Extensive Hesperian-aged south polar ice sheet on Mars: Evidence for massive melting and retreat, and lateral flow and ponding of meltwater

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Abstract. Local and regional Mars Orbiter Laser Altimeter (MOLA) topographic data support the presence of an extensive Hesperian-aged volatile-rich south polar deposit (the Dorsa Argentea Formation, Hd, and related units) underlying the present Amazonian-aged polar cap (Api, residual ice, and Apl, layered terrain) and covering a surface area that could be as large as 2.94 x 10^8 km^2 (about 2% of the surface of Mars), over twice the area of the present Amazonian-aged south polar deposits. The deposit characteristics indicate that it contained significant quantities of water ice in amounts comparable to present-day polar deposits. Several lines of evidence for melting indicate that the ice sheet deposits underwent melt back and liquid water drainage into surrounding lows, including a large valley near the crater Schmidt and the Argyre basin. Narrow sinuous ridges lie in a broad linear depression extending from a high near the present polar cap continuously downslope to near the distal portion of Hd. The new topographic data support the interpretation of these ridges as eskers, representing meltwater distribution networks at the base of the receding deposit. Extensive development of large pits and depressions (cavi) have previously been interpreted as eolian etching or basal melting of ice-rich deposits. Analysis of MOLA topography supports the interpretation that they represent basal melting of ice-rich deposits and shows that they have links to the esker systems. Volumetric considerations and topographic lineations suggest that some of the basal melting occurred beneath regions presently occupied by Apl, and that some of the liquid water formed ponds and lakes in the distal parts of Hd. The presence of pedestal craters is further evidence of the removal of extensive volatile-rich deposits and contributes to the quantitative measure of the former deposit thickness. Where did the melt products go? Inspection of the margins of the Dorsa Argentee Formation reveals several large channels that begin there and drain downslope for distances between 900 and 1600 km onto the floor of the Argyre basin, some 3.5-4.0 km below their origin. These channels do not exhibit tributaries. Their broad lateral distribution supports other evidence that deposit melting was areally very widespread and volumetrically significant, and that a large part of the meltwater entered the surface distribution system and was deposited on the floor of the Argyre basin over 1000 km away. Estimates of the present deposit thickness together with amounts of the deposit removed by meltback suggest that the original volume could have been as much as 5.9 x 10^8 km^3, equivalent to a global layer of water ~20 m deep if the deposit consisted of ~50% volatiles. A portion of these volatiles migrated across the surface to pond in adjacent valleys and basins, and into the groundwater system. A significant portion of the volatiles remain in the deposit, representing a net removal from the atmosphere and from the active hydrologic system in early to middle Mars history, and forming an accessible record of aqueous conditions and possible biological environments dating from that time.

1. Introduction

Present volatile-rich south polar cap deposits [Thomas et al., 1992] are of Late Amazonian age and are areally [Tanaka and Scott, 1987] and topographically [Smith et al., 1999] centered near the present south rotational pole. These deposits consist of Api (residual ice) and Apl (layered terrain) [Tanaka and Scott, 1987] and are thought to be relatively young [Plaut et al., 1988; Herkenhoff and Plaut, 2000]. Comparison to the north polar cap shows some similarities and significant differences [Tanaka and Scott, 1987; Zuber et al., 1998; Smith et al., 1999; Plaut et al., 1988; Herkenhoff and Plaut, 2000; Fishbaugh and Head, 2000]. In our discussions about polar deposits we follow many workers [e.g., Fisher, 1993; Zuber et al., 1998] and describe the topography represented by the Amazonian-aged layered terrain and residual ice as the present "polar cap," and we follow terrestrial convention [e.g., Benn and Evans, 1998] and refer to "ice sheets" if the area of a deposit is >50,000 km^2, and "ice caps" if the area of the deposit is less than that.

A number of researchers have proposed that the evolution of water on Mars must have involved more extensive polar ice deposits accumulating at a cold trap at the south pole in earlier Mars history [e.g., Fanale et al., 1982, 1986; Plaut et al., 1988; Baker et al., 1991; Clifford, 1987, 1993; Kargel and Strom, 1992; Schenk and Moore, 1999; Clifford and Parker, 2001; Baker et al., 2000]. Specifically, Kargel and Storm
[1992] noted an abundance of anomalous landforms that they attributed to glaciation; the distribution of these features suggested the presence of large high-latitude ice sheets as late as the Middle Amazonian Period, perhaps covering a significant part of the southern high latitudes [Baker et al., 1991]. However, major questions and uncertainties exist about the nature and extent of such deposits [e.g., Carr, 1996] and whether they would even accumulate at or near the present rotational pole [Schultz and Lutz, 1988]. The availability of new very high resolution altimetry data from the Mars Orbiter Laser Altimeter (MOLA) experiment [Smith et al., 1998, 1999; Zuber et al., 1998] permits reassessment of the local topographic characteristics of individual features and regional topographic relationships of stratigraphic units related to these hypotheses. In this analysis we examine the characteristics of the Hesperian-aged Dorsa Argentea Formation [Tanaka and Scott, 1987] and related units underlying the present polar deposits in the south polar region. We outline the characteristics of the deposits and associated landforms, use MOLA data to address the origins and evolution of the deposit and its structure, and compare this to the present polar deposits.

2. South Polar Regional Deposits and Topography

The youngest extensive stratigraphic units [Tanaka and Scott, 1987] in the south polar region are Late Amazonian in age (Plate 1 and Figure 1) and consist of two closely related deposits. The youngest unit, polar ice deposits (Apl), forms a residual high-albedo feature at the Martian south polar summit and is offset about 200 km north from the pole position at 45°W. Underlying this unit are polar layered deposits (ApI), which are smooth, sparsely cratered moderate-albedo materials that form Planum Australe; in places, ApI is characterized by alternating light and dark layers tens of meters thick with the complete sequence as much as 3 km thick. ApI surrounds Apl in a more extensive deposit and is topographically lower [Smith et al., 1999], with its base generally at about 1600 m elevation. Apl extends north almost continuously to about ~80°N, and is asymmetrically distributed between ~70° and ~80° in the longitude zone 140°-245°W.

Lowest in the stratigraphic column are units of Noachian age associated with the Hellas and Argyre impact basins, followed by several members of a heavily cratered plateau sequence, the youngest member of which extends into the Hesperian [Tanaka and Scott, 1987]. Stratigraphically between the Noachian heavily cratered units and the Amazonian polar deposits is the Hesperian Dorsa Argentea Formation (Hd), which forms polar plains which embryo older highland rocks and ridged plains material. The Dorsa Argentea Formation is stratigraphically below ApI, and is predominantly topographically lower than ApI (about 1000-1800 m elevation), as low as 800 m. The areal distribution of Hd is significant; it is adjacent to ApI, extends northward, and forms a large deposit asymmetrical about the present pole in a direction (~83° to ~57°; 225°-130°W) complementary to the asymmetrical distribution of ApI (Plate 1).

An additional unit, Hesperian-Noachian undivided (HNu), occurs largely surrounded by Hd and has been assigned a Noachian-Hesperian age [Tanaka and Scott, 1987]. HNu is undivided material composed of rough massive deposits exposed in walls and floors of large irregular pits (Angusti and Sisyphi Cavi) and forming closely spaced rounded hills a few kilometers across. HNu tends to occur close to and abutting ApI and within Hd is generally closer to the pole than to the higher-latitude margins of Hd. HNu has a large topographic range extending from lows less than 500 m in the bottom of pits to elevations greater than 2500 m. A significant percentage of the outcrops of this unit occur at elevations in excess of 1800 m, the same elevations as ApI, the polar layered deposits.

Geologic evidence indicates a Hesperian age for the Hd deposits [Tanaka and Scott, 1987] (Figure 1b), and topographic evidence suggests that the deposits underlie much of the present area of Amazonian-aged polar layered terrain. Tanaka and Scott [1987] designated two members of the Dorsa Argentea Formation. The upper member (Hdu) consists of broad smooth deposits embaying surrounding cratered terrain and interpreted to be lava flows. The lower member is similar to the upper member but is more degraded, forming mostly smooth plains around the cavi, and interpreted to be an eolian mantle or lava flows. Early in our analysis we found that the designation of the Dorsa Argentea Formation into upper and lower members was not consistently supported by the topographic data, in that the upper member often was seen to underlie the lower member and vice versa. We thus treated the two members as a single unit. Similarly, topographic relationships showed that unit HNu was laterally continuous with the Dorsa Argentea Formation or formed pits which cut into it [see also Plaut et al., 1988], and thus was less likely to have formed partly in the Noachian period, and to underlie Hd, as proposed by Tanaka and Scott [1987]. Topographic relationships also show that parts of Hd and HNu lie at the same elevation as adjacent exposures of ApI (for example in the area poleward of Cavi Angusti), suggesting that there may be some lateral equivalence and possible overlap in time of these units.

The Dorsa Argentea Formation presently is exposed over a surface area of 1.18 x 10^6 km², while HNu covers 0.34 x 10^6 km², for a combined area of 1.52 x 10^6 km², slightly larger than the present south polar deposits (ApI and Apl combined). If the Hesperian units underlie the present south polar deposits (ApI and Apl), then the total area covered would be about 2.94 x 10^6 km², about 2% of the surface of Mars.

In summary, on the basis of areal distribution, surface morphology, stratigraphic relationships, topographic relationships, and complementary asymmetry about the present pole, we conclude that of the units mapped in the south polar region [Tanaka and Scott, 1987], Hd and HNu represent the best candidates for ancient polar deposits (Plate 1). We now proceed to assess the nature of associated features and their topographic distribution.

3. Structures Associated With Hd (Dorsa Argentea Formation) and HNu (Hesperian-Noachian Undivided)

Several distinctive features and structures have been noted in these two units, and these may provide insight into processes associated with the emplacement of the units. We assess these using MOLA data.

3.1. Dorsa Argentea

Dorsa Argentea form a NW-SE trending sinuous ridge system about 50-250 km wide, 850 km long, and located at about 70°-80°S and 0°-60°W (Figures 1 and 2). The ridges are located exclusively in the Hesperian-aged Dorsa Argentea Formation (Hd) [Tanaka and Scott, 1987], which forms polar plains near Angusti and Sisyphi Cavi, embays older highland rocks and ridged plains material, and underlies Amazonian-aged polar
**Plate 1.** Distribution of geologic units and topography around the south pole of Mars poleward of 55° south latitude. Projection is stereographic. (a) Generalized geologic map of the south polar region [after Tanaka and Scott, 1987]. Amazonian polar residual ice (Api) is white; Amazonian polar layered terrain (Apl) is gray; Hesperian Dorsa Argentea Formation is yellow; Hesperian-Noachian undivided (HNu) is purple; Hesperian ridged volcanic plains (Hr) are green; Noachian cratered terrain and plateau sequence is brown. (b) Mars Orbiter Laser Altimeter (MOLA) topography of the south polar region, from MOLA 54° south stereographic projection generated 5/20/00. Resolution is ~600 m/pixel. Image is overlain on gradient map. Gradient maps here and in other figures are generated from the same data sets used for topography and consist of directional point-to-point slope maps encoded as grayscale. Gray denotes flat, white denotes upslope; and black denotes downslope.
layered deposits (Apl) (Plate 1a). The Dorsa Argentea Formation has been interpreted as eolian mantle or lava flows, with a possible role of ground ice [Tanaka and Scott, 1987] (Plate 1 and Figure 1). We mapped the detailed nature and distribution of these ridges in the central part of their occurrence (Figures 2 and 3) and constructed a MOLA topographic map of the same area. Using these maps, and individual MOLA profiles (Figure 3d), we analyzed the morphology of these features and assessed their regional topographic setting.

In this region the Dorsa Argentea Formation occurs in a shallow linear depression about 1250 km long and about 250 km wide (Plates 1, 2, and 3a) which we designate the Schmidt valley, named for the impact crater located along its western margin; the ridges of Dorsa Argentea generally parallel the trend of this valley. At the polar end of the linear depression, it divides into two valleys and rises to the layered terrain (Apl) and to HNu (undivided terrain of Hesperian-Noachian age). At its northward end, the valley ends in a circular arc (portion of an old crater rim about 450 km in diameter). The floor of the valley lies at about 900 m elevation at the northern end and at about 1200 m in the area of Figure 3, and rises toward 1500 m and above as it approaches HNu and Apl (Plates 1 and 3a). A series of individual profiles across the valley in the area of Figure 2 illustrates the setting and details of the individual ridges (Figure 3). The individual ridges of Dorsa Argentea lie within and along the margins of a broad concave low about 140 km wide and 200-250 m deep. To the east (at about −76.5°) are exposed units of the older Noachian highlands. To the west (−78.5° and to the south), higher topography of Hd occurs (Plates 1 and 3a).

The individual profiles (Figure 3d) show that the ridges have a variety of cross-sectional shapes (simple symmetrical...
inverted-v, flat topped, median trough, asymmetrical, complex), that they are typically 50-100 m in height but range up to 120 m, and that their widths typically are about 2-4 km but some are up to 6 km. The more prominent individual ridges can be traced laterally in map view for typically several hundred kilometers; less prominent ridges can be traced 50-100 km. Comparison to topography shows that the individual ridges typically maintain similar heights for tens of kilometers, but migrate up and down in elevation several tens of meters. In addition, ridges run generally parallel to the strike of the valley, but do not show a dendritic pattern converging on the lowest point in the valley, as might be expected from surface flow of streams. Instead, the ridges generally parallel contours and when they are sinuous, tend to wind back and forth across the valley irrespective of slope (Figure 3). The ridges often fork or branch downslope, and show a variety of intersection relationships: Some merge or branch to and from one another, while some appear to cross and be superposed on others.

An important contribution of the new topography data is the ability to assess the relationship of these ridges to other units.
Plate 2. Color perspective views. (a) Perspective view looking SSE toward the south polar region (toward the bottom part of Plate 3c). Red region to the right is the margin of Cavi Angusti (see Plate 3a for vertical view); the purple area in the center of the image is the southeastern part of Schmidt valley (see Plate 3c for vertical view). Note the braided and sinuous ridges in the upper middle part of the scene emerging from beneath the mantle deposits. These are the ridges interpreted to be eskers, and a more detailed view is shown in Plate 2b. Vertical exaggeration is about 15 x. (b) Enlargement of the central part of Plate 2a, showing the esker-like ridges and their relation to the overlying mantle deposit. This is the same area as that seen in Figure 3. Vertical exaggeration is about 15 x. (c) Perspective view looking NNE from over the south polar region toward the Cavi Angusti, which is in the middle of the image. Individual cavi are seen as depressions (blue and purple floors). The edge of Schmidt valley is seen in the upper part of the image, and the flat plains of parts of the Dorsa Argentea Formation are seen in the foreground. Vertical exaggeration is about 15 x. A topographic map is seen in Plate 3a, and topographic profiles in Figure 4.
Figures 3a-3c show that the ridges are prominent in the north-east part of the image but are absent or obscure in the south-west. The unit to the southwest is part of Hd, but is characterized by a variety of textures different from those in the area of the sinuous ridges, including lineated, braided, smooth, and pitted. Detailed examination of the contact between these units shows that the boundary is transitional in some places (central region), but sharp and scarp-like in others (southeast and northwest). In the northwest the scarp-like contact is accompanied by pits on the southern side of the contact, and on the
Figure 3. Distribution of sinuous ridges in a portion of the Desa Argentena Formation (see Figure 1a for location). (a) Viking orbiter image 421B3. Illumination is from the top right, and the circular numbers refer to individual ridges. (b) Geologic sketch map of the same area showing the location of the sinuous ridge system. Profile locations are shown in Figure 3b. (c) MOLA altimetric profiles across the sinuous ridge system. Profile locations are shown in Figure 3b.
northern side of the contact several small plateau-like outliers of similar terrain are seen. Taken together, these observations (remnant plateau outliers and pits near the contact) suggest that the unit to the southwest overlies the unit to the northeast, and that it is being stripped back toward the south. This observation is supported by the altimetric profiles (Figure 3d) and topographic map (Figure 3c), which show that the surface of the unit to the southwest lies at a higher elevation than the sinuous ridge-bearing unit to the northeast.

These data also provide evidence for the relationship of the sinuous ridges to this unit; the sinuous ridges can be traced up to this boundary and show two types of relationships: (1) They disappear rather abruptly, and (2) they tend to become progressively more mantled until they disappear completely within the overlying unit. These data further support the idea of the unit to the southwest as mantling the sinuous-ridged plains. In addition, within the overlying unit there are occasional sinuous ridge segments exposed and other evidence (albedo streaks and subtle topography) that suggest that the sinuous ridges lie within or under this mantling unit.

Several other occurrences of similar sinuous ridges are seen in the Dorsa Argentea Formation (Plates 1 and 2; Figures 1, 2b, and 2c). One of these (near the crater Lau in the area 72°-79°S; 95°-115°W; Figure 2c) is a single prominent sinuous ridge that extends for over 500 km, emerging from beneath Apl in a region of yardangs and extending for about 140 km in a patchy, discontinuous distribution, apparently due to overlying deposits of a discontinuous nature. It then emerges from beneath these overlying deposits and is very crisp and prominent for about 110 km where it enters Lau, a 100 km degraded crater, crosses the floor, and exits. From this point it winds to the northeast for another 200 km until it is lost in the surrounding terrain. The feature begins at an altitude of about 1700 m and extends down a slope of about 0.06°, to a low point of about 1500 m elevation. Its height is about 40 m, and its width is generally about 6-8 km. In summary, this feature is very similar in width and morphology along the vast majority of its length and shows little evidence for braiding or other morphologic complexity.

Another occurrence of sinuous ridges similar to Dorsa Argentea is found in the Prometheus basin along its western margin (79°-82°S; 300°-315°W; Figure 2b). Here a distinctive sinuous ridge emerges from beneath the layered terrain (Apl) at an elevation of about 1380 m and descends down a slope of about 0.12° to the floor of Prometheus (about 1080 m), extending for a total distance of about 150 km. About 80 km from where it emerges from beneath the polar layered deposits, it bifurcates into two prominent ridges, the westernmost of which appears to form a trellis-like pattern, and the easternmost terminates in a delta-shaped structure. Both terminate at about the same elevation, equivalent to the base of the slope where it turns into the floor of the basin. These sinuous ridge structures are typically about 5-6 km across and 40-60 m high.

3.2. Origin

Sinuous, braided, kilometer-wide ridges have long been noted to occur in a number of places on Mars, preferentially at middle to high latitudes, and the features making up Dorsa Argentea are among the most distinctive [Clifford, 1980; Howard, 1981; Lucchitta, 1982; Tanaka and Scott, 1987; Baker et al., 1991; Kargel and Strom, 1991, 1992]. These sinuous ridges occur singly and in groups or systems up to 800 km long, and display a variety of patterns including cross-cutting, braided, and dendritic. On the basis of their similarities to terrestrial eskers, these sinuous ridges have been interpreted by several workers [Howard, 1981; Kargel and Strom, 1991, 1992; Metzger, 1991] to be Martian eskers associated with glacial processes, but this interpretation has not been widely accepted, and other candidate origins have been proposed [Howard, 1981; Tanaka and Scott, 1987; Carr and Evans, 1980; Ruff and Greeley, 1990; Parker et al., 1986]. The new MOLA data showing the topography and regional setting of the Dorsa Argentea and related sinuous ridges near the south pole allow us to assess candidate origins for these features.

The range of candidate origins proposed for these sinuous ridges includes lava-flow-related ridges or wrinkle ridges

Wrinkle ridges [Tanaka and Scott, 1987] typically differ in their cross-sectional shape, lateral continuity, and detailed structure and texture, and commonly show a topographic stepdown across the ridge [Golombok et al., 1991] that is not prominently seen in these examples, and more linear and orthogonal pattern development in contrast to the braiding seen here. This interpretation cannot be conclusively ruled out, however, particularly if these are partly exhumed and thus modified in some unusual way. The topographic symmetry and map patterns of the ridges argue against lava-flow-related ridges [Tanaka and Scott, 1987]. Exhumed dikes of igneous origin [Carr and Evans, 1980] seem unlikely because of the braided and sinuous patterns, relatively atypical of dike propagation patterns [Pollard, 1987]. Clastic dike origins [Ruff and Greeley, 1990] are more difficult to assess without specific models of formation, but the scale (complex systems, some elements continuous over hundreds of kilometers) seems unlikely. Similarly, although abundant sand dunes are seen on Mars [Greeley et al., 1992], individual laterally continuous linear sand dunes many hundreds of kilometers long [e.g., Parker et al., 1986] seem less likely, although occasional individual features are observed (e.g., in Chasma Boreale [Benito et al., 1997]).

The topography of individual ridges and their relation to the topography of the valley argue against an origin as fluvial and/or lacustrine spits or bars [Parker et al., 1986]. Inverted stream topography [Howard, 1981] might be favored by the relationship to the overlying mantled unit, but the lack of relation of the ridge patterns to underlying topography seems to argue against this hypothesis.

Criteria for assessment of esker origins have been outlined by Metzger [1991], Kargel and Strom [1991], and Kargel [1993], supported by an abundant literature on terrestrial esker morphology [Price, 1973; Benn and Evans, 1998; Warren and Ashley, 1994; Brennand, 2000] and physics [Shreve, 1972, 1985]. On Earth, eskers are long, narrow sinuous ridges composed of stratified glacial drift deposited by a stream of meltwater flowing beneath a glacier in a tunnel or subglacial streambed. Eskers form from glacial meltwater streams which can be supraglacial, englacial, or subglacial. Broken or discontinuous eskers suggest englacial flow and disruption as the esker settles to the valley floor during melt back. Most continuous eskers are interpreted to have formed in subglacial streams forming tunnels in which the sediment is deposited to ultimately form the esker. Eskers form in stagnating glaciers when glacial retreat exposes the ridges of sediment on the subglacial stream floor. Two different types of terminus retreat are commonly observed: (1) active retreat, in which the continuous or episodic forward flow of the glacier deforms and distorts the esker system and produces end and recessional moraines; and (2) stagnant retreat, in which generally long continuously exposed sections of eskers are well preserved, and end moraines and re-
Plate 3. Topographic maps of key areas in the Dorsa Argentea Formation. (a) The Cavi Angusti region. Image is from MOLA 72° south stereographic projection generated June 3, 2000. Resolution is ~300 m/pixel. Image is overlay on gradient map. Numbers along margins show location of profiles seen in Figure 4. (b) The Cavi Sisyphi region. Image is from MOLA 54° south stereographic projection generated May 20, 2000. Resolution is ~600 m/pixel. Image is overlay on gradient map. Numbers along margins show location of profiles seen in Figure 5. (c) The Schmidt valley region. Image is from MOLA 54° south stereographic projection generated May 20, 2000. Resolution is ~600 m/pixel. Image is overlay on gradient map. Numbers along margins show location of profiles seen in Figure 6.
Plate 3. (continued)
cessional moraines are minimal. An important part of the tunnel nature of subglacial streams is that the water is under hydraulic head, accounting for why eskers commonly climb and cross topographic divides [Shreve, 1972, 1985].

Eskers can occur as lone sinuous ridges, in alignments of head-like hills, or in complex systems, including dendritic and anastomosing planimetric structures reflecting meltwater drainage, or rectilinear or orthorhombic systems reflecting drainage through glacial crevasses. Eskers can have confluent and diffuent branches, can cross subglacial topographic divides and may cross-cut one another, and in cross section are known to be sharp-crested, rounded, flat-topped, and double-ridged. Morphology is often related to flow processes [Shreve, 1972, 1985; Brennand, 2000]. Terrestrial eskers are known to range to over 400 km in length, up to 200 m high, and up to 6 km wide. Because of their distinctive and unusual properties and associations, eskers are extremely diagnostic of the presence of ice sheets and glaciation in different areas of the Earth.

The Dorsa Argentae sinuous ridges, and those seen elsewhere in the Dorsa Argentae Formation (Plate 1 and Figure 1a), share many of these characteristics (Figures 2 and 3). They fall in the same range of dimensions and display the same range of shapes. They are sinuous, and form anastomosing, somewhat dendritic systems, with both confluent and diffuent branches. Their sinuosity is similar to terrestrial eskers [Kargel, 1993]. They cross one another and meander and migrate upslope and downslope, consistent with a hydraulic head driving force, in contrast to simple surface flow. In addition, recent work has shown that ridge morphology and cross-sectional shape change as a function of proximity to topographic barriers in a manner consistent with glacial physics and observations of terrestrial eskers [Head and Hallet, 2001a, 2001b]. The stratigraphic relations revealed by the topography (Figure 3) indicate that the sinuous ridges composing Dorsa Argentae emerge from beneath an overlying layer to the southwest. This layer appears to be fragmental and volatile-rich, occurs toward the pole, and is stratigraphically below the Amazonian-aged polar layered terrain. In summary, on the basis of the available data on morphology, topography, stratigraphy, and stratigraphic and areal proximity to present polar deposits, we favor the interpretation of the Dorsa Argentae sinuous ridges as Martian eskers [e.g., Howard, 1981; Kargel and Strom, 1991, 1992; Metzger, 1991].

If this interpretation is correct, then there are several implications. First, the presence of eskers implies that the area of esker exposure was once covered with ice-rich deposits which have subsequently melted back. The extension of the ridges for a distance of over 800 km to a latitude of about 70° then suggests that during the Hesperian, polar ice-rich deposits would have extended to those latitudes. Esker height of at least 100 m is then a minimum thickness for these deposits. Secondly, the preservation of numerous eskers extending many hundreds of kilometers would imply that the polar deposits consisted of significant amounts of water ice, their continuity and lack of abundant end and recessional moraines suggest a period of significant melt back at these latitudes. Further, if polar deposits occurred here during the Hesperian and are so intimately associated with later Amazonian-aged layered deposits, it suggests that the present south pole has been a cold trap at least during significant parts of these two periods. These structures also imply that basal melting was an important process at some point in the Hesperian, a conclusion consistent with other findings at the south pole.

If these features are correctly interpreted as eskers, what accounts for their areal distribution? A number of factors are important in the distribution and characteristics of eskers [Brennand, 2000]. One factor is likely to be topography, with broad topographic gradients controlling the thickness and distribution of superposed volatile-rich deposits, and the pathways taken by their melt products. MOLA data show that the valley in which Dorsa Argentae occur is one of the lowest regions of the Dorsa Argentae Formation (Plate 1), and thus meltwater channels, although ultimately controlled by hydraulic head in detail, would be controlled regionally by the broad topographic trends. A second factor influencing esker development is the presence of a coherent substrate, which may favor formation of eskers relative to a more porous substrate [Clark and Walder, 1994]. Superposition of the volatile-rich layer on a substrate of low porosity (volcanic plains and impact melt sheets) may cause the preferential development of eskers, while in other areas, highly porous substrate (regolith and eolian deposits) might absorb meltwater directly into the groundwater table.

3.3. Angusti and Sisyphi Cavi

Several areas of the south polar region are characterized by a concentration of a range of sizes of irregular steep-sided pits with rounded or flat bottoms which appear to be cut into a relatively smooth upland surface (Plate 1). Regions with this texture were first noted in Mariner 9 data and were described as pitted and etched plains or terrain [Condit and Soderblom, 1978; Murray et al., 1972; Sharp, 1973; Cutts, 1973]. The most prominent occurrences of this texture are in Angusti Cavi (55°W, -77°) and Sisyphi Cavi (5°W, -72°), two regions located adjacent to the present polar layered terrain (Plate 1 and Figures 1, 4, and 5). Later work designated these areas as part of an individual unit (HNu) which occurs largely surrounded by Hd (the Hesperian Dorsa Argentae Formation) and has been assigned a Noachian-Hesperian age [Tanaka and Scott, 1987]. HNu is undivided material composed of rough massive deposits exposed in walls and floors of large irregular pits (Angusti and Sisyphi Cavi) and forming closely spaced rounded hills a few kilometers across. HNu tends to form close to and abutting Apl, and within Hd is generally closer to the pole than to the higher-latitude margins of Hd (Plate 1).

Early in the analysis of the polar regions, the pits and related features were interpreted to be formed through eolian processes, representing an erosional modification of a massively structured sedimentary blanket of eolian or volcanic origin [Condit and Soderblom, 1978; Murray et al., 1972; Sharp, 1973; Cutts, 1973]. Evidence favoring this origin included other eolian landforms in the region, similarities to terrestrial deflation basins, absence of structural control, and exhumation of subjacent units. Evidence of scarp wall sharpness and modification processes [Sharp, 1973] suggested a high ice content in the deposits in which the pits were formed.

Using Viking data, Howard [1981] reexamined these features and interpreted them to be formed by basal melting of ground ice, implying that ice was a major component of the etched plains unit. Howard [1981] pointed out that ridges exposed on pit floors were similar to braided ridges seen nearby, which he interpreted as eskers. Although providing a distinct alternative to the eolian origin for the pits, he cited several problems with his interpretation of the pits and braided ridges as due to subglacial melting: (1) the area of braided ridges shows no other evidence of an ice cover (e.g., grooving, moraines, or
remnant deposits), (2) many contacts of the etched plains deposits with the adjacent smooth plains near the braided ridges seem uneroded (e.g., the size of the pits diminishes gradually near the contact, suggesting an original thinning of the deposit), and (3) there is no clearly identifiable lateral extension of the ridge system into outwash channels or fans. Despite these reservations, Howard [1981] interpreted the braided ridges and etched plains as evidence of subglacial melting and pointed out that they represented melting from the bottom rather than the top. Viking data also revealed that the pits were quite deep (~1 km) [Plaut et al., 1988] and that they appeared to have exposed units underlying the Dorsa Argentea Formation, with which they are associated [Tanaka and Scott, 1987]. As the Dorsa Argentea Formation was interpreted by Tanaka and Scott [1987] to be volcanic, heat associated with the volcanism was suggested as the cause of ground ice melting.

MOLA data provide insight into the morphology, structure, and stratigraphic relationships. We focus first on Cavi Angusti, a 600 x 600 km region located adjacent to Apl and within unit Hl; much of the structure has been mapped as HNu [Tanaka and Scott, 1987]. MOLA data show that deep, irregular depressions commonly having depths in excess of 1000 m characterize the region (Plates 1, 2, and 3a; Figure 4), with some floors reaching as low as 480 m elevation (Figure 4). The area in which Cavi Angusti lies forms a broad region of high topography about 550 km wide extending away from the pole for a distance of about 600 km, and sloping down to the north (Plates 1, 2a, and 3a); Cavi Angusti form large depressions in the summit and flanks of this rise. Individual profiles, perspective views, and topographic maps show that the depressions are larger and more irregular in the center, and more linear toward and away from the polar cap. Toward the polar cap these linear valleys form parallel and slightly sinuous troughs that extend hundreds of kilometers into Apl. Northward of the depressions, the elevation of the surface decreases systematically, the large pits give way to linear chains and channel-like features, and at the margin of the structures the topography gently slopes down to the valley dominated by sinuous ridges (to the northeast) and pitted plains and smooth mantling deposits (to the north). The lowest part of the adjacent Schmidt valley containing the esker-like ridges is often at the level of, or within a few hundred meters of, the depth of the depressions forming the cavi (Figures 3 and 4b). Long-baseline slopes on the walls of the depressions commonly exceed 6°. Locally, some of the depressions are deeper (~480 m elevation) than the adjacent valley (~900-1000 m elevation) (see profile 3 in Figure 4a).

Within the depressions, material is exposed at several different levels (Plate 3a and Figure 4). The deepest level (below about 750 m; 1750 m below the surface; black in Plate 3a) appears in the bottom of rounded to irregular pits primarily in the central parts of the unit. The second level (about 850-1000 m; 1500 m below the surface; purple in Plate 3a) is broadly distributed and is similar in elevation to the adjacent plains region to the northeast, where the esker-like ridges occur. The broad distribution of this level, and its similarity in elevation to the adjacent plains, suggest that it may represent a close approximation to the base of the volatile-rich deposit. The third level (about 1300-1500 m; 1000 m below the surface; light blue in Plate 3a) lies close to the level of much of the other exposed area adjacent to the broad high that constitutes the Cavi Angusti region (Plate 1) and constitutes a small percentage of the total area inside the depressions. Two additional higher
topographic levels can be discerned in the broad rise in which the cavi occur. One lies at about the 1900 m level (light green in Plate 3a) and is part of a broad smooth plain that makes up much of the surface of the Dorsa Argentea Formation. Superposed on this is a contrasting, much rougher terrain (~2000-2600 m level; light green to yellowish in the central part of Plate 3a) forming two types of textures: (1) groups of blocks and knobs forming broadly linear and arcuate regions of varying orientation, and (2) fine-scale linear and arcuate textures composed of ridges hundreds of meters across and up to tens of kilometers long that form parallel and sometimes orthogonally intersecting patterns (the "Inca City" texture first observed in Mariner 9 [see Sharp, 1973]; Plate 3a and Figure 4a, profile 4).

These data provide the basis on which to estimate both the original thickness of the deposits in this area and the processes responsible for the formation of the pits. A plausible maximum thickness of the deposits in this region is given by the difference between the deeper parts of the cavi (~600 m) and the highest level of the surface deposits (~2500 m), or about 1900 m. An estimate of the volume of the material removed to form the cavi can be obtained by assuming the original upper level of the deposits was at about 1500 m elevation on the eastern side where the contour encloses the vast majority of the depressions (about at the edge of the blue-green color in Plate 3a), and at about 2000 m elevation around the rest of the cavi (about at the edge of the green color in Plate 3a) and calculating the present volume of the cavi below this level. The cavi make up an area of about 70,000 km$^2$ (about 20% larger than the area of Lake Michigan) and contain a volume of ~4.0 x 10$^7$ m$^3$.

Cavi Sisyphi is a similar but more scattered collection of pits and etched terrain centered about 1000 km to the east (Plates 1 and 3b, and Figure 1) and lying in a very smooth plain near 1500 m elevation (green in Plate 3b). The smooth plain is marked by several large impact craters, both superposed on, and exhumed from below, the Dorsa Argentea Formation and HNn. Higher topography is seen in the form of cratered terrain at the margins of the unit (in excess of 2000 m, yellow and red color at left of Plate 3b) and clusters of degraded cratered topography locally protruding through the plains (yellow color in right of Plate 3b). In addition, several isolated peaks that were mapped by Tanaka and Scott [1987] as mountains or possible volcanoes rise above the plains (yellow peaks in top center of Plate 3b and Figure 5, profile 6). Detailed examination of Mars Global Surveyor (MGS) data has led Ghataan and Head [2001] to conclude that many of these are excellent candidates for volcanoes that erupted largely subglacially. Cavi Sisyphi themselves show many similar characteristics to Cavi Angusti (e.g., depressions, associated pitted terrain and valleys, several topographic levels within the cavi), although the pits are generally somewhat shallower (~500-1000 m deep), as pointed out by Piault et al. [1988].

The pits that modify the otherwise smooth plains are linear to very irregular in shape, range from a few to several hundred kilometers across, and have several relatively flat topographic levels on their floors. The deepest (750-1000 m deep) are near 500-750 m elevation (black and purple in upper left and upper right of Plate 3b), while some floors are at elevations of about 800 m (light purple, central part of Plate 3b) and are thus about 700 m deep, and still others have floors that are around 1200 m elevation (light blue in Plate 3b and Figure 5b, profile 2), about 250 m deep.

As with Cavi Angusti, these data provide information about the deposit in which they occur and the mode of formation. First, the nature of the pits (steep-sided flat-floored irregular depressions formed in a smooth flat-surfaced unit with little indications of lava flood plain formation) provides further evidence that the deposit in which they occur is formed of fine-grained fragmental material, with a significant component of volatiles [e.g., Sharp, 1973]; the nature of the smooth floors and scalloped margins suggests removal of volatiles (sublimation and/or melting) and collapse of material to form the pits. Gullies observed in the pit walls in Mars Orbiter Camera (MOC) images provide further evidence for the volatile-rich nature of this deposit [Malin and Edgett, 2000]. The several topographic levels within the pits suggest that the deposit may be layered at the scale of hundreds of meters. The total thickness of the deposit can be estimated by the depth of the pits (a maximum of about 1000 m below the smooth plain), under the assumption that the maximum depth is the base of the deposit and the plain is the top. The paucity of Noachian-aged crater rims protruding from the central part of the deposit, together with the highly
the degraded nature of terrain that emerges above the smooth plains (lower left part of Plate 3b and Figure 5, profile 6), suggests that the 1000 m thickness may be a minimum. Association of several pits with suspected volcanic edifices [Ghata and Head, 2001] suggests that there may be a local correlation between volcanism and volcanic loss.

3.4. Origin

The pits and depressions in Cavi Angelus and Cavi Sisyphus are thus very large, commonly very deep, and very widespread. They appear to represent the removal of a significant volume of the volcanic-rich deposit in which they occur (as much as 3 x 10^7 km^3 in Cavi Angelus alone). The great depths and significant lengths of these pits seem unlikely to be due solely to wind erosion, although local eolian modification has clearly taken place on a smaller scale, as seen in MOC images. A compelling correlation is well-illustrated by the MOLA data in the Cavi Angelus region is the linear nature of the valleys extending into Apl (Plates 1, 2, and 3a), and the linear, channel-like nature of the northern depressions transitional to the adjacent valley (Plate 3a). Earlier we mapped a series of sinuous and braided ridges emerging from the distal slopes of the deposits adjacent to the valley, and extending downslope out into the valley (Plate 1 and Figures 2 and 3). We found supporting evidence for the origin of similar features elsewhere in the same valley as eskers, subglacial streams transporting meltwater and sediment (Figure 3). We thus conclude that basal melting of a volcanic-rich layer, as originally proposed by Howard [1981], is the most plausible origin for Cavi Angelus. The linear valleys extending from beneath the polar layered terrain (Apl), the depth of the depressions, the esker-like ridges emerging from the basal parts of the deposit at about the same elevation, all strongly support lateral transport of water from below the present layered terrain, through units Hd and HNu, and out onto the surface of Hd at latitudes of about 75° (Plates 1, 2, 3a, and 3c; Figures 1, 4, and 6). The local very deep portions of the depressions (Plate 2c and black areas in Plate 3a) are below the regional surface on which the esker-like features occur and suggest that at least some of the water drained vertically into the water table [e.g., Clifford, 1993]. The relationships of the depressions to the poleward linear valleys and the equatorward esker-like ridges suggest that the basal melting was not limited to the cavi; nonetheless, in the area of the cavi, up to about 1000 m of overlying unit was removed, indicating that these deposits must have been very volcanic rich or that the subsurface sediment transport system (e.g., the basal meltwater stream producing the esker-like features) may have been very efficient. Cavi Sisyphus shows many similar characteristics and is also a candidate for volcanic loss and basal melting processes, perhaps aided by subdepositional volcanic activity [e.g., Ghata and Head, 2001].

These new MOLA data also address Howard's [1981] initial reservations concerning an interpretation of pits and braided ridges as due to glacial melting. These reservations, listed first, and new data relevant to them, in parentheses, are as follows: (1) no other evidence of an ice cover (topographic evidence is now clearly seen of remnant deposits overlying the braided ridges), (2) many contacts of the etched plains deposits with the adjacent smooth plains seem uneroded (topography now shows that they overlie smooth plains and that the contact is gradational), and (3) no clearly identifiable lateral extension of the ridge system into outwash channels or fans (topography data now suggest ponding of water in the lower parts of valley). Thus we conclude that the new MOLA data address the reservations of Howard [1981], strengthening his initial interpretation of subglacial melting in these deposits.

4. Nature and Fate of Melt Products

Several analyses using MOLA data thus support the general concept that significant volcanic-rich deposits formed an extensive ice sheet earlier in Mars history in the south polar area. Geologic evidence indicates a Hesperian age [Tanaka and Scott, 1987; Plaut et al., 1988] for these deposits, and topographic evidence suggests that the deposits underlie much of the present area of Amazonian-aged polar layered terrain. The associated geological features (cavi and esker-like ridges) are consistent with the presence of volcanic-rich materials. Significant volatiles were lost when portions of the ice sheet underwent melt back to approximately the present position of the edge of the layered terrain. Evidence that a significant part of these volatiles was water is supported by (1) the presence of sinuous ridges, plausibly interpreted as eskers and marking the location of subglacial streams that drained much of the retreating polar ice sheet, and (2) cavi, structures plausibly interpreted to be the surface manifestation of subglacial melting and vertical and lateral movement of water.

Examination of the distal portions of the esker-like ridges and Cavi Angelus provides evidence relevant to the fate of the meltwater (Plate 3 and Figure 6). Both of these features end in a north-northwest oriented valley just east of the crater Schmidt. The Schmidt valley here is approximately 350 km wide and extends from the edge of the present polar cap north toward the Argyre basin for a distance of over 750 km (Plate 3c). It is bounded on the west, north, and east by irregularly shaped, topographically high segments of Noachian cratered terrain [Tanaka and Scott, 1987]. These cratered terrain units form a topographic barrier to the drainage of any meltwater associated with the esker-like features or cavi formation. A series of topographic profiles illustrate the nature of the basin (Figure 6a). The northern margin of the basin rises to over 2000 m (profile 1), then descends to the south to about 1200 m (profile 2), and then descends further to the regionally flat basin floor which lies at a level of about 900-1000 m (profiles 3-5). In the central part of the basin, some impact craters are almost completely buried, while others are superposed on the deposits (compare two similar-sized craters in profile 4). Older impact craters around the margins of the basin usually have their rims breached or embayed and destroyed toward the basin (craters between profiles 2 and 3). The valley narrows in the vicinity of the crater Schmidt (profile 6), and then widens again east of the distal deposits of the Cavi Angelus region (profiles 7 and 8). At the southern edge of the valley, the topography begins to rise toward the south pole region, and the esker-like features begin to be developed on this upward slope. Longitudinal profiles along the strike of the valley also illustrate these trends (profiles 9-12), with profile 10 approximating the center of the valley and profiles 9 and 11 its western and eastern margins [see also Head and Pratt, 2001].

Products from melt back of the volcanic-rich deposits and the formation of the esker-like features and cavi would have entered Schmidt valley at its southern end (Plate 3c and Figure 6a). On Earth, eskers represent the former position of subglacial transport of meltwater and deposition of sediment carried by these streams. Melt back and loss of the overlying volcanic material then exposes the channel floor as a sinuous ridge. These contained streams, and thus the eskers, change morphology or
Figure 6. Schmidt Valley. (a) Altimetric profiles illustrating the characteristics of topography in the Schmidt valley region. Map showing the characteristics of Schmidt valley and the location of profiles is seen in Plate 3c. Perspective views of this region are shown in Plate 2. (b) Interpretive block diagram showing the relationships of the esker-like ridges, overlying volatile-rich debris mantle, and ponding of meltwater in a proglacial lake.
terminate when they flow into an outwash plain or a standing body of water in front of the glacier. Using the termination of the distribution of the esker-like features as an estimate of the extent of the volatile-rich deposits, we find that they define a level of about 1000 m elevation (Figure 6a). Examination of the eastern and northeastern margin of the Cavi Angusti region shows that esker-like features emerging from below this deposit also terminate at about this same level (Plate 3a and Figure 4b).

If this interval represents the termination of the subsurface flow of meltwater at the edge of the volatile-rich deposit, then the morphology of features at lower levels should provide information about the meltwater fate. Examination of the region in the valley below this level shows no evidence for extensive channels or braided deposits that might be expected if such streams emerged onto an outwash plain. Instead, the regional surface is very flat (Plate 3c and Figure 6a) and in detail is extremely hummocky (Figure 6a, profiles 3, 5, and 10). We interpret these observations to mean that the sediment-laden meltwater emptied into the Schmidt valley and ponded there (Figure 6b). The present topography is interpreted to be the result of regional smoothing by this emplacement process, and the hummocky roughness may be due to the sublimation and removal of portions of the volatile-rich material subsequent to its deposition in this low valley. The implication of these observations is that the meltwater was deposited in an extensive lake filling the Schmidt valley during the period of meltwater formation (Figure 6b) [Head and Pratt, 2001].

An unusual set of features and deposits is located just to the north of this zone along the western margin of Schmidt valley south of the crater Schmidt and at the edge of the Cavi Angusti deposits. This region consists of five parallel zones (upper right part of Plate 3a, Plate 3c, and Figure 4a, profile 1) the boundaries of which parallel the topography. At the highest topographic level (in excess of 1700 m) the morphology is characterized by the deep troughs and pits of Cavi Angusti. At the margins of this zone, and generally parallel, occurs a transition region dominated by smooth plains interrupted by linear pits. A second zone is seen between about 1400 m and 1150 m elevation; it consists of a 250 km long, 40-80 km wide region of pitted plains sloping toward Schmidt valley. The pits in this zone are uniformly smaller than those making up Cavi Angusti, ranging from as large as 10 km down to several hundred meters, and typically having diameters of about 1-3 km. A narrow trough a few kilometers wide and a few tens of meters deep separates this zone from the next lowest one, which has a much smoother surface and almost no pits. This zone is about 30 km wide and is separated from the regionally smooth, but locally rough topography of the valley floor by a scarp/trench about 100 m in height.

Previous interpretations of the origin of the pits in this terrain have focused on eolian processes and deflation pits [e.g., Murray et al., 1972]. The localized distribution of these features, their location at the edge of an extensive deposit of volatile-rich material that is undergoing melting (Cavi Angusti), and their formation along the margin of a valley that was the likely recipient of an extensive volume of meltwater forming a lake suggest an alternate interpretation. On the basis of our observations, we interpret these zones to represent the location of a former marginal ice-rich deposit, with the outer zone representing the equivalent of an outwash plain, and the inner pitted zone representing the location of a decaying ice sheet which left isolated ice blocks surrounded by sediment, which then melted or sublimed to produce the Martian equivalent of terrestrial kettles [Head and Pratt, 2001]. Thus we interpret these features to have formed at the junction between the margin of the ice-rich deposit (Cavi Angusti) and the lake forming on the floor of Schmidt valley.

If these interpretations are correct, then Schmidt valley could have been the site of an extensive lake [Head and Pratt, 2001]. Topographic data permit us to assess the size and depth of such a lake. Using the margins of the occurrences of esker-like features and distal Cavi Angusti deposits, we outlined the topographic contour of the valley if it were filled to this level (about the 1050 m contour). This level produces a flooded area about 360 km wide and 750 km long, or about 270,000 km², about 73% the size of the Caspian Sea, and almost 5 times the size of Lake Michigan. The presently lowest point in the valley (excluding impact craters) is about 710 m, and thus if this were the floor of the valley at that time, the water in the lake could have been as much as 340 m deep.

A conceptual diagram of these relationships is shown in Figure 6b. In this scenario, meltwater from the receding volatile-rich layer was distributed by subsurface channels (which ultimately became eskers) and drained into the adjacent low to form an extensive lake. The distal end of the valley in which the esker-like ridges occur is very flat and smooth compared to others in the southern hemisphere [Kreslavsky and Head, 2000], supporting the interpretation that there was ponding and sedimentation. In addition, a notch-like breach occurs at the distal end of the valley. A topographically distinct channel (channel 1, described below) has been traced from this breach down into the adjacent Argyre basin, supporting the hypothesis that liquid water flowed from the distal parts of the retreating polar deposits, onto the floor of Schmidt valley, and through a pass and channel, down onto the floor of the Argyre basin, over a 1000 km away.

In this example and others we have examined, the presence of cavi, esker-like ridges, pits, and smooth marginal areas strongly suggests that much of the water derived from melting of the Dorsa Argentae Formation and related deposits migrated laterally to the edges of the deposits. We asked the question, If a significant amount of melt back of these deposits and some ponding of the meltwater occurred, where did the water ultimately go? To address this question we analyzed the regions in the directions downslope from the margins of the Dorsa Argentae Formation (Hd). We located five major sinuous channels emerging from the distal margins of the circumpolar deposits and extending downslope over hundreds of kilometers into adjacent lows (Figure 7).

Channel 1 emerges from the distal end of the occurrence of Hd containing Dorsa Argentae (the Schmidt valley described above). It is first clearly observed as a notch in an ancient, heavily degraded crater rim between the craters Von Karman and Fontana at the eastern edge of Argyre Dorsum (Plate 1 and Figure 1). It emerges at about 1000 m elevation and winds its way 400-500 km northward across a relatively flat several hundred meter deep plain just north of Argyre Dorsum. The channel sometimes becomes indistinct on the topographic map, usually corresponding to flat areas where ponding is likely to have occurred. As it reaches the edge of the Argyre basin rim (Charitum Montes), it winds through a low in the rim topography and plunges some 3500 m over a distance of about 400 km to the basin floor where it loses its identity.

Channel 2 (Surius Valles) emerges from the distal end of the Hd/NHv deposit at an elevation of about 1500 m (Figure 7),
flows into a crater and out the other side through a notch, turns westward and flows about 150 km downslope, and then turns northwestward down a flat valley mapped partly as Hd, just below 1000 m elevation. A distinctive narrow linear trough can be seen in the topography over parts of this 650-700 km segment, and as the channel reaches the area between Von Karman and Phillips, it becomes more and more topographically distinctive. At this point, it turns northward again, and flows about 700 km down to the floor of the Argyre basin through Charitum Montes, dropping about 3200 m in elevation. Over the majority of this last segment, it is distinctive enough to be mapped as Hesperian-aged channel material (Hch) [Tanaka and Scott, 1987].

Channel 3 (Dzigai Valles) emerges at the distal end of the major continuous occurrence of Hd/HNu just northwest of Sisyphi Cavi in a specific protrusion of HNu lying inside a crater (Figure 7). The inner rim of the crater is at about 1500 m elevation, and the channel cuts a notch in the western rim, extending westward down a linear valley about 300 km until it turns north and flows about 400 km down a valley to a low point below 500 m elevation just southeast of Miraldi crater, where it is joined by channel 4. Together, the combined channel turns northwest, flows through a notch just east of the Miraldi crater rim, and then plunges through Charitum Montes to the Argyre basin floor, dropping about 3000 m in elevation over about 400 km distance. The total elevation drop for channel 3 is about 4000 m over a distance of about 1250 km.

Channel 4 (Doanus Valles) emerges from the margin of the major continuous occurrence of Hd/HNu just northwest of the crater Wegener (Figure 7). It begins in a notch in highland topography at about 1500 m elevation, adjacent to the Hd/HNu unit, and flows northwest for about 75 km, then turns west and traces a sinuous course down a distinctive valley for about 300 km until it reaches the basin just southeast of Miraldi crater, where it is joined by channel 3. From its origin to the edge of this valley it drops about 500 m in elevation, and then continues to decrease in elevation another 500 m to the point where it joins channel 3. The combined channel 3/4 plunges through Charitum Montes to the Argyre basin floor. The total elevation drop for channel 4 is about 4000 m over a distance of about 1100 km.

Channel 5 (Palacopas Valles) does not show a clear correlation with the margin of the Hd/HNu continuous deposit. It is most distinctive in a valley to the east of Darwin crater (Figure 7). This 200 km long segment appears to flow into Darwin, and exits Darwin through a notch in its northwest rim, continuing through Charitum Montes toward the basin floor (Figure 7). A notch in the western rim of a subdued crater at the edge of the Hd/HNu deposit due north of Wegener is a candidate for a portion of this channel, but no evidence has been found for the channel in the relatively flat valley between the edge of the deposit and where it begins southeast of Darwin.

On the basis of these characteristics and relationships, we draw the following conclusions: (1) The four major channels are closely associated with the distinctive Hd/HNu deposits interpreted to be volatile-rich [Thomas et al., 1992; Carr, 1996] and of polar origin. They begin at the margins of the regional deposit at elevations of 1000-1500 m and drain downslope for distances between 900 and 1600 km onto the floor of the Argyre basin, some 3.5-4.0 km below their origin. (2) These channels do not exhibit minor tributaries, dendritic patterns, or other associated features that would suggest runoff from re-
gional precipitation or extensive groundwater sapping over their course. (3) These channels are interpreted to represent the distal downslope distribution systems of the meltwater associated with extensive and varied basal melting within the Hld/HNu unit (e.g., esker-like ridges, cavi, and lake). (4) The broad lateral distribution of these channels along the margins of the Hld/HNu deposit (each separated by 200-400 km lateral distance) supports other evidence that basal melting was areally very widespread and volumetrically significant. (5) The distance over which the channels flowed (900-1600 km) indicates that a significant part of the basal meltwater entered the surface water distribution system. The amount that entered the subsurface groundwater system below the ice sheet is unknown. (6) The notch-like depressions formed where the channels cut through preexisting topography in several places along their courses suggest that ponding, overflow, and down-cutting may have occurred. Thus some of the unusually flat nature of the intercrater plains may be due to ponding and sedimentation, and perhaps to the solidification of ponded water to form ice. (7) All of these data strongly support the presence of a large, water ice-rich deposit (Hld/HNu) in the south polar region prior to the present Amazonian-aged deposits (Api, Apl), and suggest that volatile-rich portions of this older deposit underwent basal melting and melt back earlier in Mars history during the Hesperian. The areal extent of these deposits is significant, but much less than the Austral ice sheet proposed by Baker et al. [1991]. (8) The termini of these channels uniformly are situated on the floor of the Argyre basin, indicating that a significant part of the meltwater drained by these channels flowed into this basin. This supports previous hypotheses that the floor of Argyre involved standing bodies of water and perhaps ice and glacial processes [e.g., Kargel and Strom, 1992].

5. Discussion and Conclusions

5.1. Areal Extent of the Volatile-Rich Deposit

The geological features associated with these units are consistent with volatile-rich materials, and their distribution suggests that such deposits extended to latitudes as low as about −57°. The Dorsa Argentia Formation and HNu as presently exposed cover a combined surface area of 1.52 x 10⁶ km², slightly larger than the present south polar deposits (Api and Apl combined). If they underlie the present south polar deposits (Api and Apl), which seems plausible given the geometric relationships documented by the MOLA topography data, then the total area covered would be about 2.94 x 10⁶ km², slightly over 2% of the surface of Mars. This is approximately 50% larger than the present Greenland ice sheet (~2 x 10⁶ km²) and about 22% of the area of the Antarctic ice sheet (13.5 x 10⁶ km²).

5.2. Evidence of Melt Back and Its Style

Significant volatiles were lost when portions of the deposits underwent meltback to as far as the present position of the edge of the layered terrain. That a significant part of these volatiles was water is supported by (1) the presence of sinuous ridges plausibly interpreted as eskers and marking the location of subglacial streams that drained much of the retreating polar ice sheet, (2) cavi, interpreted as the surface manifestation of subsurface melting and vertical and lateral movement of water, (3) the presence of a lake filling the Schmidt valley and draining out to the north, and (4) large channels exiting from the margins of the deposit and draining into the Argyre basin to the north.

Two features also indicate that a significant amount of sediment was transported by the water: (1) The presence of esker-like features shows that the subdeposits channels transported and deposited sediment for distances of hundreds of kilometers, and (2) deposits at the termination of the esker-like features and channels (on the floor of Schmidt valley and on the floor of Argyre) also show that sediment was transported over many hundreds of kilometers.

On Earth, glaciers that accomplish net retreat through repeated stages of advance and retreat tend to form eskers during retreat and then obliterate or segment them during advance. In contrast to this active retreat behavior, eskers are often more well-preserved when continuous melt back occurs unaccompanied by glacial advance (stagnant retreat), leaving the basal deposit structures, such as eskers, essentially unmodified. On Mars, the preservation and continuity of esker-like ridges for extensive lateral distances (many hundreds of kilometers) and lack of evidence for abundant terminal and recessional moraines suggest that retreat may have been stagnant rather than active. If true, melt back may have represented a distinctive step-like global change in conditions, rather than oscillating conditions on shorter timescales or local melting.

5.3. Original Volume of Deposits

The present volume of deposits considered here is clearly less than their original volume due to the loss of volatiles and redistribution of sediment during the melt back process. The following data are relevant to estimating the original volume of the deposits: (1) the depth of the pits in Cavi Angusti suggests a present thickness in this area of ~1500 m; (2) the depth of the pits in Cavi Sisyphi suggests a present thickness in this area of ~1000 m; (3) the height of the edge of a distinctive pedestal crater near the crater Lau (Figure 2c) suggests that a thickness of about 520 m has been removed [see also Bleacher et al., 2000]; (4) the esker-like features indicate that the volatile-rich deposit formerly overlaid them, implying a previous thickness of at least hundreds of meters; and (5) the topographic relations between the base of the Amazonian-aged layered terrain (Apl) and the top of Hld and HNu in the Cavi Angusti area suggest that the deposit may have an additional thickness of about 1000 m beneath the present polar cap (Apl) (e.g., Plate 3a and Figure 4a, profile 6). These data suggest that the present thickness of the deposit ranges from a minimum of 0 at the margins to possibly as much as 2500 m below the layered terrain (Apl), and that as much as 500 m has been removed from the upper parts of the deposit (for example in the area of the large pedestal crater). Any amount of ice sheet loading and flexure [e.g., Johnson et al., 2000] will represent potential additional volume.

Using the present areal coverage of these deposits described above (~2.94 x 10⁶ km²) and these estimates, we can approximate the original volume of the deposits under the following assumptions: (1) If the deposit averaged 1000 m thick, then the volume would be ~2.94 x 10⁶ km³ (equivalent to a global layer ~20 m thick). (2) If the deposit averaged 2000 m thick, then the volume would be ~5.88 x 10⁶ km³ (equivalent to a global layer ~40 m thick). (3) If the deposit averaged 3000 m thick, then the volume would be ~8.82 x 10⁶ km³ (equivalent to a global layer ~60 m thick). MOLA topography permits an estimate to be made of the volume remaining at present. We used the thickness estimated between the present Amazonian polar cap deposits (Apl and Apl) and the base of the Dorsa Argentia Formation and related units to obtain a reasonable approximation of this present volume. The base of Apl is estimated to lie
at the 1800 m contour, providing a volume of \(-1.33 \times 10^6\) km\(^3\) above this datum (includes both Apl and Api). We estimated the base of the Dorsa Argentea Formation and related units to lie at about the 750 m contour on the basis of the deepest parts of the Schmidt valley and Cavi Angusti. Using the mapped margins of the deposit and this basal contour, we calculated a volume for the deposit above the base of the Dorsa Argentea Formation and below the base of the overlying Amazonian layered terrain. We obtained a volume of \(-2.19 \times 10^6\) km\(^3\) (equivalent to a global layer \(-15\) m thick). These values are estimates of the present volume of the deposits. The original volume was certainly larger, and may have been larger by a factor of 2 or more. Depending on the percentage of sediment contained in the present deposit, the volume of volatiles represented by the remaining material in this unit could range from 0.73 \(\times 10^6\) km\(^3\), if the volatiles form one third of the deposit (equivalent to a global layer \(-5\) m thick), to \(-1.1 \times 10^6\) km\(^3\), if the volatiles form as much as one half of the deposit (equivalent to a global layer \(-7.5\) m thick).

For comparison, the estimated present volume of these deposits \((2.19 \times 10^6\) km\(^3\)) is 1.2 to 1.8 times the present volume of the north polar cap [Smith et al., 1999; Zuber et al., 1998], 1.6 times the present volume of the south polar Amazonian deposits as calculated here, and one sixth of the volume estimated for the area below Contact 2, suggested by Parker et al. [1989, 1993] to be the shoreline of an ancient ocean in the Hesperian period [Head et al., 1999]. This is about 92% of the present volume of the Greenland ice sheet, and about 7% of the volume of the Antarctic ice sheet. Estimates for the total amount of water on Mars are variable and wide-ranging [see Carr, 1996, Figures 7-11]; however, it is clear from these comparisons that the present volume of this deposit \((2.19 \times 10^6\) km\(^3\)) indicates that this reservoir was very significant in the Hesperian, and remains so even today.

5.4. Fate of Volatiles

If the deposit originally contained such large amounts of water, where did the water go? Very deep cavi provide evidence that at least some water reentered the subpolar aquifer, but chasmas and esker-like ridges show that much of the water migrated laterally to the edges of the deposits. The analysis of the distal parts of these deposits using MOLA data shows that sinuous channels emerge from the distal margins of the circumpolar deposits and extend downslope over hundreds of kilometers into adjacent lows, such as the Argyre basin. These channels are interpreted to result from drainage of water from the receding volatile-rich deposit and the associated areas such as the lake apparently occupying the Schmidt valley during the melt back period [Head and Pratt, 2001]. We presently do not have sufficient data to determine the relationship between the amount of water that entered the groundwater table vertically and the amount that was transported laterally away from the margins of the deposits. Two other factors are important. First, sublimation and possibly evaporation must have been important during this melt back period, and because of the widespread nature of the deposit, and its melt back history, a significant amount of water must have migrated into the atmosphere. Second, it seems highly unlikely that the melt back phase could have totally depleted the volatiles from this unit. A significant amount of volatiles must still remain in this deposit and represent a volume permanently removed from the atmosphere and active hydrologic system since the Hesperian Period. This is consistent with recent data from Mars meteorites that suggest that the Martian crust may contain 2-3 times more water than previously thought [e.g., Leshin, 2000].

5.5. Formation of the Volatile-Rich Deposit

On the basis of the ages and geological information about the nature and configuration of the deposits, we conclude that a very volumetrically significant deposit of frozen volatiles and sediment formed in the region of the present south pole in the Hesperian Period, and underwent melt back soon thereafter. On the basis of the thickness and configuration of this deposit and its similarities to the present polar deposits, it seems reasonable to conclude that this ice sheet formed at a cold trap located at the pole.

5.6. Ages

On the basis of an analysis of impact crater size-frequency distribution in the south polar region in the area we have analyzed, Plaut et al. [1988] outlined the following history. Cratered terrain was formed across the entire region from about 4.2-4.0 Ga, followed by the emplacement of plains material which obliterated a large area of cratered terrain and imprinted a zonal asymmetry on the region during the Hesperian \((3.5-0.1\) Ga). Just subsequent to this \((3.5-3.3\) Ga) occurred a period of intense debris deposition; pitted terrain materials (Hd and HNUs) accumulated to a thickness of at least a kilometer in a few 100 Myr, and had essentially stopped accumulating by \(-3.3\) Ga. No discernable obliteration has occurred since formation, but large stripped areas and pits suggest that significant erosion occurred on this material. Plaut et al. [1988] pointed out that the overlying Amazonian layered deposits are thought to have an age of \(-100\) Ma (updated to \(-10\) Ma [Herkenhoff and Plaut, 2000]), but that there are no tight constraints on when they began to form; they proposed that the simple (uniformitarian) model suggests that accumulation has continued from this time until the present. Thus Plaut et al. [1988] documented that there was significant secular variation in the net accumulation of south polar debris, with an early period of enhanced crater obliteration followed by a period of steady state accumulation at a rate 1-2 orders of magnitude lower.

Our analysis leads to the interpretation (Figure 8) that in the early part of the Hesperian (Figure 1b), a massive volatile-rich deposit (ice sheet) extended from near the south pole, where it was thickest, northward to the approximate present boundaries of the Dorsa Argentea Formation. Later in the Hesperian (Figure 8, bottom), large-scale melt back occurred, causing water to migrate to the base of the ice sheet and flow laterally in subglacial channels which deposited sediment on their floors. In other places, such as Sisypheus and Angusti Cavi, large collapse depressions were formed and meltwater migrated both laterally in basal channels and vertically into the subsurface groundwater table. As the ice sheet melted back, the subsurface meltwater channel deposits were exposed, leaving an outwash plain dominated by eskers. In some places the meltwater was contained by regional topography, forming proglacial lakes [Head et al., 2001], such as near the crater Schmidt [Head and Pratt, 2001]. Meltwater breached the barriers at their lowest points around these lakes, forming channels which extended downslope into the Argyre basin. Elsewhere, similar channels drained the meltwater directly into the Argyre basin (Figure 7).

The phase of melt back apparently stabilized, leaving the deposits that are observed today (Plate 1a). The Hesperian age of this activity at the south pole is based on the interpretations of Tanaka and Scott [1987] and Plaut et al. [1988]; a reassessment
of the crater counts is presently under way using the new MOLA data and the interpretative scheme developed here in order to distinguish buried, exhumed, and superposed craters.

5.7. Causes of Melback

What factors might have caused melting and retreat of these volatile-rich deposits? Among the candidate causes are loading-induced basal melting, enhanced geothermal gradient, global climate change, and/or orbital variations. Whatever the mechanism, it needs to be consistent with the formation of a significant volume of volatile-rich deposits (considerably greater than the present polar caps) in the Hesperian Period, the melt back of these deposits slightly later in the Hesperian Period, their stabilization, and their preservation until the present (Figure 8).

We find some evidence for local associated extrusive volcanic activity in the form of edifices that might have erupted below and within the volatile-rich deposit [Ghutan and Head, 2001]. This volcanic activity could be accompanied by an enhanced geothermal gradient that could serve as a heat source for melting of portions of the deposit [e.g., Head and Basilevsky, 2001]. Loading-induced basal melting [Clifford, 1987, 1993] could also have been important, particularly earlier in the history of Mars when thermal gradients were enhanced. Orbital variations (obliquity and eccentricity) are responsible for changes in insolation and, in the case of obliquity variations, perhaps produce alternating periods responsible for the layers in the layered terrain [e.g., Thomas et al., 1992]. Extremes in obliquity variations could also cause larger-scale melting of polar deposits [e.g., Ward, 1992; Lasker, 1989] and thus could be responsible for more regional melting. We have found little evidence for the type of layering typical of the overlying Apl deposits in the Dorsa Argentea Formation or HNx. Although such structure could be destroyed in the melt back phase, its lack may also suggest that the role of obliquity variations was not as significant as it was later in the Amazonian. In addition, obliquity variations would be expected to be repetitive, and unless we are observing the record of only the last melt back phase, the evidence suggests that the extreme melt back was a distinctive one-time event.

Causal factors which are plausible, but for which we have insufficient information to assess, include (1) true polar wander [e.g., Schultz and Lutz, 1988], which could bring volatile-rich polar deposits into a more equatorial region, triggering extensive melting, and (2) major global climate change [e.g., Baker et al., 1991], in which elevation of global atmospheric temperatures could cause a widespread melting event in volatile-rich deposits.

5.8. Geological History of the South Polar Region

Our analysis confirms and elaborates on the general geological history scenario described by Tanaka and Scott [1987] and Plaut et al. [1988] (Plate 1, and Figure 1b). Following formation of Noachian cratered terrain, ridged plains material was emplaced in the Early Hesperian. Immediately following this (~3.5-3.3 Ga) [Plaut et al., 1988] occurred a period of formation and modification of the extensive volatile-rich debris mantle to a thickness in excess of a kilometer. This period of accumulation and modification had essentially ceased by the Middle Hesperian (~3.3 Ga ago) [Plaut et al., 1988]. There is a major hiatus between these deposits and the overlying Amazonian polar layered deposits, thought to have an age of ~100 Ma [Plaut et al., 1988; Herkenhoff and Plaut, 2000].
If polar-like deposits formed here during the Hesperian and again in the Late Amazonian (Apl, layered deposits), it suggests that the present south pole has been a cold trap at least during parts of these two periods. If the time between the Middle Hesperian and the Late Amazonian is indeed a period of nondeposition (Figure 1b), it is not clear why polar deposits did not form here in this time interval, which is possibly as long as several billion years. If they did form, and have been removed, evidence for their former presence does not seem to have been preserved. Alternatively, polar deposit flow might have occurred, obscuring or destroying the cratering record of the intervening period [Head, 2000a, 2000b; 2001].

In the northern hemisphere lowlands, the Vastitas Borealis Formation was emplaced in the Hesperian, and a major hiatus exists between this unit and the Amazonian north polar layered terrain (Apl) (Figure 1b). Tanaka and Scott [1987] found that the Vastitas Borealis Formation is younger than, but may overlap slightly with, the Dorsa Argentia Formation (Figure 1b). Parker et al. [1989, 1993] interpreted the margins of much of the Vastitas Borealis Formation as shorelines of an ancient ocean, and interpreted details of its facies as supporting the presence of an ocean-scale body of water in the northern lowlands at that time [see also Head et al., 1999]. The presence of a significant volatile reservoir at the south pole during the Hesperian in the form of the Dorsa Argentia Formation could have played a very important role in the hydrologic cycle and in recharging the groundwater reservoir [e.g., Clifford, 1987, 1993]. In this scenario [Carr, 1996], water reentering the groundwater system may have recharged the global aquifer sufficiently to initiate the outflow channels and provide water to the southern lowlands [e.g., Baker et al., 1991], known to be a regional topographic low well before that time [Smith et al., 1999; Zuber et al., 1998].

5.9. Implications for the Search for Life

Water is thought to be a fundamental factor in environments favorable to the formation and evolution of life [e.g., McKay and Stoker, 1989; Farmer, 1998; Farmer and Des Marais, 1999]. If life originated in the early history of Mars, then the Dorsa Argentia Formation and related units may provide an excellent location to explore for fossil evidence in any of the deposits representing ancient aqueous environments documented here (e.g., massive volatile-rich deposits, glacial, stream, or lake). These deposits are exposed at and near the surface and should contain a record of aqueous conditions characteristic of earlier Mars history (the Hesperian Period).

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