ≥10 m Yellowstone geoid anomaly, which is the anomaly height circumscribing typical oceanic hotspots (Smith and Braile, 1994).

Timing of magmatic activity

At both hotspots, outpouring of large volumes of basaltic lava flows occurred contemporaneously with dike-swarm emplacement in a short time span at the onset of plume activity. At the Mackenzie plume, most of the magmatic activity (emplacement of the Muskox intrusion, the Coppermine River basalts, and the Mackenzie dike swarm) took place over a period not longer than 2 m.y. at 1267 Ma (LeCheminant and Heaman, 1989; Heaman and LeCheminant, 1993). The subsequent volcanic activity is not known. It is possible, however, that the 727–721 Ma Franklin dike swarm (Heaman et al., 1992) was produced by some mechanism of plume recurrence (Fahrig, 1987).

At the Yellowstone mantle plume, most of the initially

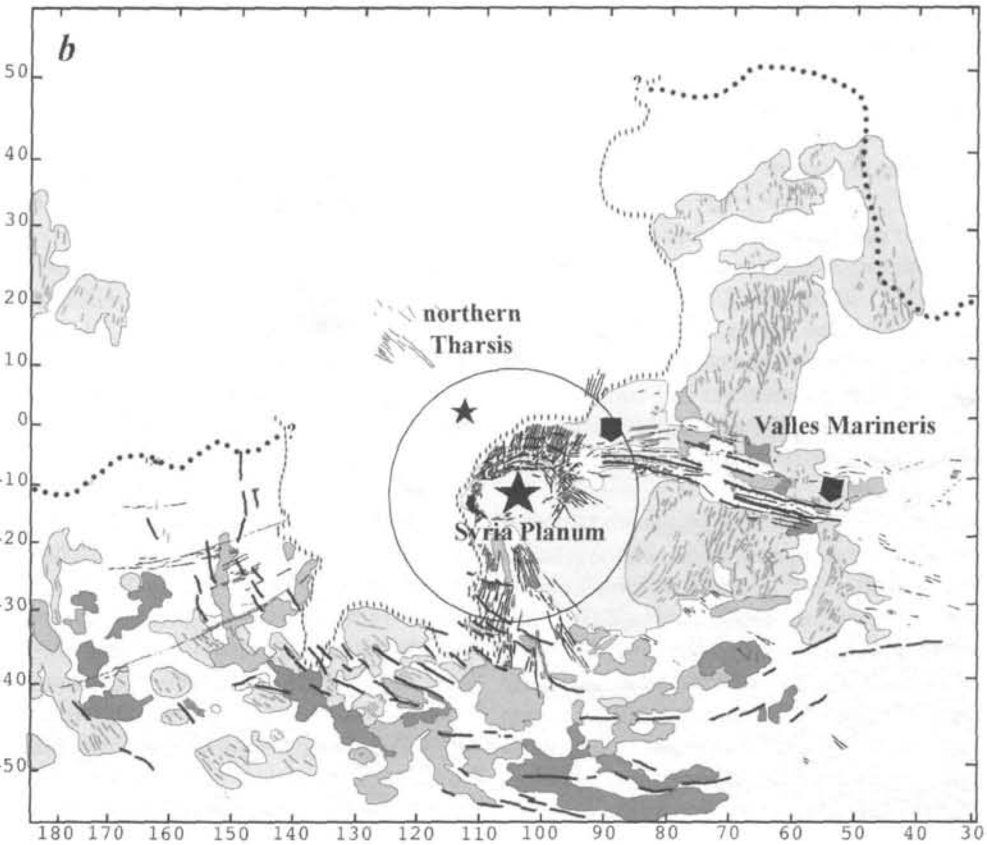

Figure 3 (continued).

164 000 km^2 surface area (Tolan et al., 1989) of the Columbia River basalt flows and their feeder dikes—the Chief Joseph and Monument dike swarms—were emplaced between 17 and 15 Ma (Reidel, 1984; Reidel et al., 1989; Tolan et al., 1989). Along the Yellowstone hotspot track, the earliest volcanic activity was produced at 16.1 Ma at the McDermitt volcanic field, at the junction of Nevada, Oregon, and Idaho (Pierce and Morgan, 1992; Smith and Braile, 1994). More than 50% of the 149 000 km^3 of the Grande Ronde basalts (Tolan et al., 1989), i.e., 43% of the whole Columbia River Basalt Group, was erupted in 0.3 m.y. (Baksi, 1989), which is consistent with field evidence of very high effusion rates (Swanson et al., 1975; Self et al., 1997), resulting in flow durations on the order of weeks to months (Reidel, 1998). New ages (P.R. Hooper, 1999, written communication) suggest that the whole basaltic succession—primarily the Steens, Imnaha, and Grande Ronde basalts—was erupted in 0.5 m.y. between 16.6 and 16.1 Ma. Such short time spans, both for the Mackenzie igneous province and the Columbia River Basalt Group, converge with growing evidence that flood-basalt emplacement is a process typically taking not more than 1 or 2 m.y. Emplacement of 80% of the two largest continental flood-basalt provinces, the Deccan and Siberian Traps, took less than 1 m.y. (e.g., White and McKenzie, 1995, and references therein).

Dike length/thickness ratio

The Mackenzie dike swarm is more than 2500 km long, and mean dike thickness is on the order of 30 m. Mean dike length observed at the Columbia Plateau is on the order of 110 km (S.P. Reidel, 2000, written communication). However, if the dikes were fed from the main magma chamber beneath the McDermitt volcanic field area, they must continue south of the Columbia Plateau in the subsurface as nonfeeder dikes. Their length would need to be on the order of 500 km, which is similar to the minimum length of the dike swarm associated with the northern Nevada rift south of the McDermitt volcanic field (Zoback et al., 1994). The early Yellowstone plume has consequently two associated converging mafic dike swarms, each ~500 km long. Mean dike thickness is <10 m at the northern Nevada rift (Zoback et al., 1994) and 30 m at the Columbia Plateau (S.P. Reidel, 2000, written communication). Statistical analysis of dike length and thickness does not appear to have been published. If it is assumed that the mean Mackenzie dike length and thickness are 1500–2000 km and 30 m, respectively, and those values for the early Yellowstone plume are 400–500 km and 10–30 m, respectively, then crude estimates for dike length/thickness ratio are within the ranges $5–6.7 \times 10^4$ for the Mackenzie dikes and $1.3–5 \times 10^4$ for the early Yellowstone.

Although these values are rough estimates, they are of the same order. The difference between these two ranges is partly due to uncertainty on mean dike thickness. It may also reflect depth dependence of dike length and thickness (Lister, 1991).

Crustal uplift and subsidence history

Early thermal surface uplift and subsequent subsidence are common features of plume-tectonics models (White and McKenzie, 1989; Hill, 1991; Olson, 1994). Evidence was found that the Mackenzie Coppermine River flood basalts were erupted shortly after a period of uplift had taken place, followed by a short period of subsidence (LeCheminant and Heaman, 1989, 1991). The present location of the Yellowstone hotspot corresponds to a 600-m-high topographic anomaly (Smith and Braile, 1994), and it has been argued that the present-day topography of the western United States still shows the trace of the swell that formed when the ascending mantle plume reached the lithosphere (Parsons et al., 1994). However, plume-induced uplift at the Columbia Plateau area appears to have been impeded by the general subsidence of the Pacific Northwest since the early Tertiary. Columbia Plateau subsidence dramatically increased when eruption of the Columbia River Basalt Group began (Reidel et al., 1989). There is no general agreement on the reason why subsidence increased at this very time. Although the geodynamic context of the Pacific Northwest has probably played a role, Mège and Ernst (this volume) have reviewed various hypotheses and argue that this strong correlation should reflect in part the waning plume's thermal anomaly. The Snake River Plain, which represents the Miocene-Pleistocene hotspot track, is also gradually subsiding isostatically, as lithospheric cooling proceeds owing to the plate motion relative to the hotspot (Kirkham, 1931; Brott et al., 1981). The subsidence rate at the Columbia Plateau decreased rapidly after 16 Ma and is correlated with the Columbia River Basalt Group eruption rate (Reidel, 1984; Reidel et al., 1989). This fact suggests that in addition to thermal subsidence, subsidence due to lava loading may have played a major role in the total subsidence observed (Mège and Ernst, this volume).

Extensional tectonics

The Mackenzie and Yellowstone plumes both collided with a lithosphere that was already in an extensional regime that allowed rifting to occur at the onset of magmatic activity. Passive rifting has been interpreted as the mechanism that produced the opening of the postulated Poseidon Ocean, the geometry of which would have been partly controlled by dike-swarm geometry. Following Burke and Dewey (1973), Fahrig (1987) proposed that the plume impact resulted in emplacement of three giant mafic dike swarms, each paralleled by rifts; two of the rift arms formed the Poseidon Ocean basin and the third one failed.

At the Yellowstone hotspot, most extension has occurred at the northern Nevada rift and Oregon Plateau (Zoback and Thompson, 1978; Hart and Carlson, 1987; Pierce and Morgan, 1992; Zoback et al., 1994). However, grabens 17 Ma in age or younger and having the same trend are also observed northward where they can be traced to the western Snake River plain (Mabey, 1976; Hart and Carlson, 1987), the Columbia Plateau, and adjacent areas (e.g., Hooper, 1997). Evidence for the Pacific Northwest Eocene extension (e.g., Heller et al., 1987) has been found beneath the Columbia River Basalt Group (Catchings and Mooney, 1988; Campbell, 1989). However, Eocene extension is probably unrelated to the hotspot, as shown by the difference in magma composition, which denotes a separate tectonic history (Hooper and Hawkesworth, 1993) and existence of a period of tectonic and volcanic quiescence between 20 and 17 Ma between the extensional event and Columbia River Basalt Group eruption (McKee et al., 1970). From 15.5 Ma to the present, extension at the Columbia Plateau was probably due to the influence of the nearby Basin and Range (Hooper and Conrey, 1989).

A mechanism, deriving from Fahrig (1987), for tectonic extension and dike-swarm emplacement for the Mackenzie hotspot can thus account for the early volcano-tectonic evolution of the Yellowstone hotspot, with development of two arms (Columbia Plateau-western Snake River Plain and northern Nevada rift) instead of three, followed by failure of both arms. At both hotspots, rifting is thought to have been passive and to have taken place in that part of the crust above the hotspot that should have been weakened by the plume (White and McKenzie, 1989, 1995). Crustal uplift may have also provided stresses contributing to rifting (Houseman and England, 1986; Griffiths and Campbell, 1991; Hill, 1991).

Contractional tectonics

Formation of the Yakima fold-and-thrust belt on the western Columbia plateau is evidence of compressive deformation in the Columbia River Basalt Group throughout the period of volcanic activity (Camp and Hooper, 1981). The origin of folds is debated owing to the multiplicity of possible stress sources and structural discontinuities in the crust of the Pacific Northwest. In the companion paper, Mège and Ernst (this volume) show that subsidence associated with the incipient Yellowstone plume is likely to have played a major role in Yakima fold-and-thrust belt formation. Fold initiation prior to lava outpouring would be a consequence of early thermal subsidence followed by isostatic adjustment of the Columbia River Basalt Group load. Isostatic subsidence was enhanced by intrinsically thin oceanic crust and subsequent Eocene crustal thinning. From eruption of the earliest flood lavas on, the ridge-growth rate followed the rate of flood-lava emplacement (Reidel, 1984; Reidel et al., 1989). At the onset of plume activity, plume-related stress would have dominated over preexisting regional stress so that the ridges would have formed concentrically about the plume center and normal to the dike swarms (Hooper and Conrey, 1989; Tolan et al., 1989).

Although contractional tectonics has not been observed in the field at the Mackenzie igneous province, it has been argued that dike dilation may have been balanced by formation of contractional structures at the periphery of the uplifted topography (Mège and Ernst, this volume).

SYRIA PLANUM VOLCANIC PROVINCE

The Tharsis volcanic province (Fig. 3A) has been chosen for this study because of the quantity of geologic and geophysical studies published since 1973 (Hartmann, 1973), the intensity and variety of tectonic deformation (Scott and Tanaka, 1986), which helps to elucidate geodynamic processes, and the availability of extensive stratigraphic mapping (Scott and Tanaka, 1986). The geologic history of the Tharsis area is complex, but one of the first events, at the Syria Planum dome, has been rather well preserved. It occurred while the crustal structure in the area should have been essentially pristine, apart from a precursory event at the nearby Thaumasia region (Frey, 1979) and moderate crustal extension northward along the Tharsis rift line (Crumpler and Aubele, 1978). Evidence has been found of recurring magmatic and related tectonic activity (Frey, 1979; Plescia and Saunders, 1982; Scott and Dohm, 1990; Mège and Masson, 1996a) separated by periods of quiescence (Hiller et al., 1982; Mège and Masson, 1996a). The Syria Planum event (discussed later in this paper) corresponds to the main period of tectonic activity of Tharsis. Discussion of later tectonic evolution, including the tectonic evolution during the formation of the Tharsis Montes (Fig. 4), can be found in Mège and Masson (1996a) and in papers referenced therein. The term *Tharsis plume* is used in the text to refer to the mantle source for all the igneous events in the Tharsis region, with the exception of the events related to the Alba Patera and Olympus Mons volcanoes, which appear to have a distinct geologic history (Scott and Tanaka, 1986; Banerdt and Golombek, 2000).

Evidence for plume origin

The Tharsis dome contains most of the tectonic features of the planet as well as three giant shield volcanoes, the Tharsis Montes, having elevations between 14.1 and 18.2 km (Smith et al., 1999). The topographic uplift culminates at ~8 km elevation southeast of the Tharsis Montes in the Syria Planum area. The origin of Tharsis, from comparison with terrestrial hotspots, has for long been attributed to the impingement of a mantle plume on the bottom of the lithosphere (Hartmann, 1973; Plescia and Saunders, 1982). The earliest arguments behind this interpretation have been regional uplift and observation of widespread lava plateaus (Hartmann, 1973; Mutch et al., 1976; Wise et al., 1979) and clues to mobile lithosphere above a fixed thermal anomaly in the mantle (Plescia and Saunders, 1982). More detailed arguments were given by Mège and Masson (1996a), and further evidence is given in the present study from the analogy with the Mackenzie and Yellowstone plumes. Numerical models of mantle convection predict stable, large-scale, mantle upwelling in the martian mantle that can explain the voluminous and polyphase volcanism observed at Tharsis (Breuer et al., 1996, 1998; Harder and Christensen, 1996).

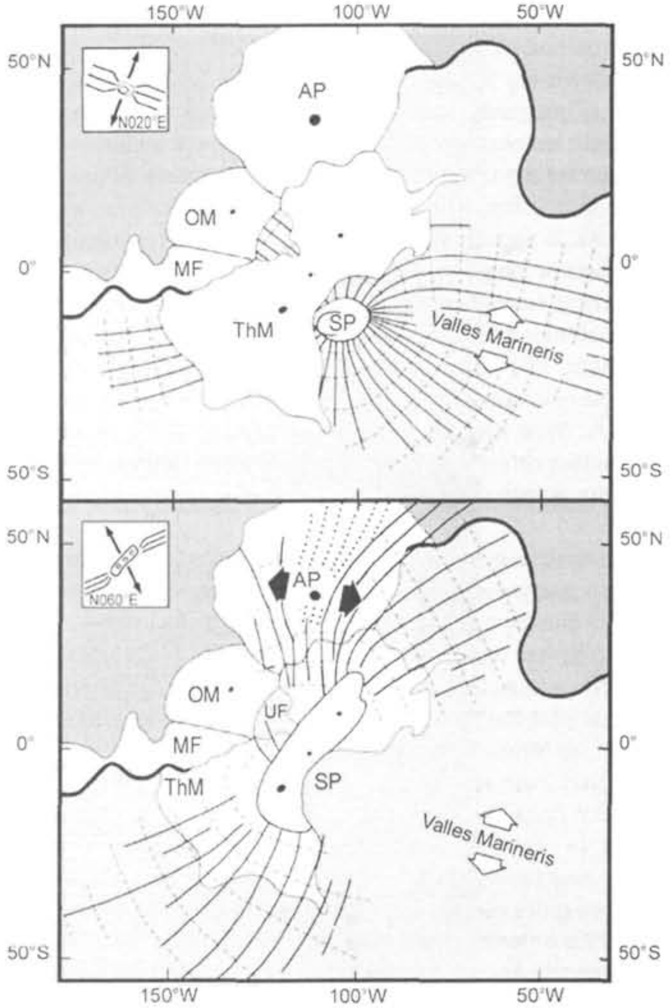

Figure 4. Horizontal principal-stress trajectories (Mège and Masson, 1996a) revealed by statistical analysis of dike-swarm orientation for the Syria Planum (stage 1, up), Tharsis, and Alba Patera magma centers (stage 2, down). AP—Alba Patera; MF—Medusae Fossae formation; OM—Olympus Mons; SP—Syria Planum; ThM—Tharsis Montes; UF—Uranius Fossae. Dark gray—intensely cratered uplands; light gray—smooth lowlands; white—volcanic, eolian, and other deposits subsequent to the Syria Planum volcanic activity. The thickest line locates the martian dichotomy boundary. The inset gives the orientation of the least compressional principal stress of the regional stress field deduced from the geometric analysis of dike patterns (see also Fig. 6). 1° latitude = 55 km.

Volcanic and tectonic activity took place during a considerable period of time, maybe the whole geologic part of the planet's history (Scott and Tanaka, 1986). Nevertheless, a few peaks of magmatic and tectonic activity can be distinguished and make necessary a recurrence mechanism at a temporal scale that can be reproduced with difficulty in numerical models of convection using expected ranges of physical and chemical parameters. It appears likely, however, that both exothermic and endothermic phase transitions in the mantle played a role in stabilizing convection and plume patterns (Breuer et al., 1996, 1998; Harder and Christensen, 1996). The earliest widespread magmatic and tectonic events took place during the Syria Planum volcanic stage.

The Syria Planum volcanic province, whose age is at least 3.5 Ga (see references in Tanaka, 1986), has followed a history in many points similar to the evolution of terrestrial hotspots associated with failed rifts. It may seem paradoxical that an extraterrestrial hotspot of Archean age displays similarities with modern plume tectonics on Earth. A possible reason is that the smaller diameter of Mars (6778 km, Smith et al., 1999) would have allowed more rapid planetary cooling and lithospheric thickening (e.g., Schubert et al., 1992), so that the present-day stage of planetary evolution on Earth may have been attained earlier on Mars. However, the differences in heat-loss conditions may have resulted in a thermal evolution fundamentally different from the thermal evolution of Earth. Thermal blanketing of the crust made possible by the absence of mobile plates may have induced sudden magmatic pulses interspersed between quiet periods (Stevenson and Bittker, 1990; Breuer et al., 1993; Solomatov and Moresi, 1997; Reese et al., 1998).

The volcanic and tectonic events associated with the Syria Planum magma center and discussed herein include generation of huge mafic volcanic flows (Scott and Tanaka, 1986); fanning grabens (Carr, 1974; Frey, 1979; Plescia and Saunders, 1982) displaying geomorphologic evidence of underlying tension fractures (Tanaka and Golombek, 1989) or dikes (Davis et al., 1995; Mège and Masson, 1996a; Mège, 1999a, 1999b; Mège et al., 2000; Wilson and Head, 2000); Valles Marineris, a 2000-km-long canyon system interpreted to be akin to continental rifts on Earth (e.g., Blasius et al., 1977; Masson, 1977; Schultz, 1991; Peulvast and Masson, 1993; Mège and Masson, 1996b; Schultz, 2000a) or to result from gravitational collapse or pseudokarst (Tanaka and Golombek, 1989; Tanaka, 1997; Tanaka and McKinnon, 2000); wrinkle ridges normal to the fanning grabens (Banerdt et al., 1992; Tanaka et al., 1991; Watters, 1993; Mège and Masson, 1996a), and a peripheral circumferential contractional belt (Schultz and Tanaka, 1994; Mège and Masson, 1996a; Mège and Ernst, this volume).

Dike-swarm length and flood-basalt province extent

Two dike swarms have been identified at the Syria Planum province (Fig. 4). Each of them has a minimum length of 3000 km, on the basis of geomorphologic and structural effects at the surface (Mège and Masson, 1996a). Squyres et al. (1987) suggested that emplacement of large sills at Tharsis over long distances may have also significantly contributed to crustal growth, similar to the Mackenzie sills (Mandler and Clowes, 1997). The Syria Planum lava floods include the *ridged-plains materials* (Scott and Tanaka, 1986), which cover $\sim 60 \times 10^6$ km^2 (see Watters, 1993), as well as other smooth terrains of similar and slightly older ages (Fig. 3B). The occurrence of ridged units in impact-crater infills throughout the planet's surface shows, however, that some of them are probably not related to the Syria Planum events. The thickness of the ridged-plains material and other volcanic formations suspected to be of similar origin has been estimated from meteorite-crater burial depths as a maximum of 2 km (De Hon, 1982). However, preliminary observations of layered basement cropping out over the entire 8-km-deep Valles Marineris walls (Fig. 5) and displaying striking similarities to voluminous flood-basalt sequences on Earth (McEwen et al., 1999) may lead scientists to reconsider these estimates.

Plume diameter

Syria Planum has a present-day 100–500 mGal positive free-air anomaly (Zuber et al., 2000), which probably results from the accretion of dense material generated by the plume to the crust. The size of the anomaly may reflect a large part of the whole magmatic and tectonic activity at the hotspot and partly depends on the minimum depth of mantle melting. Consequently, it cannot be used to infer the plume diameter of the Syria Planum magma pulse. Figure 4 shows that the domain of influence of the Syria Planum hotspot-related stress field defines a circle 2000 km in diameter. This diameter is similar to the Mackenzie plume diameter as determined by the same method and may be indicative of the maximum lateral extent of the plume thermal anomaly at the bottom of the lithosphere.

Timing of magmatic activity

The oldest volcanic and tectonic events in the Tharsis region are probably as old as the period of late heavy bombardment (Tanaka, 1986; Barlow, 1988), maybe older, and the most recent lava flows have been dated at 100–200 Ma (Hartmann et al., 1999). The Syria Planum events were preceded by Noachian uplift at the Thaumasia plateau, which may have been associated with early volcanic activity (Frey, 1979; Mège and Masson, 1996a). The Syria Planum dike swarm is probably contemporaneous (Mège and Masson, 1996a) with Syria Planum flood-lava outpouring, as the dikes likely constitute one of the lava flows' primary feeders. Given the size of the Syria Planum plume already discussed, there is probably no requirement for intrusive and extrusive events lasting longer than such events at terrestrial hotspots, i.e., 1 or 2 m.y.

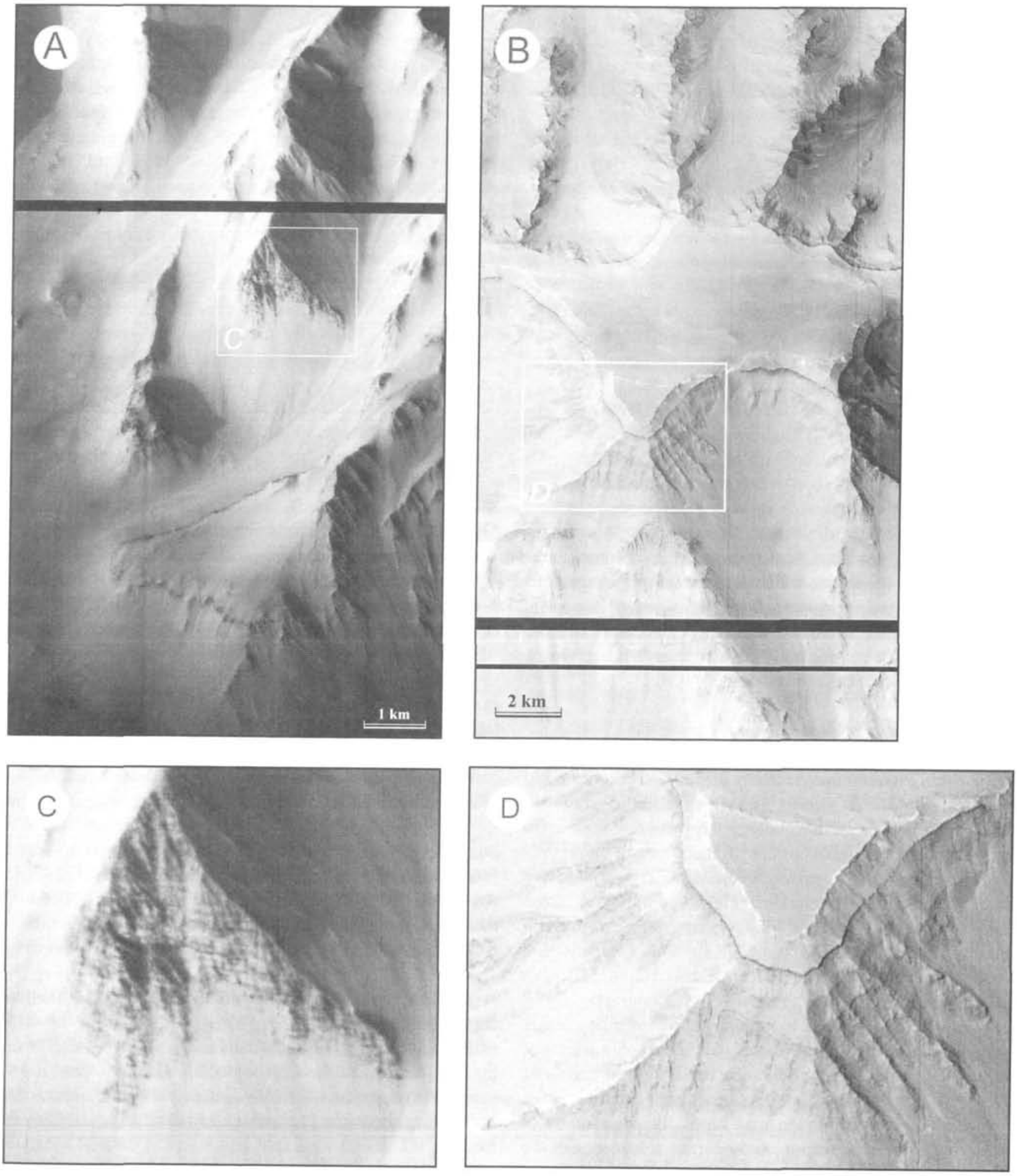

Figure 5. Views (north at the top) of the layered deposits in Valles Marineris. A, Northern wall of Ius Chasma (location: south of left arrow in Fig. 3B). B, Northern wall of Coprates Chasma (location: south of right arrow in Fig. 3B). C, Faceted spur at mid-depth of Ius Chasma wall displaying finely layered stratified materials. D, Plateau and subsurface stratified materials at Coprates Chasma. Wall stratification is observed over the Valles Marineris walls from the top to the bottom, i.e., for a vertical distance of ~8 km, and may represent flood-lava flows that erupted during the Syria Planum plume events or earlier. NASA/MSSS MOC images, ~5 m/pixel.

Dike length/thickness ratio

Most dike tops at the Syria Planum region are located beneath the surface and are detected by collapse pits and troughs formed within grabens (Mège and Masson, 1996a). If it is assumed that dikes propagate at a level of neutral buoyancy in the crust, the average thickness of a dike that is being emplaced in the martian crust has been calculated to be 40% larger than on Earth (Wilson and Head, 2000). However, this thickness can be achieved only if instantaneous (days) crustal stretching can accommodate such dilation. It is not clear whether the dikes are individual dikes that propagated laterally from a central magma chamber (Tanaka et al., 1991; Mège and Masson, 1996a; Wilson and Head, 2000) or individual linear "subswarms" that followed the grabens at depth and were vertically fed by elongated magma chambers that collapsed after magma withdrawal (Mège et al., 2000, and work in progress), similar to magmatic systems at spreading rift zones on Earth (Sinton and Detrick, 1992; Phipps Morgan et al., 1994; Gudmundsson, 1995; Juteau and Maury, 1999; Lagabrielle and Cormier, 1999). If comparison to terrestrial dike swarms is appropriate, the data reported here on dike thickness at the Mackenzie igneous province and the Yellowstone hotspot and the theoretical calculations by Wilson and Head (2000) suggest that the mean thickness of the Syria Planum dikes is 40–45 m, which is noticeably too small to produce the collapse structures whose widths are of kilometer scale observed at the martian grabens. Emplacement of dike subswarms fed by a magma body beneath every graben would then be a more appropriate interpretation.

Crustal uplift and subsidence history

The Syria Planum topography is domed and reaches 8 km in elevation (Smith et al., 1999). The direction of the lava flows inferred from flow-front morphology suggests that they were all emplaced on topography similar to today's. The current topography suggests that considerable underplating should have taken place in order to maintain a high topography until the present. The mafic volcanic plains surrounding the Syria Planum center have subsided until full isostatic compensation has been achieved (see Mège and Ernst, this volume). Consequently, evolution of the Syria Planum uplift topography is consistent with the evolution of topography above mantle plumes inferred from numerical modeling of the topographic effects of plumes; such modeling highlights the initial central and flank uplift before lava flooding, lava flooding from the plume center, central magmatic underplating that keeps the topography at more or less constant elevation, and isostatic subsidence of the lava plateaus around the central area (Olson, 1994).

Extensional tectonics

Extensional patterns contemporaneous with the Syria Planum events are of two kinds (Fig. 3B), (1) the grabens associated with the Valles Marineris canyon system, which is 2500 km long, 600 km wide, 4–10 km deep, and narrow (<10 km, usually ≤5 km), and (2) the shallow (usually ≤300 m), segmented grabens of similar cumulative length. Both the Valles Marineris and narrow grabens are composed of mechanically linked fault segments (Schultz, 2000a); this fact suggests that every segmented graben has evolved as a coherent structural unit. Their geometry is radial about the Syria Planum center. Commonly, one end is located on the high part of the uplift flank, and the other end is located beyond the uplift periphery, in the cratered uplands whose topography may have remained undisturbed by the plume events.

Opening of Valles Marineris began at the onset of Syria Planum volcanic activity or even earlier while the region was already being uplifted by the Thaumasia events (Fig. 3A); crustal stretching proceeded throughout the period of volcanic activity in the Tharsis region until the Amazonian (Tanaka, 1986; Lucchitta et al., 1992; Mège and Masson, 1994; Peulvast et al., 2000). Most tectonic movements, however, date back to the period of activity of the Syria Planum magma center. Strain at Valles Marineris has been calculated to be on the same order as strain at continental rifts such as the Rhine, Oslo, and East African rifts, i.e., only a few percent (Schultz, 1995; Mège and Masson, 1996b).

The narrow grabens associated with the Syria Planum events can be grouped into two swarms, each one corresponding to a major magma center (Fig. 4). The grabens associated with the Syria Planum events display geomorphologic and structural evidence of dike emplacement at depth (McGetchin and Ullrich, 1973; Schultz, 1988; Mège, 1994; Davis et al., 1995; Mège and Masson, 1996a; Mège, et al., 2000). Analysis of dike-geometry pattern suggests that crustal stretching occurred in response to magma center-related stress such as stress produced by inflation of an axially symmetric magma chamber (McKenzie et al., 1992; Chadwick and Dieterich, 1995) and body forces (Merle and Borgia, 1996) superimposed on a regional tensile stress field with stable stress trajectories (Fig. 6). An alternative to the regional tensile stress field hypothesis is that the magma chamber is not axially symmetric, so that the resulting stress field geometry departs from axial symmetry as well.

Crustal stretching was accommodated by dike dilation at depth simultaneous with fracturing and faulting at the surface (Mège and Masson, 1996a). Analysis and interpretation of similar dike patterns at the Galapagos Islands (Crumpler and Aubele, 1978) may provide strong constraints on the stress sources involved at the Syria Planum center (Chadwick and Dieterich, 1995).

Contractional tectonics

Wrinkle ridges (Fig. 3A) are contractional structures observed in flood-lava units associated with the Syria Planum magma center (Scott and Tanaka, 1986). Similar structures are

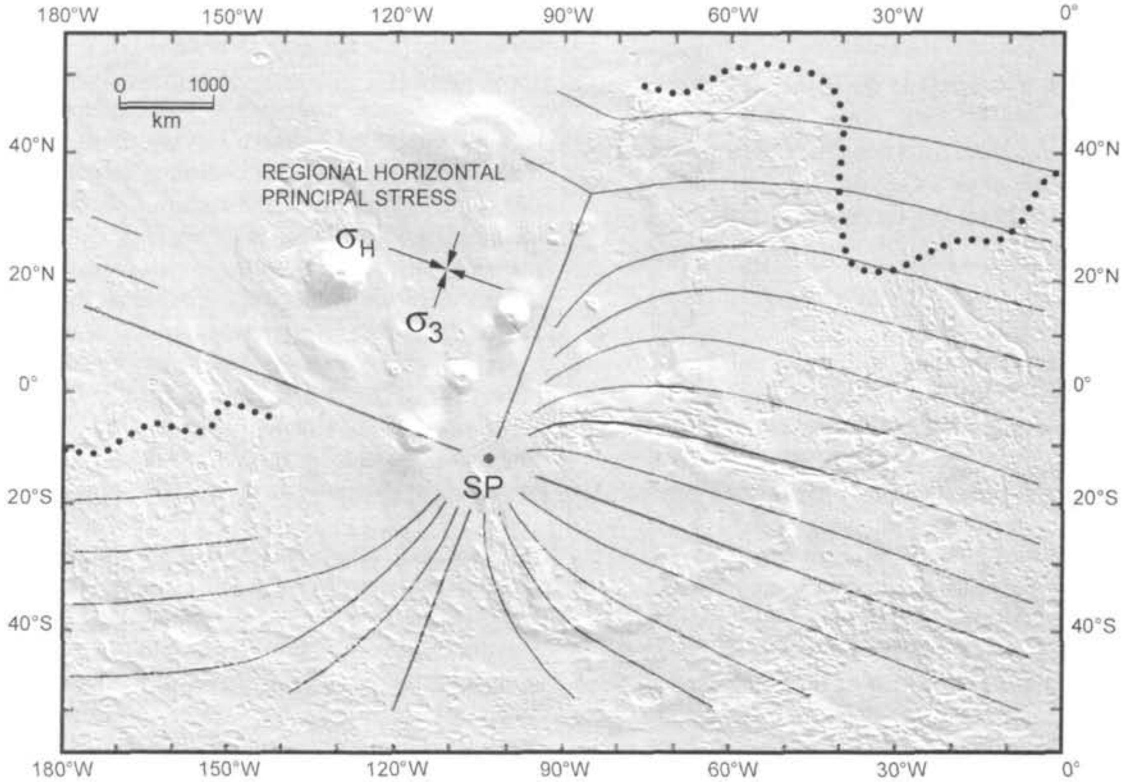

Figure 6. Horizontal principal-stress trajectories in the Tharsis hemisphere during the Syria Planum magmatic and tectonic events. Trajectories were deduced from statistical analysis of the orientation of regional dikes (Mège and Masson, 1996a) superposed on a shaded relief map (MOLA Science Team, 2000). The stress field is analyzed in terms of (1) axially symmetric stress centered at Syria Planum (SP) and a superposed stable regional stress source with σ_3 bearing N020°E (arrows) or (2) stress centered at an imperfectly axially symmetric Syria Planum magma center (not shown).

also observed to be spatially associated with mantle upwellings on Venus (e.g., Basilevsky, 1994; Rosenblatt et al., 1998; Bilotti and Suppe, 1999). They typically form topographic ridges 100 km long, 10 km wide, and a few hundred meters high. Current ideas in regard to their structure call for a combination of folding and thrust faulting (Plescia and Golombek, 1986; Watters, 1988, 1992; Suppe and Narr, 1989; Mège and Reidel, 2000; Schultz, 2000b). Substratum stratification seems to play a key role in their geometry (Schultz, 2000b), and mafic lava floods appear to be the only setting in which they form (Scott and Tanaka, 1986; Watters, 1993). Initiation of most wrinkle ridges in the Tharsis hemisphere began as soon as the flood lavas were erupted (Watters and Maxwell, 1983), and their development stopped before eruption of the oldest lava flows attributed to the Tharsis Montes (Watters and Maxwell, 1986). Their orientation is normal to contemporaneous narrow grabens from the Syria Planum events (Watters and Maxwell, 1986; Mège and Masson, 1996a). Wrinkle-ridge formation has been attributed to thermal subsidence induced by the waning thermal anomaly due to the plume following the initial thermal peak and enhanced by isostatic adjustment of the flood-lava load on a thermally thinned brittle crust (Mège and Masson, 1996a; Mège and Ernst, this volume).

In addition to wrinkle ridges, a peripheral contractional annulus that began forming before the eruption of the Syria Planum lava flows limits the Syria Planum uplift to the south (Fig. 3B). This circumferential pattern has been called the south Syria Planum ridge belt by Mège and Ernst (this volume) and is part of the south Tharsis ridge belt defined by Schultz and Tanaka (1994). The ridges are topographic rises of greater dimension than the wrinkle ridges (elevation on the order of 1 km), and stratigraphic analysis has revealed that their formation began before the main Syria Planum flooding events. In contrast to the wrinkle ridges, they are mostly observed in the intensely cratered basement and are embayed by the Syria Planum lava flows. Their topography has been explained by lithospheric thrust faulting (Schultz and Tanaka, 1994), and their distribution and periodic spacing are consistent with folding and thrusting at the periphery of volcanic loads induced by gravitational spreading of volcanic topography (Merle and Borgia, 1996). Mège and Ernst (this volume) have attributed the south Syria Planum ridge belt to gravitational spreading of the Syria Planum uplift.

DISCUSSION

In this section I recapitulate the common points between the Syria Planum mantle-plume events and the events at the Yellowstone and Mackenzie plumes. Then I test these conclusions against previous models of Syria Planum and Tharsis evolution and discuss implications for further models.

Synthesis

The magmatic and tectonic features at the Syria Planum province share several striking characteristics with the Mackenzie igneous province and the Yellowstone hotspot, including the following:

Plume diameter and flood-basalt province extent. Although some of the products of the volcanism associated with the early Yellowstone plume flood-basalt events have been removed by erosion and other products may not have been identified as such in the Basin and Range region, the initial extent of flood-basalt volcanism is probably two or three orders of magnitude smaller than the extent of the Syria Planum flood lavas. Both the diameter of the plume required for the Syria Planum events and the indicated dike-swarm length are of the same order of magnitude as those at the Mackenzie igneous province. Despite the smaller planetary radius of Mars (half of Earth's), there is probably no need for a megaplume. The huge volume of magmas emplaced at the Syria Planum volcanic province, as well as the length of the dikes, may result from a plume significantly smaller than the Mackenzie plume that nevertheless generated significantly more melt by adiabatic mantle decompression (McKenzie and O'Nions, 1991). For instance, a plume 300 °C hotter than a 1300 °C normal mantle that would impinge on a martian lithosphere 100 km thick would extract two times more mafic melt from the mantle and generate three times more melt than the same plume impinging on a continental lithosphere having similar thickness on Earth (Fig. 7).

Plume recurrence. Plume recurrence and renewal of dike-swarm emplacement occurred at the Tharsis region. Plume recurrence is also observed at some terrestrial flood-basalt provinces (e.g., Heaman and Tarney, 1989; Bercovici and Mahoney, 1994; George et al., 1998; Ebinger et al., 2000) and has been hypothesized for the post-Mackenzie Franklin igneous events. The recurrence time interval is usually on the order of 10^7–10^8 years. Uncertainty on the absolute age of martian stratigraphic units (Tanaka, 1986) allows this range to match the recurrence time interval in the Tharsis region, although longer recurrence times cannot be dismissed.

The mechanism for the recurrence phenomenon is still unclear, however. Separation between the plume head and the plume conduit at the lower-mantle-upper-mantle boundary (Bercovici and Mahoney, 1994) is a possible explanation.

Number of dike swarms. Two dike swarms are associated with the Syria Planum plume. This pattern is similar to that of the Yellowstone plume, which has the Columbia Plateau and northern Nevada rift dyke swarms.

Extensional tectonics. At both the Syria Planum and Yellowstone plumes, extension is concentrated in two opposite regions. The failed northern Nevada rift is south of the early Yellowstone plume center, whereas distributed extension is dominant in the central and eastern Columbia Plateau and in the western Snake River Plain north of the early Yellowstone plume. Distributed extension is also featured on the western side of the Syria Planum uplift, whereas the Valles Marineris failed rift is on the eastern side.

Comparison with the Mackenzie events is inappropriate because of the ensuing successful oceanic spreading.

Contractional tectonics. The Columbia Plateau's Yakima fold-and-thrust belt and the Syria Planum wrinkle ridges present numerous similarities, some of which pointed out by Watters (1988, 1992) and some others by Mège and Ernst (this volume): (1) ridge size (length, width, height), (2) concentric ridge distribution about the plume center in the absence of remote stress, and (3) ridge development only in stratified flood-basalts (compressive structures akin to wrinkle ridges have also been found to form in layered sedimentary rocks [Plescia and Golombek, 1986] but are smaller by at least two orders of magnitude), (4) flood-lava emplacement on a thin, brittle crust underlain by a ductile layer allowing rapid flood-lava isostatic subsidence and ridge development, (5) ridge-growth rate parallel to lava-flow eruption rate, (6) periodic ridge spacing (Yakima fold-and-thrust belt: 20 km, Watters, 1989; martian wrinkle ridges: 30 km, Watters, 1991), (7) ridge formation by combination of folding and thrust faulting, and (8) thin-skinned deformation (e.g., Watters, 1991; Thomas and Allemand, 1993; Mangold, 1997; Schultz, 2000b; Mège and Ernst, this volume), an interpretation that is controversial as, e.g., Zuber (1995) and Golombek et al. (2000) have argued in favor of thick-skinned deformation for the martian wrinkle ridges.

Formation of a peripheral belt at the Tharsis case has been explained by gravitational spreading of the topography generated by the plume (Mège and Masson, 1996a). The same mechanism may have operated at the Mackenzie igneous province (Mège and Ernst, this volume).

Comparison with other tectonic models

Most studies dealing with the origin of tectonic structures in the Tharsis region have been based on elastic models of plate deflection with topography and gravity data providing the constraints; from the model results, stress was calculated (Banerdt et al., 1982, 1992; Sleep and Phillips, 1985; Banerdt and Golombek, 1992). Of three lithosphere-support modes investigated in these studies—dynamic, isostatic, and flexural—a combination of the two latter modes was shown to explain the orientation of a number of tectonic structures. Agreement and disagreement with the observed geology have been discussed by Schultz (1985), Tanaka et al. (1991), Watters (1993), Schultz

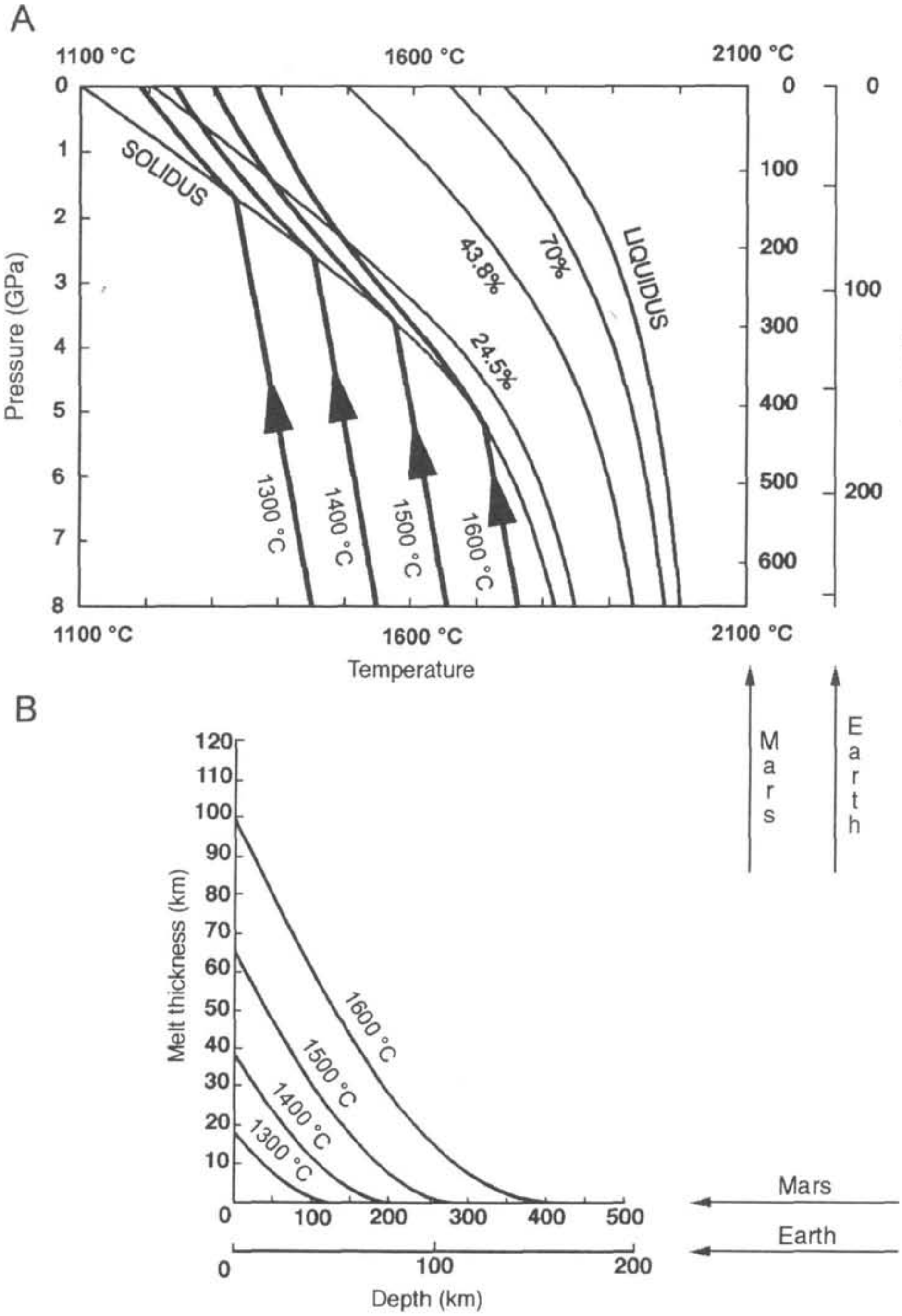

Figure 7. Melt generated by adiabatic mantle decompression at four potential temperatures in the martian and terrestrial mantles. A, Degree of partial melting as a function of minimum melting depth in the mantle (i.e., mechanical-lithosphere thickness). B, Thickness of the extracted melt. Modified after McKenzie and O'Nions (1991).

and Zuber (1994), and Mège and Masson (1996c). Here I focus on a comparison between the Syria Planum plume model and the more recent elastic model published by Banerdt and Golombek (2000) established from the new Mars Orbiter Laser Altimeter topography and gravity data (Smith et al., 1999; Zuber et al., 2000), which supersedes the previous models (Fig. 8). The improvement of Banerdt and Golombek's (2000) models made possible by the improved topography and gravity data shows that the partially isostatic support required to explain part of the tectonic structures in the Tharsis region in earlier models is no longer necessary. A large number of tectonic structures are correlated with strain patterns and magnitudes predicted by lithospheric flexure, confirming earlier results suggesting that Tharsis was formed primarily by addition of magmatic materials to the crust and lithosphere.

Stress sources in plume tectonics and flexural models. Before discussing consistency between the flexural model by Banerdt and Golombek (2000) and geologic observations, I

Figure 8. Magnitude and direction of extensional strain in the Tharsis hemisphere of Mars according to the flexural-lithospheric-support model by Banerdt and Golombek (2000), superimposed on extensional structures mapped by Scott and Tanaka (1986). Gravity and topography boundary conditions are consistent with global gravity and topography models by Zuber et al. (2000). Once the vertical deflection of the lithosphere is determined from gravity and topography, strain is determined by the displacement field. Courtesy of W.B. Banerdt.

compare the stress sources required in the plume-tectonics model described herein and the stress sources accounted for in the flexural model. The stress sources required in the Syria Planum plume-tectonics model primarily include (after Mège and Masson, 1996a, and Mège and Ernst, this volume) the following:

1. Some stress was generated by waning of thermal support, which contributed to wrinkle-ridge formation.

2. Loading stress (isostatic or flexural) induced by equilibration of dense flood lavas also contributed to wrinkle-ridge formation.

3. Stress generated by body forces induced gravitational spreading of the topographic uplift and resulted in development of the peripheral contractional belt. This stress was shown by Merle and Borgia (1996) to simultaneously produce radial grabens in Mohr-Coulomb materials and, therefore, may have contributed to radial graben development.

4. Stress was generated by magma overpressure in the central magma chamber. Such stress was shown to favor emplacement of dikes radially oriented about the magma chamber (Muller and Pollard, 1977; McKenzie et al., 1992) at some lateral distance. Above the magma chamber and at a short lateral distance from it, magma overpressure predicts concentric dike formation (Chadwick and Dieterich, 1995), which is in good agreement with the concentric orientation of dikes and grabens at Noctis Labyrinthus, close to the Syria Planum magma center.

5. Stress generated by magma deflation in the central magma chamber would also predict concentric dike formation

at Noctis Labyrinthus (see McKenzie et al., 1992) between periods of magma-chamber overpressurization.

6. Coherent, remote, deviatoric stress, having stable trajectories over the whole Tharsis hemisphere, has been revealed by geometrical analysis of dike swarms. Such stress contributed to dike propagation to great distances from the main magma center and simultaneous graben formation. This stress provides an explanation (Mège and Masson, 1996a) for the deviation of wrinkle-ridge orientation from a perfectly concentric distribution about Syria Planum (Watters, 1993).

7. Stress due to thermal uplift was initially produced while the plume's hot blob was spreading out and should have induced extensional strain on the uplift very early in the Syria Planum history. In elastic models, such stress favors the development of concentric extensional structures (e.g., Banerdt et al., 1992). At Syria Planum, more recent lava flows (Scott and Tanaka, 1986) have removed any trace of these putative structures, but the geometry of the Noctis Labyrinthus fossae is consistent with reactivation of such structures.

Of the stress sources listed, 1 through 3 are explicitly included in the thin-shell theory used by Banerdt and Golombek (2000). Stresses 4 and 5 were not included, and 6 could have been included but was not found necessary to include. However, neglecting these stress sources may provide a reason why a good agreement is found with some geologic observations and why agreement is imperfect for others, as shown hereafter.

Strength and limitation of the flexural model. Although the flexural model presented by Banerdt and Golombek (2000) has not been fully published to date (in particular with regard to contractional strain), its present form allows a discussion of its consistency with a number of geologic observations. I show next that although the consistency is good for some tectonic structures, a dynamic lithospheric support provided by a Syria Planum-centered mantle plume is probably required to explain the full range of tectonic deformations formed during the Syria Planum magmatic events. Existence of a remote stress source having coherent principal trajectories should also help to boost the agreement between the observed tectonic structures and the theoretical predictions. Then I discuss requirements for a flexural model and a plume-tectonics model to complement each other harmoniously.

Consistency between graben sets and the flexural model can be investigated by comparing the predicted and observed distribution of strain, the orientation of principal strain axes, and calculated vs. modeled strain magnitude in areas where strain has been calculated from imagery and topography data. First, from a strain-distribution point of view, the location of grabens should coincide with the location of predicted extensional peaks. Second, many graben sets are hundreds to thousands of kilometers long. Their structure, morphology, and topography show that strain has not changed significantly along strike, which can be compared with model predictions. A third consistency criterion can be determined from the orientation of predicted principal strain axes and comparison with observed fault kinematics. With the exceptions of strike-slip faulting found in the Coprates ridged plain (Schultz, 1989) and locally in Valles Marineris (Mège, 1994), no evidence of strike-slip faulting has been found at martian grabens. Every crosscutting relationship between grabens and lava flows has demonstrated that strike-slip movements are beyond the resolution of images, i.e., can be neglected at the observation scale. Therefore, the modeled principal-strain trajectories should be normal to the graben trends. Fourth, Banerdt and Golombek (2000) have provided an estimate of compressional strain accommodated by wrinkle ridges at Lunae Planum whose comparison with calculated shortening from available data may limit further the relative role of lithospheric flexure and other stress sources in the deformation observed. These criteria (strain distribution, along-strike strain gradient, orientation of principal strain axes compared to fault kinematics, shortening accommodated by wrinkle ridges) are used here to assess the consistency between the flexural model of Tharsis tectonics as currently published (Banerdt and Golombek, 2000) and geologic observations at the Syria Planum plume. The tectonic provinces that need to be discussed as far as the Syria Planum events are considered include (Mège and Masson, 1996a) the Memnonia and Sirenum Fossae, Hesperian ridged plains such as the Coprates ridged plain or Lunae Planum, Valles Marineris, Noctis Labyrinthus, Ulysses Fossae, Fortuna Fossae, and Claritas Fossae (locations in Fig. 3A).

Strain distribution and along-strike strain gradient. Strain distribution predicted by the flexural model coincides with the observed distribution of the Valles Marineris graben system (Fig. 8). The pattern of increasing strain toward the center of Valles Marineris (Schultz, 1995; Mège and Masson, 1996b) is correctly predicted. East of Valles Marineris, the northeast-trending zone displaying chaotic terrains (Scott and Tanaka, 1986; Witbeck et al., 1991)—the geometry and morphology of which were probably partly guided by tectonic structures (Tanaka and Golombek, 1989; Mège, 1994; Peulvast et al., 2000)—coincides with a northeast-trending area of predicted high extensional strain in the flexural model.

Consistency between the flexural model and Memnonia Fossae is more ambiguous. Although extensional peaks are observed west of Syria Planum, most of Memnonia Fossae is observed to lie between areas of predicted high strain, sometimes as far as 1000 km away from the predicted extensional peaks (Fig. 8). More exhaustive description of the flexural model is required to determine whether this inaccuracy lies within the error bars of the flexural model. Another concern is that the flexural model cannot predict constant strain along the Memnonia and Sirenum Fossae. According to the model, similar to Valles Marineris, strain should gradually increase from the western and eastern graben ends toward the graben centers, which is not observed. In addition, it has been argued that some of these grabens continue farther east in areas where the flexural model predicts only very weak strain but which were covered by subsequent lava flows (Mège and Masson, 1996a).

Noctis Labyrinthus, Fortuna Fossae, the eastern Ulysses Fossae, and the northern Claritas Fossae are located in areas where the flexural model does not predict any extensional strain (Fig. 8). The domed topography of the Syria Planum-Noctis Labyrinthus-northern Claritas Fossae area (e.g., Smith et al., 1999), centered at Syria Planum, is correlated with a positive gravity anomaly of similar diameter, i.e., 1000–1500 km (Zuber et al., 2000). Although other interpretations could be given, it is plausible that the domed topography was created by dynamic uplift centered at Syria Planum upon the arrival of a mantle plume and has been partly or wholly isostatically maintained by magmatic intrusions and underplating that must have occurred simultaneously with the eruption of the observed voluminous lava flows.

It is thus possible that tectonic activity in the Tharsis region is partly due to stress provided by magmatic construction (by volcanism, magmatic intrusions, and/or magmatic underplating), but this mechanism alone probably cannot explain all the extensional deformation observed around Tharsis. To this end, a dynamic support provided by a mantle plume may complement the stress provided by lithospheric flexure. The pattern of constant strain over thousands of kilometers along major grabens such as Sirenum Fossae is consistent with the existence of a regional deviatoric stress field with stable trajectories that would interact with an axially symmetric stress field centered in the Tharsis-Syria Planum area, such as inferred from the plume-tectonics model (Fig. 6).

Predicted and observed fault kinematics. The flexural model by Banerdt and Golombek (2000) predicts the orientation of maximum-principal-strain axes throughout the Tharsis hemisphere. At Valles Marineris, the least contraction axis is in good agreement with the normal kinematics of the Valles Marineris main border faults (Mège and Masson, 1994). At the Memnonia and Sirenum Fossae, the predicted kinematics is normal and sinistral. Many other grabens demonstrated by crosscutting relationships with lava flows to have purely normal kinematics (or expected to be so) are predicted to have a strike-slip component up to 1/3 the total slip movement, e.g., at Icaria, Thaumasia, Mareotis, and Tempe Fossae (Fig. 8). This systematic discrepancy between predicted and observed fault kinematics may reflect either a slight change in the load geometry since the time these tectonic structures formed or the existence of a regional stress field having stable trajectories superposed on the stress field centered on the Tharsis-Syria Planum area. The regional stress source with N020°E-oriented minimum principal stress trajectory (Fig. 6) would tend to remove the predicted strike-slip component on the graben border faults.

Wrinkle-ridge shortening. At Lunae Planum, the flexural model predicts strain on the order of 0.2%–0.3%. Plescia (1991) used photoclinometric profiles and a model of wrinkle-ridge structure combining folding and thrusting to calculate strain in the same region and found 0.29%. Although these values are similar, they cannot be compared directly because strain is two-dimensional (2D, in map view) in the flexural model, whereas Plescia (1991) used shortening along ridge profiles, which gives one-dimensional (1D) values. Strain in 1D is overestimated compared with strain in 2D for two reasons. Although the strain obtained by Plescia (1991) at individual ridges has usually been averaged over 2 to 15 profiles, in many cases all the profiles are located at the middle part of the ridges, i.e., they do not account for strain decrease toward ridge ends where strain gradually wanes to zero. As a consequence, one can expect that 0.29% slightly overestimates strain. More important, in order to equate 1D strain and 2D strain, the interridge area must be as deformed as the ridges. Mège and Reidel (2000) have shown that in the case of the Yakima ridges, strain at the Columbia Plateau is one order of magnitude higher in 1D than in 2D. Therefore, the 0.29% 1D strain found by Plescia (1991) would probably correspond to 2D strain significantly smaller than the 0.2%–0.3% value predicted by the flexural model. At the Coprates ridged plain, 2D strain calculated by structural analysis applying the method described in Mège and Reidel (2000) to data presented in Mège (1999c) is 0.9%–1.6%. Although 2D strain at the Coprates ridged plain and Lunae Planum may differ, this range is probably an estimate of 2D strain at Lunae Planum that is good enough to suggest that 2D strain obtained in the flexural model may be only part of the whole 2D strain.

Complementary nature of both models. It turns out that the plume-tectonics model is complementary to the flexural model. Although the stress sources taken into account in the flexural model are clearly consistent with the formation of Valles Marineris, several imperfections in the correlation between the structures and the model may result from the existence of other stress sources that may be provided by the plume-tectonics model developed here. These stress sources include overpressurization and underpressurization of magma chambers, initial topographic uplift above the plume center slightly before the onset of volcanic activity, and a remote deviatoric stress field with stable trajectories over the whole Tharsis region. The main inconsistency between the plume-tectonics model and the flexural model is in the stress source expected to produce Valles Marineris passive stretching. In the plume-tectonics model, passive stretching is suggested to be partly a consequence of the existence of a significant deviatoric remote stress field, whereas the flexural model predicts that it results exclusively from Tharsis magmatic loading.

Plume tectonics during lithospheric flexural loading

The end of this section is devoted to the investigation of coexistence between dynamic forces induced by a plume and accompanying effects in the crust such as magma-chamber emplacement and evolution and flexural loading of the lithosphere. A mantle plume impinging on the base of the lithosphere beneath the Tharsis region probably would have easily induced erosion of the thermal lithosphere, and the question is whether the mechanical lithosphere will be thermomechanically eroded. Erosion of the mechanical lithosphere may result from lateral shear stress induced by diverging asthenospheric flow at the asthenosphere-lithosphere boundary (e.g., Davies, 1994), as well as plume-induced secondary convection in the lower lithosphere (e.g., Yuen and Fleitout, 1985). If thermomechanical erosion is efficient, the base of the mechanical lithosphere will rise despite its being loaded by magma accretion to the crust. The lithospheric support would thus be mainly dynamic. If it is not efficient, the magma load will deflect the base of the mechanical lithosphere downward, so that despite the existence of the plume, the lithospheric support may be flexural. Which support, either dynamic or flexural, will dominate thus depends on the balance between magmatic loading and thermomechanical lithospheric erosion.

Although removing part of the mechanical lithosphere is significantly harder than removing the thermal lithosphere, thermomechanical erosion of the mechanical lithosphere may occur and favor a dynamic support mode if secondary convection at the top of the plume takes place (Yuen and Fleitout, 1985). Moore et al. (1999) showed that lithospheric thinning on Mars by secondary convection will occur if the thermal anomaly in the plume is in excess of ~200 °C, which would correspond to the temperature in excess at average mantle plumes on Earth.

To the contrary, the large size of the Tharsis plume, the time required for significant thermomechanical erosion to occur, and the huge volume of the magma load would tend to favor building a loading support (isostatic or flexural). Davies (1994) showed that the efficiency of lithospheric thinning by thermomechanical erosion is inversely proportional to the lateral extent of the upwelling. Thus, a large plume such as the Tharsis plume will tend to impede lithospheric thinning by thermomechanical erosion. Davies (1994) also showed that the initial spreading blob is unlikely to cause efficient thinning by itself if secondary convection does not take place. If it does not, efficient lithospheric thinning probably requires that the initial thermal anomaly is fed for some time by continuing ascent of hot material in the plume conduit. Thus, early lithospheric loading and subsequent flexural downwarping of the base of the lithosphere would occur prior to significant thermomechanical erosion. In the Tharsis case, the amount of magmatic material added to the crust during emplacement of flood lavas may have exceeded by far the amount generated at terrestrial plumes, and most has not been eroded since its emplacement. If the layered crust recently observed on the 8-km-deep Valles Marineris walls by the *Mars Global Surveyor* orbiter camera is akin to terrestrial flood-basalts (McEwen et al., 1999), then the volume of magma generated at the Tharsis plume may be even larger than previously expected.

Geodynamically, it is therefore possible that plume tectonics began while the hot blob was spreading out at the bottom of the lithosphere. Dike emplacement would have occurred at this stage. Some radial, narrow grabens would have been formed above some of these dikes. The south Syria Planum ridge belt would have formed or at least begun to form. Then a flexural support would have gradually built up and reequilibrated the lithosphere being loaded. Thermomechanical erosion by the plume would have been ineffective in removing the lithospheric root created by flexural equilibration. Wrinkle-ridge formation would have accompanied downward adjustment of the volcanic plain in response to the waning thermal anomaly as well as to magmatic loading. Crustal stretching would have continued to develop in response to flexural stresses. To help assess this scenario, numerical simulation of gradual Tharsis loading by magmatic construction coupled to thermomechanical modeling of lithospheric erosion by the mantle plume is required to discuss whether emplacement and solidification of this enormous load can overcome the dynamic mantle support of the lithosphere and induce deflection of its base.

CONCLUSION

Terrestrial plume-tectonics models successfully account for the tectonic and volcanic activity associated with the Syria Planum magma center on Mars. The size of the Syria Planum plume, though huge, has terrestrial analogues. The massive amount of erupted materials is only partly related to the plume size and may be a consequence of more efficient partial melting by adiabatic mantle decompression. Similar plume-tectonics events on both planets include (1) early thermal uplift, followed by thermal subsidence after the plume head has collided with the base of the mechanical lithosphere, (2) rapid eruption of basaltic lava flows (not more than a few million years are required) at the onset of plume activity, followed by a long-lasting period (at least tens of million years) of more differentiated volcanic activity, (3) magmatic underplating and emplacement of a central intrusion and giant mafic dike swarm feeding the lava flows, (4) development of a strong positive gravity anomaly produced by the intruded and magmatically underplated materials, (5) radial tectonic extension, leading to the development of successful and/or failed rifts, and (6) development of compressive structures concentric about the main magma center. The latter is not frequent on Earth, but is expected in a number of cases discussed in the companion paper (Mège and Ernst, this volume). Better understanding of plume influence on development of tectonic structures requires numerical simulation of partial melting under martian conditions and of evolution of lithospheric support when a vigorous plume tends to dynamically support the lithosphere while it is being loaded by voluminous magma accretion to the crust. This issue is especially critical in the case of Mars because of the enormous volume of magma expected to have been accreted to the crust during the life of the Syria Planum plume.

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