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Icarus 174 (2005) 444–474
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# Origin and characteristics of the Mars north polar basal unit and implications for polar geologic history

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Received 20 February 2004; revised 16 June 2004

Available online 26 January 2005

**Abstract**

Building upon previous studies, we have used Mars Orbiter Camera and Mars Orbiter Laser Altimeter data to characterize in detail the newly discovered north polar basal unit. Lying stratigraphically between the polar layered deposits, from which it is likely separated by an unconformity, and the Vastitas Borealis Formation, this unit has introduced new complexity into north polar stratigraphy and has important implications for polar history. Exposures of the basal unit in Olympia Planitia and Chasma Boreale reveal relatively dark layers which exhibit differential erosion. Eroded primarily by wind, the basal unit may be the major if not sole source for the north polar dunes and ergs and has contributed material to the lower polar cap layers. We investigate four possible origins for the basal unit (outflow channel/oceanic deposits, basal ice, paleopolar deposits, and eolian deposits). The patchy layering within the unit, its likely sandy grain size, and presence only in the north polar basin suggest that it is primarily an eolian deposit, supporting Byrne and Murray's 2002 earlier conclusion. This implies that at some time during the Early to Late Amazonian, migrating sand was mixed with water ice, forming a relatively dark, sandy deposit. During this time, either no classic polar layered deposits were forming or smaller caps were growing and shrinking, possibly adding material to the basal unit.


*Keywords:* Mars; Mars climate; Mars surface; Geological processes

## 1. Introduction and background

For about 30 years, beginning with Mariner 9 data, scientists have known that the martian polar caps consist of layered deposits (Apl) with less sediment-rich ice layers overlying these (Api) (Tanaka and Scott, 1987) (see Thomas et al., 1992, and Clifford et al., 2000, for reviews). In the north, the Api is probably dirty water ice and almost completely covers the Apl. Hoffman (2000) proposes that $CO_2$ clathrate may be an additional component to the ices at both poles, though Mellon (1996) finds that the phase stability conditions of $CO_2$ and $CO_2$ clathrate may result in only very limited quantities within the polar caps. A thin ($\sim 10$ m) layer of $CO_2$ ice overlies the southern water ice, and these ice deposits cover much less of the Apl than in the north (Tanaka and Scott, 1987; Thomas et al., 2000; Byrne and Ingersoll, 2003). The commonly accepted mode of Apl origin is deposition of varying amounts of ice and dust (and possibly volcanic ash and impact ejecta/glass (Schultz and Mustard, 2004; Wrobel and Schultz, 2004)), with the major layers corresponding to changes in obliquity (e.g., Cutts and Lewis, 1982; Laskar et al., 2002).

Crater dating of the north polar cap surface yields a surface crater-retention age of at most 100,000 years (Herkenhoff and Plaut, 2000), dating it to the Late Amazonian. Martian geologic time is divided into three main periods which have been defined based on the ages inferred from the impact crater density of their geologic type localities (see Tanaka et al., 1992, for review). The most recent estimates by Hartmann and Neukum (2001) of the time spans covered by these periods are: Noachian (4.6 to 3.3 byr ago), Hesperian (3.3 to 2.9–3.2 byr ago), and Amazonian

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0019-1035/$ – see front matter 
doi:10.1016/j.icarus.2004.06.021

(Early: 2.9–3.2 to 1.4–2.1 byr ago; Middle: 1.4–2.1 to 0.3–0.6 byr ago; Late: 0.3–0.6 byr ago to present). The northern polar deposits have been thought to overlie the Hesperian Vastitas Borealis Formation (VBF) which consists of modified volcanics and sedimentary deposits (Lucchitta et al., 1986; Tanaka and Scott, 1987; Fishbaugh and Head, 2000; Kreslavsky and Head, 2002). As is evident in the stratigraphic column of the region published before this study (Fig. 1a), the age of the lower Apl is unknown, leaving an almost 3 byr period of poorly understood geologic history during the Amazonian. A newly discovered basal unit (BU) (Malin and Edgett, 2001; Kolb and Tanaka, 2001; Byrne and Murray, 2002) (Fig. 2), lying stratigraphically between VBF and the north polar cap, may represent the transition between formation of the VBF and Apl and thus could lend important insight into the 3 byr of "missing" north polar history. Since the polar caps to a large degree regulate the martian hydrologic and climatic cycles, the caps' influence on these cycles during the Amazonian is also uncertain.

There are clues as to what may have been taking place during past geologic history of the poles. In the south, Head and Pratt (2001) and Ghatan and Head (2002) hypothesize that the polar deposits were larger during the Hesperian and, due to subglacial volcanic eruption, underwent melting and retreat in the late Hesperian. While polar cap deposition may have been taking place during the Hesperian and earlier in the north, we have found no definite evidence thus far for an ancient polar cap of the size and type we see today. Polar cap deposition would have been punctuated by such events as episodes of possible ocean formation (Parker et al., 1993; Clifford and Parker, 2001; Kargel et al., 1995), volcanic flooding of the northern plains (Kreslavsky and Head, 2002), and high obliquity excursions (Jakosky et al., 1995; Laskar et al., 2004) to name a few. The effect of these events on the formation or ablation of any ancient polar caps is not yet completely understood. Fishbaugh and Head (2000, 2001a) and (Zuber et al., 1998) have shown that the current north polar deposits may once have been larger, extending to about 75° N (Fig. 3). Their retreat has left possible glacial features, such as kame-and-kettle terrain and remnants of polar material, in the arc of low topography south of Olympia Planitia (Fishbaugh and Head, 2001a, 2002a). Olympia Planitia itself was considered to be a remnant of polar material left by this retreat and now covered by sublimation lag reworked into dunes. Still unknown are the timing and cause of this retreat as well as how many times advance and retreat have occurred.

We have previously discussed three possibilities for Amazonian polar cap history (Fishbaugh and Head, 2001a). (1) The current polar deposits may represent young deposits having undergone one stage of retreat. This would imply that climate conditions during the Early and possibly much of the Middle Amazonian may not have been conducive to forming a classic northern polar cap. (2) Deposition of the current cap began in the Early Amazonian, having undergone at least one stage of partial retreat. In this case, some

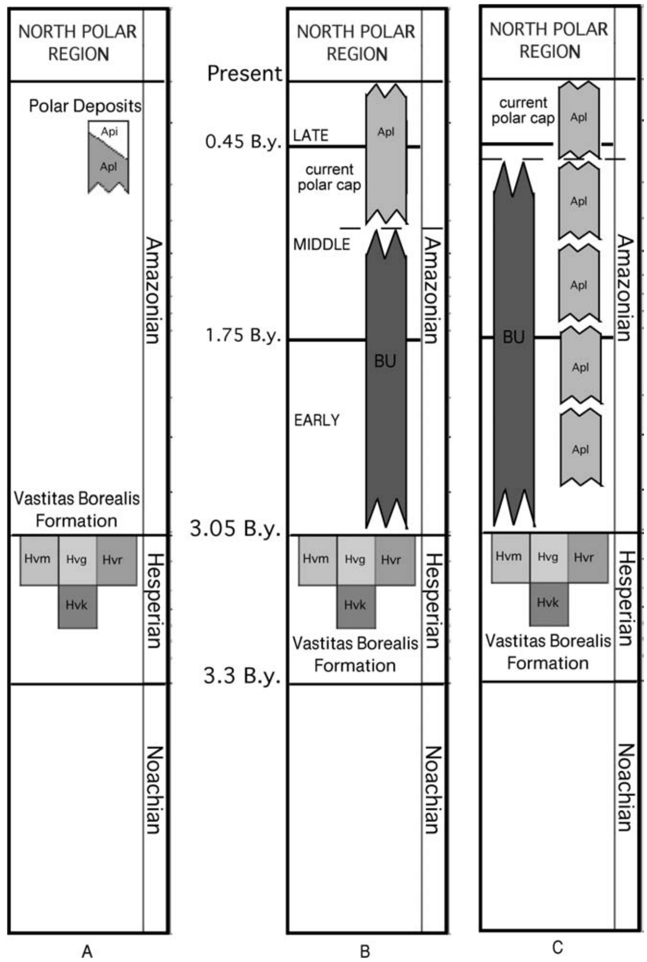



Fig. 1. Approximate stratigraphic columns for the martian north polar region. (A) Stratigraphic column from Tanaka and Scott (1987). The authors map the residual ice on top of the polar layered terrain (Apl) as a separate unit (Api). Additionally, no times are specified for the boundaries between the Noachian, Hesperian, and Amazonian. The beginning deposition time of the Apl is unknown and hence is indicated by a saw-tooth line. (B) A revised stratigraphic column according to this study which represents the possibility that the polar deposits formed after the basal unit but yet still may have partially retreated one or more times. Since it is currently widely believed that the Api merely represents the topmost layer of the Apl, these units have not been separated. Again, saw-tooth lines represent the beginning and end of polar deposit and BU deposition since these times are unknown. The dashed line represents an unconformity of unknown time length between the BU and polar deposits. The dates indicated represent the middle of the ranges given by Hartmann and Neukum (2001). The designation "current polar cap" refers to the polar cap which exists on the surface today. (C) A revised stratigraphic column according to this study which represents the possibility that polar deposits (albeit much smaller ones than that which exists today) formed throughout the Amazonian and may have even contributed to basal unit material. We have indicated several possible episodes of polar cap deposition and complete sublimation, though the actual number is completely unknown as is the number of times the caps underwent partial versus complete sublimation. Since lenses of polar layered deposits have not been found within the BU, it is likely that any small polar caps which co-existed with the BU completely sublimated. Again, saw-tooth lines represent the beginning and end of polar deposit and BU deposition since these times are unknown.

process (such as relaxation and flow during times of warmer temperatures (e.g., Pathare and Paige, 2005) or resurfacing due to deposition) has erased craters, effectively resetting the surface crater age. (3) The current cap may be the latest manifestation of caps that have grown and partially ablated since the Early Amazonian. In this case, expansion and retreat may be regulated by orbital cycles. The bottommost layers of the Apl could be Early Amazonian remnants of the oldest phase of polar deposition. Since the BU lies between the VBF and the lower Apl layers and records evidence of past stages of polar region geologic activity, investigation of its depositional history is crucial to understanding of polar cap history.

While not expounded upon in detail, according to Kolb and Tanaka (2001) this dark, platy unit represents "an earlier phase of north polar deposits" (p. 30). Byrne and Murray (2002) propose such a significant change in deposition style between the BU and Apl that the unit must represent a period in time when there was no polar cap. They believe that the BU consists of ice-rich paleoerg deposits brought by atmospheric means to the low elevation plains.

In this study, we expand upon the previous work done by Byrne and Murray (2002) and Edgett et al. (2003) in describing the main features of the BU using primarily MOC images and MOLA data, and we analyze outcrops of the BU and the unit's distribution in detail. We examine four possible initial formation mechanisms for the BU: (1) outflow channel/oceanic deposits, (2) basal ice, (3) paleopolar cap (mentioned by Kolb and Tanaka (2001) and Tanaka et al. (2003)), and (4) eolian deposit (Byrne and Murray, 2002). Finally, we discuss our findings in the context of the geologic history of the polar regions.

## 2. Observations

### 2.1. Methods of observation

Byrne and Murray (2002) have identified the BU in the walls of Chasma Boreale and in the troughs near and extending into Olympia Planitia (Fig. 3). We have extended this search by examining in great detail all available MOC images (through the E07–E12 release, January 2002) of those regions and many images in other locations: other troughs, the cap periphery, outlying remnants of polar material, and the region south of Olympia Planitia thought to contain features created by cap retreat (Fishbaugh and Head, 2000, 2001a, 2001b, 2002a). Later image releases have also been examined, though for our purposes these images did not cover significantly much more area with significantly much more detail than the previous image releases. We have recorded the presence of particular features such as deformation, presence of dunes, mass wasting, detailed layer structure, and pits (and/or impact craters), and notes on the characteristics evident in each image.

Using the GIS program, ESRI Arcview, we have mapped the recorded features on MOLA shaded relief topography, the USGS geologic map (Tanaka and Scott, 1987), and the MOC north polar image mosaic. These maps reveal patterns in the features of the BU. An example of one of these maps is shown in Fig. 4 which reveals that the BU is indeed exposed in cross-section view primarily within the Olympia Planitia troughs and Chasma Boreale.

Having chosen particular exposures of the BU which best typified its characteristics in MOC images, we have created detailed sketch maps. With these sketch maps we have noted detailed characteristics of the layers within the BU and of the BU/Apl contact. We have also aligned MOLA profiles with each sketch map and marked off unit boundaries on the profiles, creating approximate geologic cross-sections (without dip angle information) so that we could cross-correlate units from different parts of the polar cap. We describe examples of these particular outcrops below.

### 2.2. Polar layered deposits

The polar layered deposits (Apl) (Fig. 5) consist of alternating layers of more ice-rich and more sediment-rich composition created by deposition of ice/dust mixtures with ice/dust ratios varying with climatic cycles (Murray et al., 1972; Soderblom et al., 1973; Howard et al., 1982; Cutts and Lewis, 1982). The darker layers within the Apl may be laterally continuous and contain an unknown percentage of particulates. While Viking color data indicate that reddish planetary dust makes up part of the Apl particulates, Thomas et al. (1992) speculate that dark material like that found in the polar ergs is also mixed in with this dust and reduces calculated reflectance values. Small amounts of particulates can greatly decrease the albedo of ice if the ice grain size is larger than the particulate grain size so that using the albedo of the darker Apl layers does not reveal the true dust content (Kieffer, 1990). In addition to annual deposition of various amounts of dust and frost, global dust storms, spread of impact ejecta (especially by Coriolis forces causing collection of ejecta at the pole (Wrobel and Schultz, 2004; Schultz and Mustard, 2004)), and volcanic eruptions (provided that some layers are old enough to have existed when the planet was more volcanically active) may add to the Apl layers. Since much of the material which has contributed to the Apl (except for impact ejecta and glass) is transported by atmospheric suspension, it is unlikely that sand-sized particles are common within the Apl.

Layer thicknesses vary. Viking observed major layers on the order of 5–25 m thick (Blasius et al., 1982), while MOC images have revealed layers down to the limit of camera resolution (about 1.5 m/pix in this region) (Malin and Edgett, 2001; Milkovich and Head, 2003). Accumulation rates are estimated to be on the order of millimeters per year (Jakosky et al., 1993, 1995; Laskar et al., 2002; Greve et al., 2003), so that there may exist annual layers of a few millimeters thickness.

Sublimation causes most of the erosion of the Apl, though Howard (2000) has shown that wind also contributes to Apl erosion as evidenced by frost streaks and undulations; wind

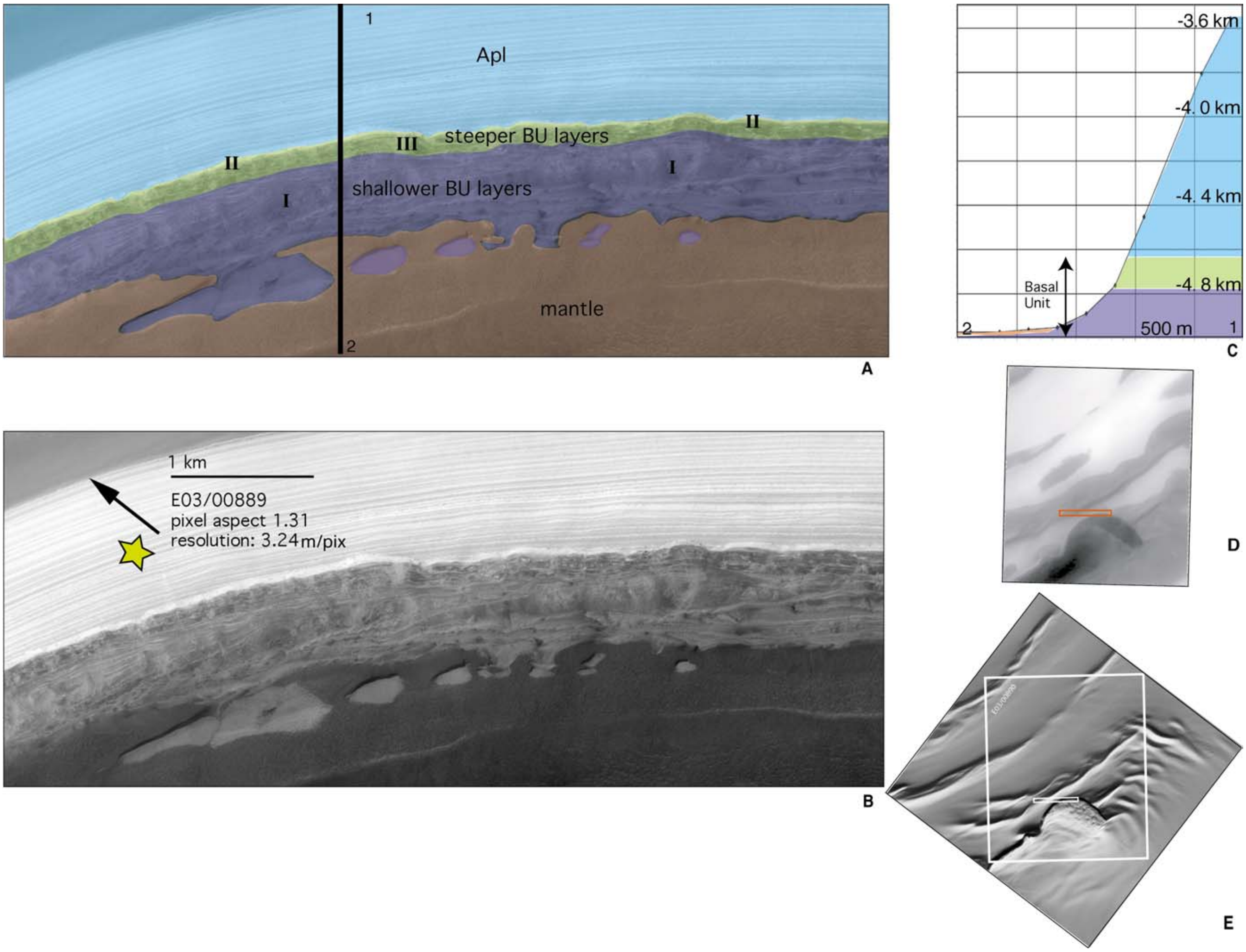



Fig. 2. Example of a basal unit outcrop in Chasma Boreale. (A) Geologic interpretation of image. Roman numerals refer to features described in text. Labeled geologic units are also described in text. The black line indicates location of the MOLA profile used for the geologic cross-section. (B) Portion of MOC image E03/00889. The arrow indicates solar azimuth. See Fig. 4 for location of this MOC image. (C) Approximate geologic cross-section corresponding to the geologic interpretation in (A) and using a MOLA profile. Note that dip angle of the geologic contacts is not indicated. This particular MOLA profile (#11685) corresponds to image M03/04769 which overlaps the MOC image in this figure. (D) Context image is MOC wide angle image E03/00890. Red rectangle shows location of *entire* narrow-angle MOC image. (E) MOLA shaded relief corresponding to context image in (D).

may also play a large role in the formation of the polar troughs which expose Apl layers. There is some evidence of differential erosion of the layers within the Apl. Some layers protrude from the others, some recede, and others are pitted (Fig. 5). Outgassing of exposed $CO_2$ clathrate or $CO_2$ ice may have created some of this pitting (Milkovich and Head, 2002). In Fig. 5, these layers with small pits lie beneath a layer characterized by small ridges; this latter layer has been termed a "marker bed," because it is easily recognized in several locations throughout the Apl (Malin and Edgett, 2001; Kolb and Tanaka, 2001). Other evidence of erosion comes in the form of mass-wasting. In many MOC images, we have observed bright streaks of material emanating from the Apl (V in Fig. 6). Talus from mass-wasted Apl material may also have built up along the BU/Apl contact in Fig. 2 as noted by Edgett et al. (2003). As we discuss below, bright material in the basal unit also appears to be undergoing mass-wasting, producing bright streaks similar to those produced by the Apl.

### 2.3. *The basal unit*

#### 2.3.1. *The basal unit in Olympia Planitia*

The most obvious difference between the basal unit and Apl is the lower albedo of the BU, yet both are layered. Layering in the BU itself has been described as thick, irregular, and platy (Byrne and Murray, 2002) with some beds showing more resistance to erosion than others (Edgett et al., 2003). Further investigation in this study reveals that the details of layering in the BU are much different than in the Apl.

Figure 6 illustrates a typical example of basal unit layers exposed in Olympia Planitia. The layering within the unit is complex, with no single layer extending across the entire image (a distance of about 3.3 km). This particular outcrop

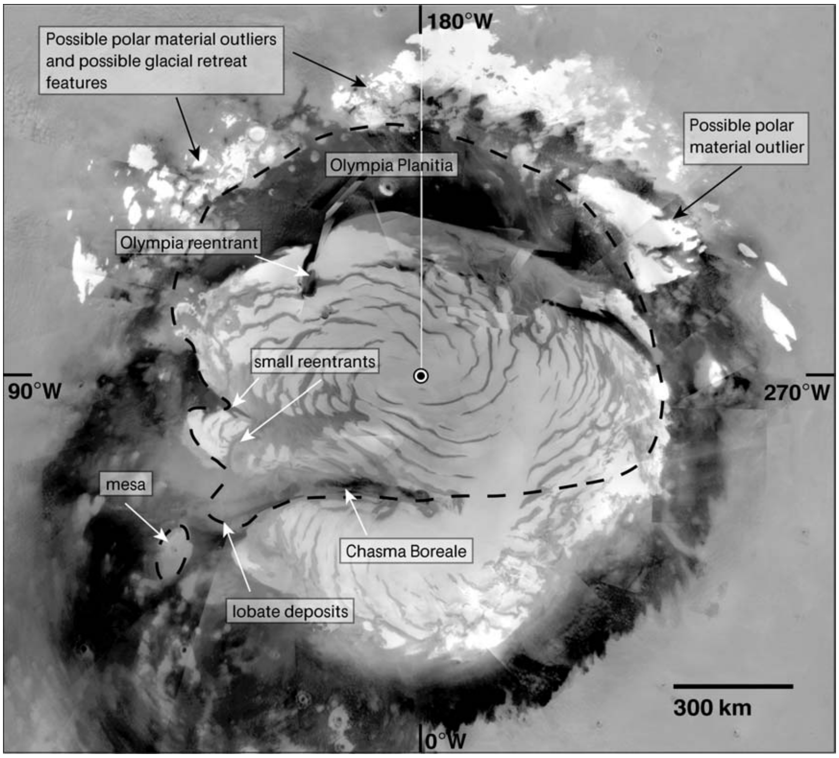



Fig. 3. MOC wide-angle image mosaic of the north polar region (75° N to pole) labeled with features and areas mentioned in the text. Dotted line shows approximate extent of the basal unit in map view. Note that the inclusion of the cratered mesa beyond Chasma Boreale is less certain than the area within the rest of the outline (see text for details).

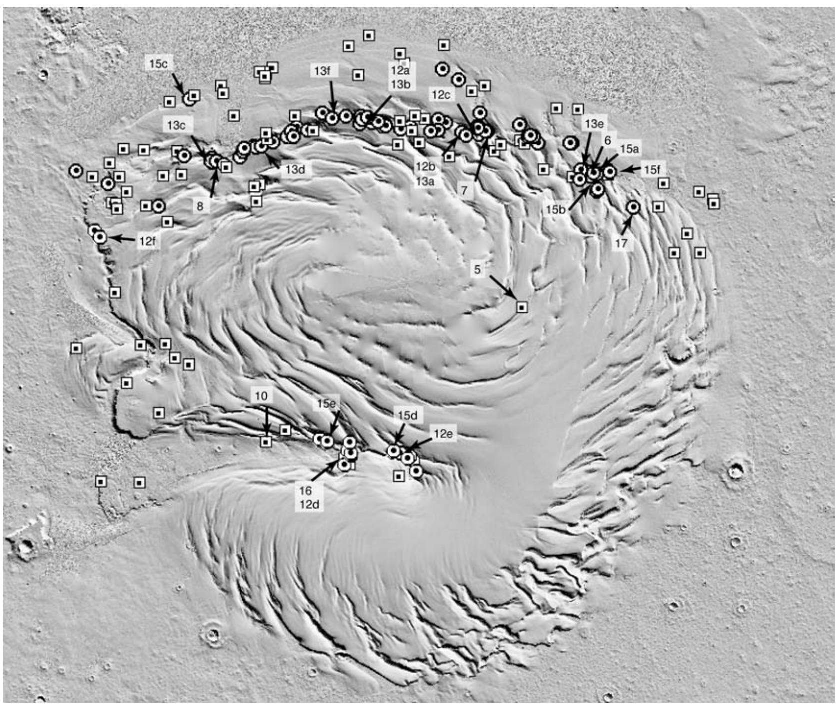



Fig. 4. MOLA shaded relief showing locations of MOC images analyzed in detail for this study and which contain a possible exposure of the BU. ⊙ = approximate location of the center of a MOC image containing a definite basal unit outcrop. ⊡ = approximate location of the center of a MOC image which may contain a basal unit outcrop. This map does not indicate the location of poor quality images or images analyzed which did not contain BU outcrops. We analyzed all available MOC images of the region south of Olympia Planitia containing polar material remnants and retreat features as well as many images in the cap margin area east of Chasma Boreale and extending to Olympia Planitia. No obvious evidence of the basal unit was found in these areas. Numbers indicate locations of MOC images shown in figures of the same number.

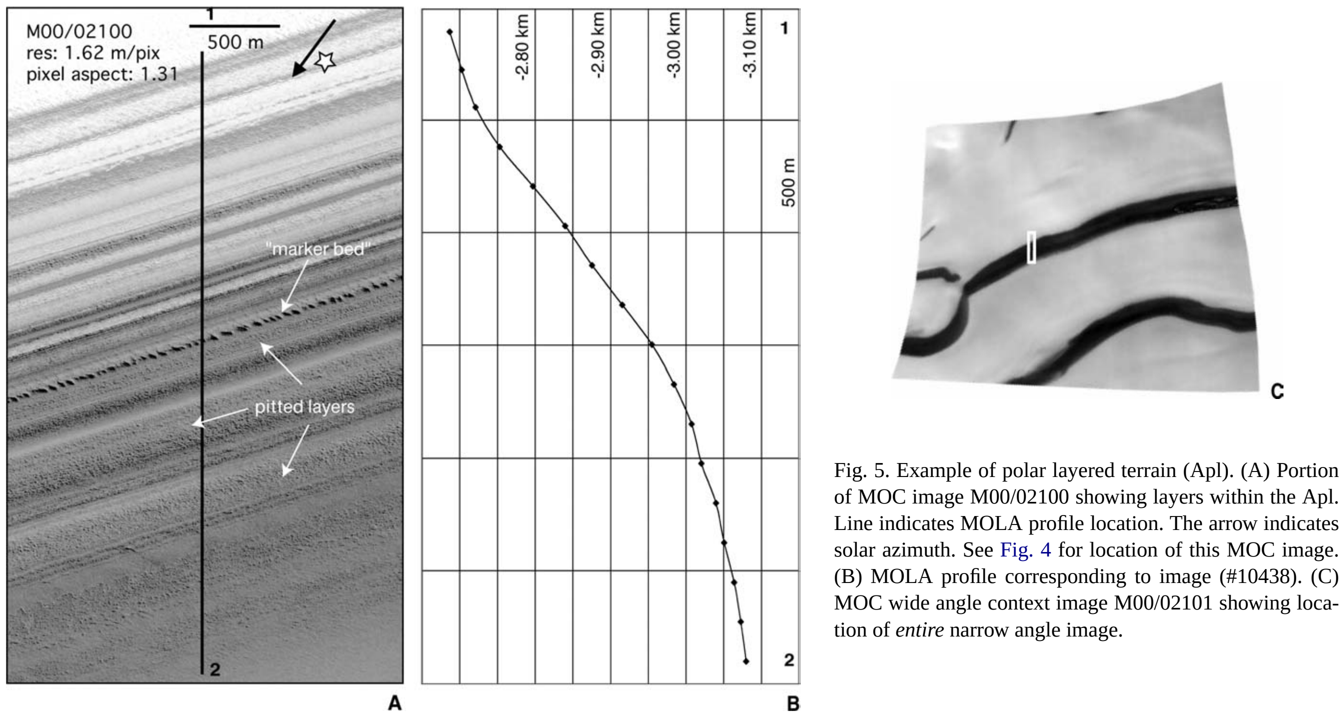



Fig. 5. Example of polar layered terrain (Apl). (A) Portion of MOC image M00/02100 showing layers within the Apl. Line indicates MOLA profile location. The arrow indicates solar azimuth. See Fig. 4 for location of this MOC image. (B) MOLA profile corresponding to image (#10438). (C) MOC wide angle context image M00/02101 showing location of *entire* narrow angle image.

exhibits 4 main sequences. At the bottom of the image is a rough-textured, dark mantle. This texture could be a result of the presence of hummocks, boulders, or small dunes, and the mantle may represent the lag left by basal unit erosion. Above this in the image lie alternating bands of bright and dark material, and above this is the brighter Apl.

Within the alternating bright and dark bands of the BU is a section just below the Apl with steeper slopes (about 40°) than the more exposed basal layers below it. There are about seven dark layers in this 180 m thick section; therefore, each dark layer has an apparent thickness of about 25 m. Because individual layers may be of a thickness bordering on or less than the limit of resolution, each layer cannot be correlated with layers in other images. However, we have observed similar layer morphology throughout the basal unit both in Olympia Planitia and within Chasma Boreale (e.g., Figs. 6, 7, 11, 12, and 13).

Eolian reworking is evident in many BU outcrops. In Fig. 6, the lowest BU layers are characterized by lineations likely of eolian origin which may either be yardangs or longitudinal dunes (I). Darker layers stratigraphically above this show no such evidence of eolian reworking. While ridges of unknown origin (III) appear in a few of the upper dark layers, they are for the most part featureless. The eolian lineations are visible through the overlying bright material (II), making it likely that the lighter material accumulations are much thinner than the darker, and in some places this light material even appears patchy on top of the darker. Without high resolution topographic data able to distinguish between individual layers, it is impossible to determine whether this bright material makes up entire layers or just accumulates on flat portions of and small terraces in the dark layers. Irrespective of image resolution, while the Apl also exhibits some layering at a similar scale, the brighter Apl layers can be as thick or thicker than the darker layers. The Apl layers are also more continuous and can be traced along the entire width of the image.

At the base of the lowest exposed BU layer, there exists a narrow, bright halo of material (IV). Either this represents (1) a very thin layer or (2) material which has mass wasted from above and collected here, possibly eroded from the brighter layers. Note its diffuse, irregular nature. Bright streaks (V) cross-cut many of the layers just below the Apl, and there is little to no evidence of mass wasting of the darker material. While mass wasting of dark material is apparent in some other images of the basal unit randomly scattered throughout the unit exposures, it is not common.

Further east in Olympia Planitia, the basal unit is exposed beneath the Apl and within a small, shallow trough (I) (Fig. 7). Again, light and dark layers are exposed in the steeper section (slope $\approx 45°$), and each dark layer has an apparent thickness of about 20 m, similar to those in Fig. 6. Within this image are mantles with varying degrees of smoothness, like that in Fig. 6. Dark material (II) is eroding from layers exposed in the trough and within parts of the steeper layers. Part of this dark material has formed a large dune (III) at the base of the steeper basal unit layers. Therefore, this dark material consists of sand-sized particles. As described below, dunes appear to emanate from the basal unit in several other locations.

Opposite to the Apl/BU exposure in Fig. 7 is a bright region containing a few dark layers (IV). At first glance, this may appear to be a shallow-sloped exposure of the Apl. However, the cross-section shows that this deposit lies at the same elevation as the steeper section of the basal unit (assuming horizontal bedding) and thus is probably composed

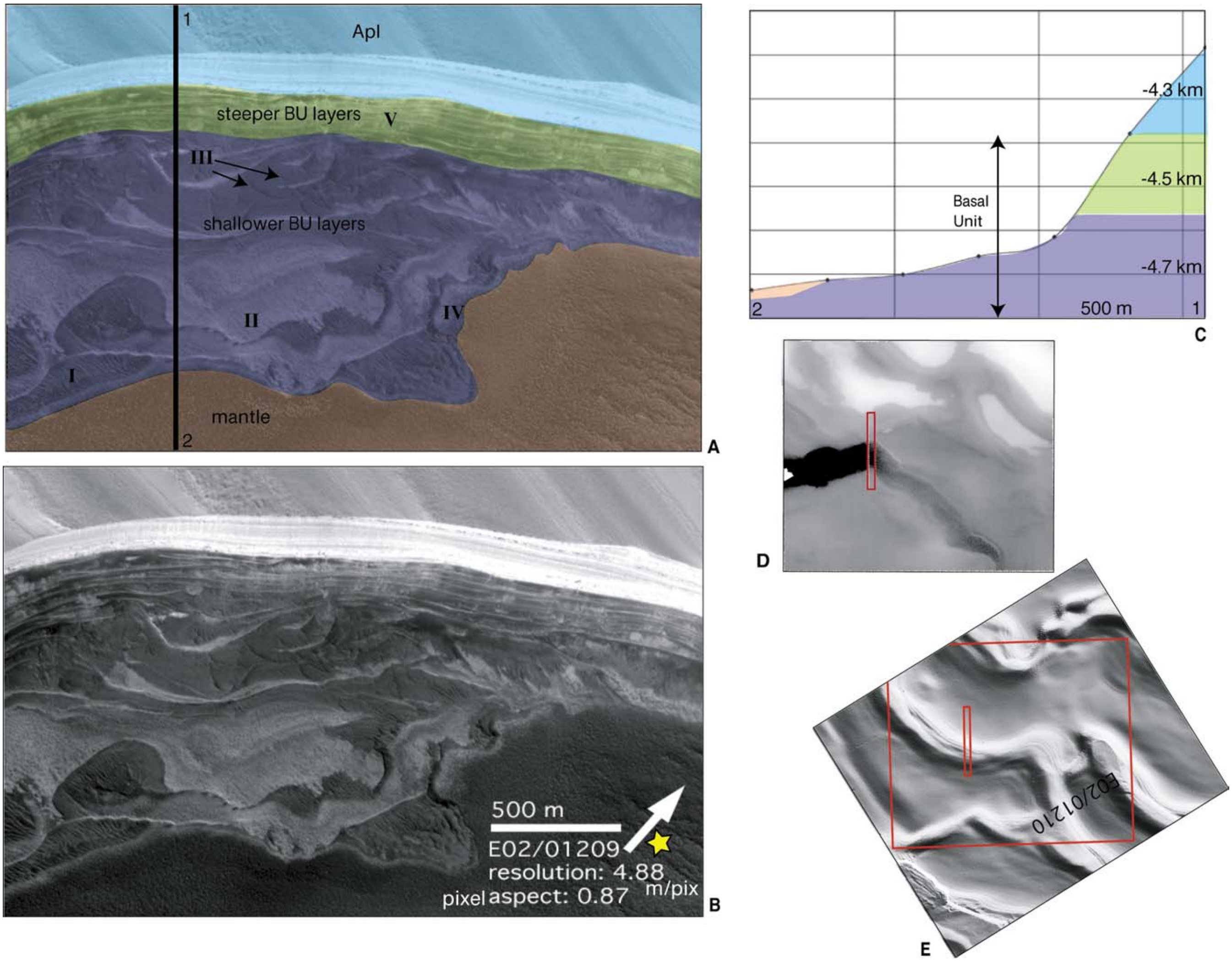


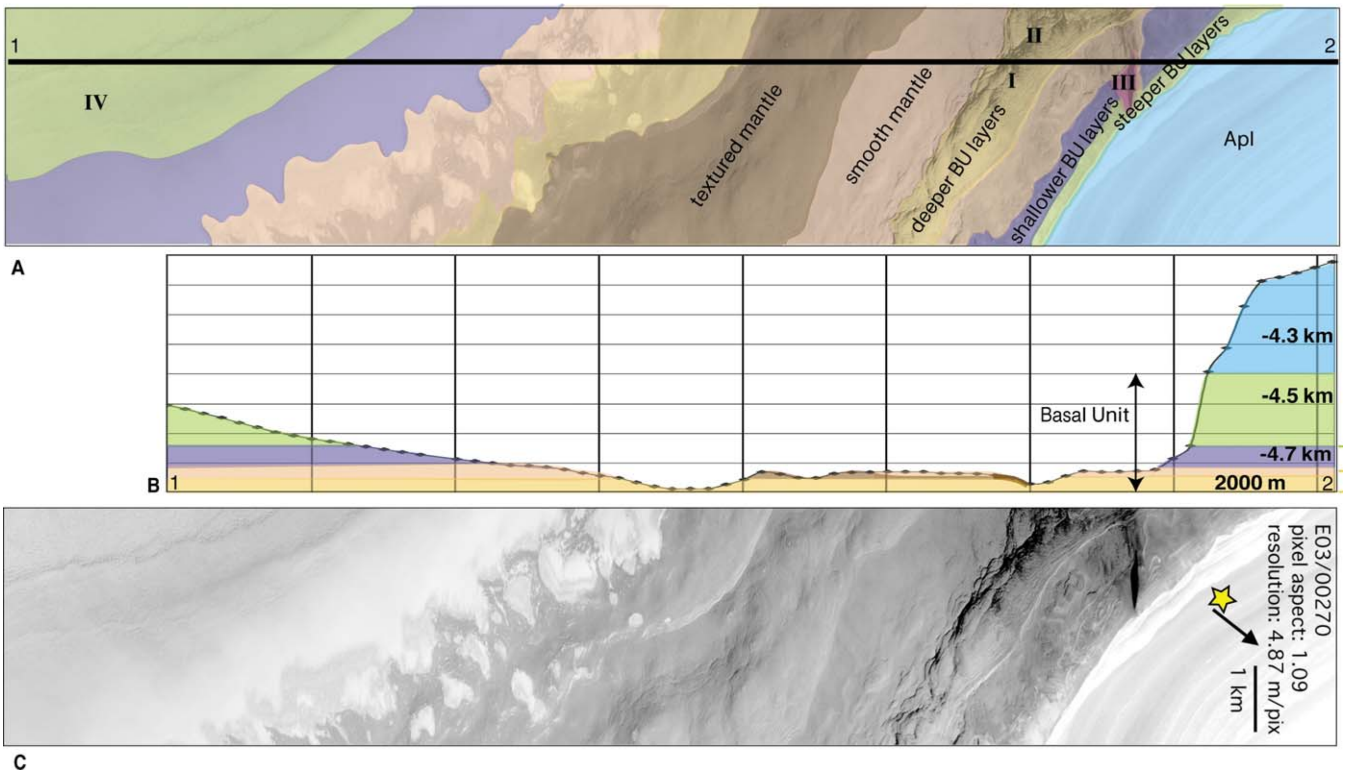

Fig. 6. Example of basal unit outcrop in Olympia Planitia. (A) Geologic interpretation of image. Roman numerals refer to features described in text. Labeled geologic units are also described in text. The black line indicates location of the MOLA profile used for the geologic cross-section. (B) Portion of MOC image E02/010209. The arrow indicates solar azimuth. See Fig. 4 for location of this MOC image. (C) Approximate geologic cross-section corresponding to the geologic interpretation in (A) and using a MOLA profile (#19016). Note that dip angle of the geologic contacts is not indicated. (D) Context image is MOC wide angle image E02/01210. Red rectangle shows location of *entire* narrow angle MOC image. (E) MOLA shaded relief corresponding to context image in (D).

of eroded BU layers covered by frost. It is important to note that one must be careful not to immediately interpret MOC images in the polar region without comparison with the associated MOLA data.

Proceeding further east, now near the small reentrant in Olympia Planitia, Fig. 8 lies within a trough. The BU outcrops beneath the Apl and in a large mesa in the center of the image. The layer thicknesses in this image appear to be 50 m, or twice that seen in Figs. 6 and 7. The scarp in which the layers are exposed may be steeper than in previously discussed images, making it difficult to see thinner layers. Some of the smaller scale scalloping and texture seen within the large BU layers in this image may therefore actually be smaller layers of the thickness seen elsewhere (e.g., in Figs. 6, 7).

The layer edges in this image appear scalloped (I in Fig. 8a) due to the fact that layering is actually discontinuous and patchy. The top BU layer just beneath the Apl is heavily pitted (II in Fig. 8a). This type of concentrated pitting appears only within BU exposures near the Olympia Planitia reentrant (Fig. 3). Therefore, the geologic sketch map shows the BU in a different color from that in other images (e.g., in Figs. 6, 7). However, the elevations at which this unit is exposed are similar to those at which the basal unit is exposed elsewhere, making it probable that this unit is also the basal unit. Small pits (diameter < 300 m) can be found in several other MOC images of the BU where erosion of overlying layers has exposed older layers beneath but are not as numerous as in this region; an impact origin for these features is uncertain.

Focusing in more closely on the pitting and scalloping in the upper BU layers (Fig. 8f), the Apl layers undulate just above the BU here (III). These undulations dampen out further up section within the Apl (IV). Undulations on the BU



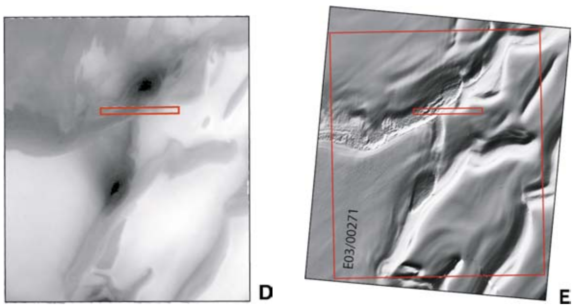




Fig. 7. Example of basal unit outcrop in Olympia Planitia. (A) Geologic interpretation of image. Roman numerals refer to features described in text. Labeled geologic units are also described in text. The black line indicates location of the MOLA profile used for the geologic cross-section. (B) Approximate geologic cross-section corresponding to the geologic interpretation in (A) and using a MOLA profile (#19266). Note that dip angle of the geologic contacts is not indicated. (C) Portion of MOC image E03/00270. The arrow indicates solar azimuth. See Fig. 4 for location of this MOC image. (D) Context image is MOC wide angle image E03/00271. Red rectangle shows location of *entire* narrow angle MOC image. (E) MOLA shaded relief corresponding to context image in (D).

(V) correlate with these Apl undulations. The pitting itself is probably erosional, but the undulations in both the BU and Apl can be interpreted in several ways. (1) The Apl layers were deposited on the eroded, irregular topography of the BU, and thus the lower Apl layers undulate (Fig. 9a). Note that these undulations damp out up section (IV in Fig. 8f). (2) Erosion of the Apl layers has led to the appearance of undulations from above, so that the "undulations" are really oriented as shown in Fig. 9b, in-and-out of the page. (3) The lower Apl layers have deformed (and are oriented as in Fig. 9a) possibly due to ice flow over the irregular contact with the BU. Due to the fact that little to no evidence of plastic deformation has yet been discovered elsewhere within the northern Apl, that the layers at (IV) in Fig. 8f do not undulate while those just below it at (III) do, and that the wavelength of undulation does not appear to correlate with layer thicknesses or particulate contents (as one might expect with deformation), we believe that the undulations in the Apl are probably erosional in nature (Fig. 9b) rather than due to folding. However, it is also possible that deformation has taken place because of differing rheologies across the interface between the Apl and BU (assuming that ice flow has occurred). Distinguishing between the situations illustrated by Fig. 9 with more certainty would be beyond the scope of this paper as it would require further detailed investigation of similar places along the BU/Apl contact together with ice flow modeling.

### 2.3.2. *The basal unit in Chasma Boreale*

The best cross-sectional exposures of the basal unit lie within the arcuate scarps at the head of Chasma Boreale. Shallower wall slopes elsewhere along the Chasma walls may increase the difficulty of recognizing the basal unit there.

In addition to the alternating light and dark bands typical of the BU, Byrne and Murray (2002) and Edgett et al. (2003) have described the BU as having a thick-layered, platy appearance in some locations. Near the center of the BU in the Fig. 2 outcrop (within Chasma Boreale) are what appear to be massive beds or large lenses of material (I), leading

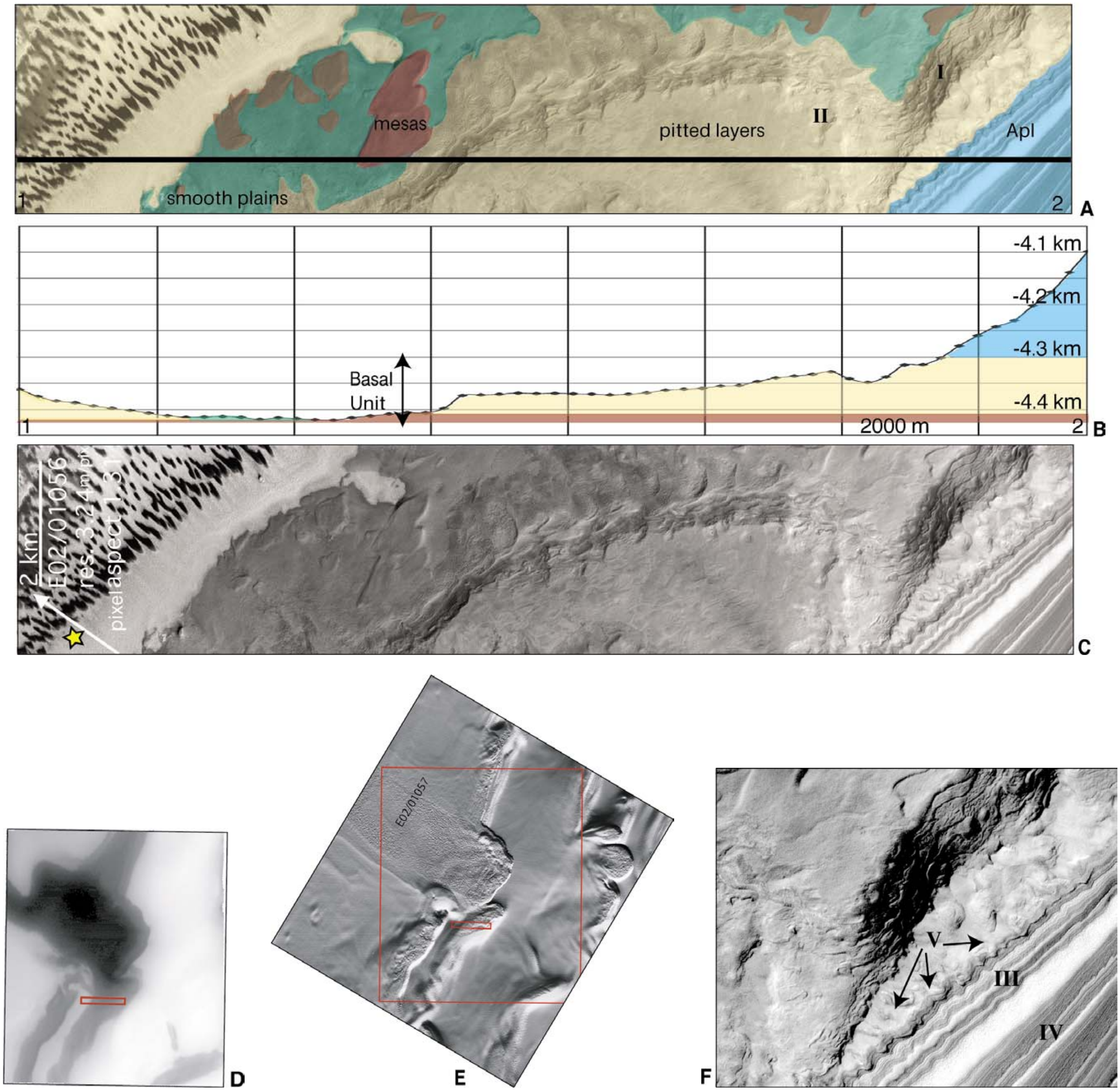



Fig. 8. Example of basal unit outcrop in Olympia Planitia. The pitting evident in the BU layers is found in images lying near the Olympia Planitia reentrant. (A) Geologic interpretation of image. Roman numerals refer to features described in text. Labeled geologic units are also described in text. The black line indicates location of the MOLA profile used for the geologic cross-section. (B) Approximate geologic cross-section corresponding to the geologic interpretation in (A) and using a MOLA profile (#18995). Note that dip angle of the geologic contacts is not indicated. (C) Portion of MOC image E02/01056. The arrow indicates solar azimuth. See Fig. 4 for location of this MOC image. (D) Context image is MOC wide angle image E02/01057. Red rectangle shows location of *entire* narrow angle MOC image. (E) MOLA shaded relief corresponding to context image in (D). (F) Close-up of lower right hand portion of MOC image in (C). Roman numerals refer to features described in text.

Edgett and Malin (2003) to conclude that the these represent a unit separate from the thinner BU layers. However, MOLA data reveal this to be a slope effect. The lenses are just layers exposed at shallower slopes than the other layers, again indicating differential erosion within the BU. Layers exposed within the steeper section (slope $\approx 45°$) of the BU in Fig. 2 are slightly thinner in some places (15 m here) than in Olympia Planitia. Since the BU profile shows the unit to pinch out near Chasma Boreale, as discussed below, layer thinning is expected. However, it is important to note these are only apparent layer thicknesses. Again, a dark mantle lies at the base of the exposed BU as seen in Olympia Planitia. There are no apparent, major differences between layering within the BU in Olympia Planitia and in Chasma Boreale.

### 2.3.3. *Basal unit surface exposures*

Thus far, we have described the basal unit only in cross-section. Identification of the BU in planform or on shallow

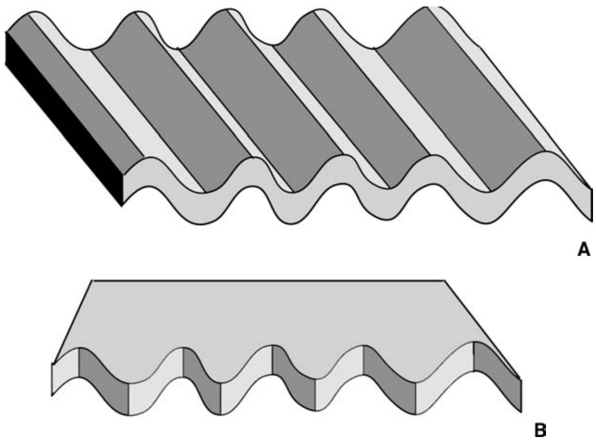



Fig. 9. Possible orientations of the wavy Apl layers in Fig. 8 (III in Fig. 8f). (A) Vertical orientation caused by deposition on irregular topography or by folding. (B) Horizontal orientation caused by erosion.

slopes is more difficult as the alternating light and dark layer pattern and sharp contrast with the overlying Apl is not always visible. In addition, very little of the basal unit surface is exposed.

Figure 10 displays an example of what is probably the BU exposed on a shallow slope near the middle of Chasma Boreale. Notice the dark layers and erosional pitting. There are few exposures of this type in the MOC images we analyzed for this study. In many places, a dark mantle lies at the base of the BU (e.g., Fig. 2). It is probable that this mantle consists of eroded BU material and masks the surface of the BU beneath it. Dune patches also hide much of the BU surface.

The floor of Chasma Boreale may reveal the largest exposed surface area of the basal unit. As discussed below, the BU pinches out somewhere near the Chasma and is exposed within the arcuate scarps in the head region. Since the exposed BU is about 350 m thick at the Chasma head (Fig. 2), a large amount of BU material has been removed by chasma formation. We do not know how the thickness of the BU changes further down chasma, so the exact volume of removed material is difficult to determine. However, if we assume that the BU was 350 thick throughout the length and width of the Chasma before erosion and had a rectangular cross-section, then the volume of material removed is about $1 \times 10^4$ km$^3$. Thus, whatever eroded the chasma, be it katabatic winds (Howard, 2000), melting (Clifford, 1987; Benito et al., 1997), or a combination of the two (Fishbaugh and Head, 2002b) has stripped away a large amount of the basal unit and revealed lower BU layers. It is important to keep in mind that while melting may have removed much of the BU in this area, later katabatic wind action may have removed even more layers, thus erasing some of the record of outflow. What is left on the Chasma floor is a surface covered in many places by dunes and dark mantle material. There are three craters on the floor of Chasma Boreale that are 10–20 km in diameter and have been exhumed by chasma formation, but there are few small craters. Edgett et al. (2003) have indicated that remnants of the BU may lie on the chasma floor. Figure 11 shows an example of a layered mesa lying on the chasma floor near an exposure of basal unit layers. This mesa may be a remnant of the BU.

The BU may also have affected the possible outflow which initiated the formation of Chasma Boreale (Benito et al., 1997; Fishbaugh and Head, 2002b) by influencing where Chasma Boreale formed (since the BU ends near Chasma Boreale) and/or by acting as a reservoir for the water as suggested by Byrne and Murray (2002).

#### *2.3.4. Layering within the basal unit*

Thus far, we have discussed the characteristics of the BU within its main exposures in Olympia Planitia and Chasma Boreale and the similarities and differences between its expression in these different locations. Here we focus on the range of layering characteristics observed, regardless of location.

In Fig. 12 are shown six examples of layer properties within the BU. Alternating bright and dark bands are common throughout the BU ((a), (b), (d), (e)). In some places, the thickness of visible layers is similar in the Apl and BU (b). The thickness of layers is not uniform throughout the BU, though the major layers have an apparent thickness of tens of meters thick (see discussion in Sections 2.3.1 and 2.3.2). As shown in Fig. 12c, thinner layers can be enclosed within thicker layer packages. Layering tends to be highly eroded (a), nearly planar ((b), (e)), irregular ((e), (f)) or even massive in appearance (f), though the latter may be due to lighting conditions and lower image resolution. There are few if any outcrops of the BU wherein a single layer can be traced along the entire outcrop within one image. Thus, unless major erosion took place every time a layer was formed, it seems that deposition of the BU was patchy and discontinuous. Small patches of material were deposited in an overlapping fashion.

Eolian erosion within layers is common (Figs. 12a, 12c, and 12f; and I in Fig. 6a), thus the material is easily eroded and reworked by wind and probably has a sandy grain size. The compositional contrast between the Apl and the BU has resulted in differential erosion between the two units. Erosion of the BU is typically characterized by pitting, residual mesas (e.g., Fig. 8), and dark mantles (e.g., Fig. 6). Since in most locations, the BU does not undercut the Apl, but rather protrudes from beneath it (even exists without Apl on top of it, Fig. 12d), it is probably more resistant to erosion than the Apl.

### *2.4. The Upper and lower BU contacts*

#### *2.4.1. Upper (BU/Apl) contact*

Most notably, the contact between the Apl and the BU exhibits a stark contrast in albedo (e.g., Fig. 2). While Byrne and Murray (2002) noted that this contact is in some places manifested as a protruding step, not all of the contact is associated with a major break in slope (e.g., Figs. 2, 6, 8). In places the contact appears distinct and sharp. How-

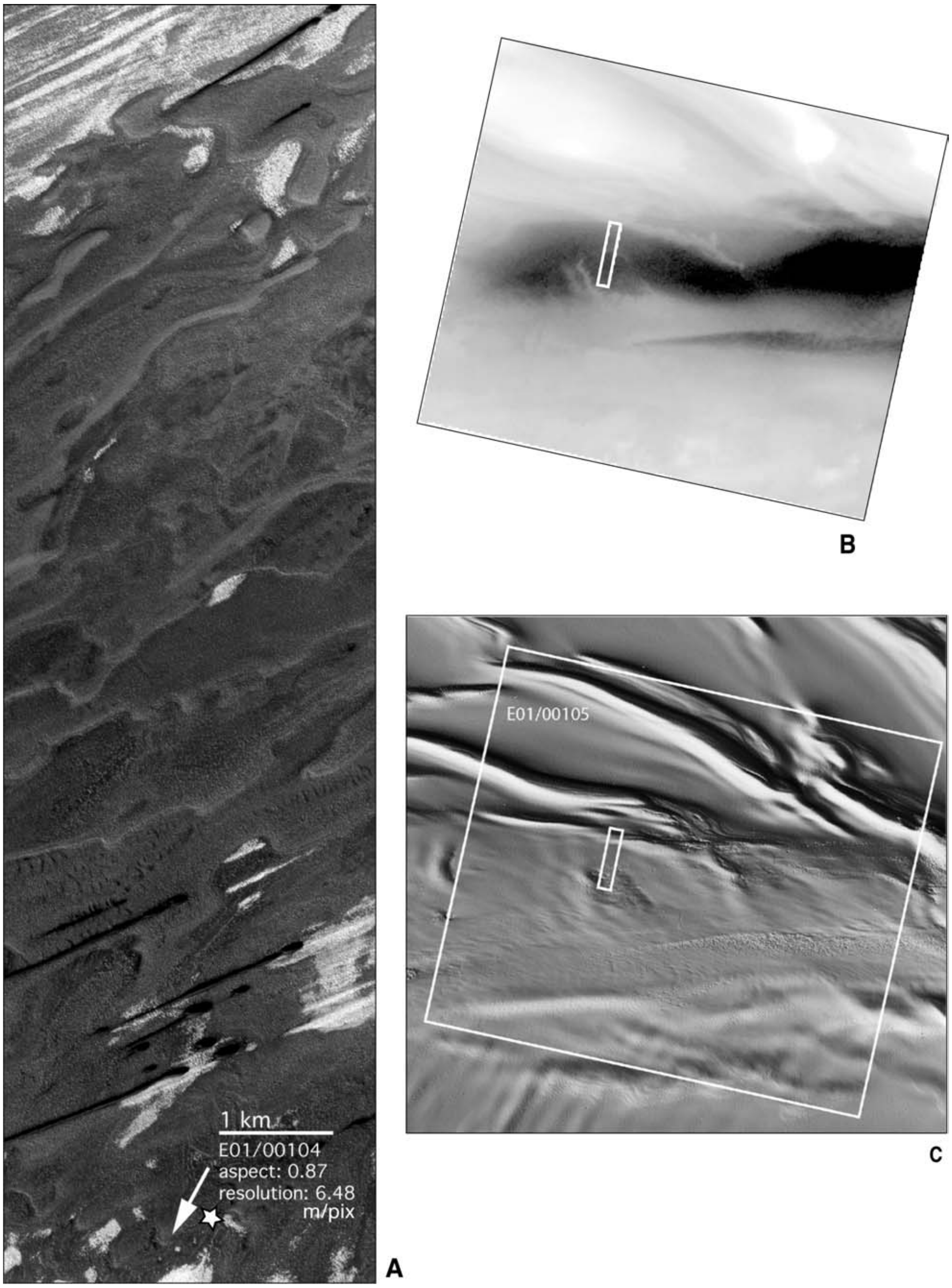
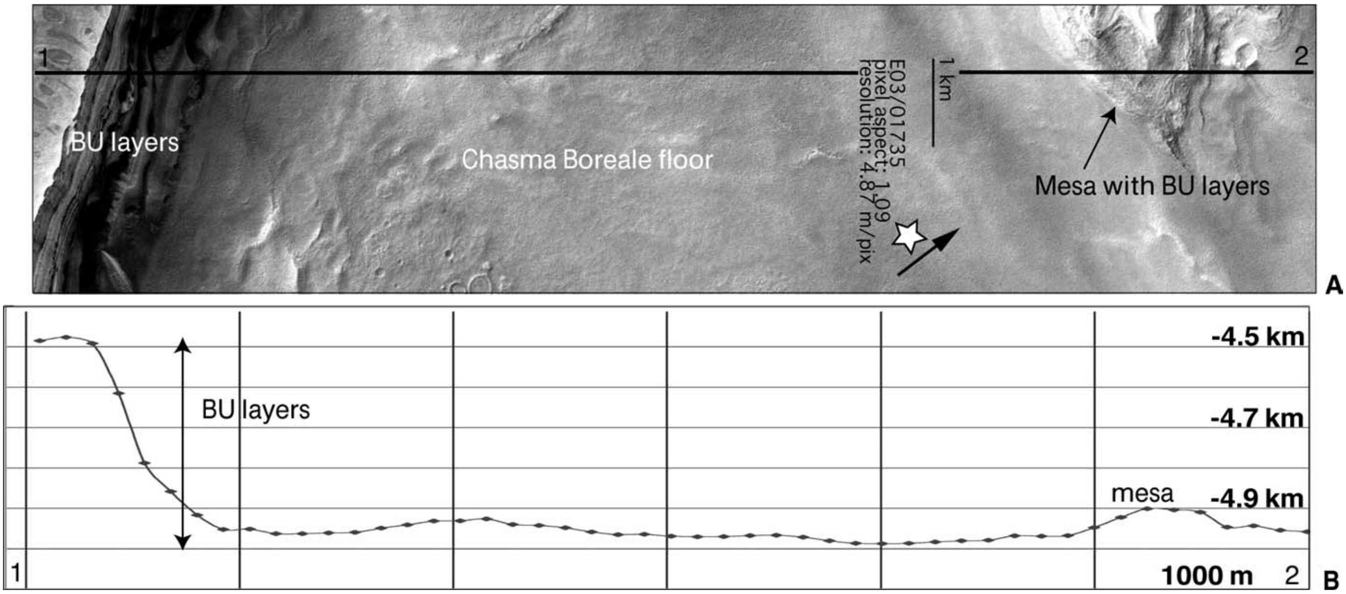



Fig. 10. Example of BU layers exposed on a shallow slope in Chasma Boreale. (A) Portion of MOC image E01/00104. See Fig. 4 for location of this MOC image. (B) MOC wide angle context image (E01/00105) for (A) showing location of *entire* narrow angle image. (C) MOLA Shaded relief map corresponding to context image.

ever, closer examination reveals more complexity. Figure 13 shows several examples of the BU/Apl contact. While the layering in each may appear to be conformable (a), there are actually few if any locations where a single basal unit layer can be traced along the entire contact. Figure 13b shows an example of the top layers of the BU pinching out just beneath the Apl. In many places, there are only remnants of BU layers left beneath the Apl ((e), (f)) or even highly eroded and degraded BU with little evidence of layers (c). While the contact appears wavy in places (e.g., Fig. 2), this does not itself signify an unconformity, but rather a post-formation erosional effect as explained by Edgett et al. (2003).

There exists a transitional zone in the Apl just above the contact in Fig. 2 (II) which may consist of talus (Edgett et al., 2003); indeed light streaks of material emanate from this transitional zone (III in Fig. 2). To allow talus to accumulate, there must be a small break in slope at the contact here, too small to be discernable in MOLA profile data. Alternatively, this transitional zone may be the uppermost layer of the BU with bright streaks emanating from it as observed in the brighter material of the BU in Olympia Planitia (Fig. 6).

Figure 13d illustrates an unusual exposure of the Apl and basal unit. In the lower Apl layers are thin, dark layers. Layers like this are observed in several images of the Apl and



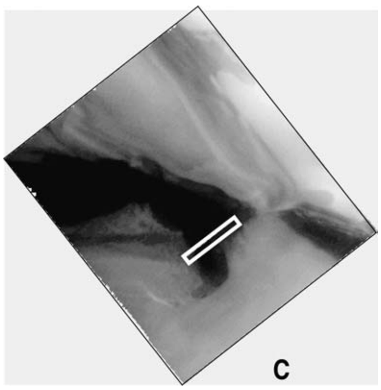




Fig. 11. Example of layered mesa on floor of Chasma Boreale. (A) Portion of MOC image E03/01735. Layers of the basal unit are exposed on the left in the image, and a layered mesa, possibly a remnant of the BU, lies at the rigth in the image. Note its streamlined appearance. Black line shows location of MOLA profile. Image location is shown in Fig. 4. (B) MOLA profile corresponding to image (#19469). (C) MOC wide angle context image (E03/01736) showing location of *entire* narrow angle image.

are discussed in more detail below. At the base of the Apl is a shelf of layered, darker material with bright layers below. A marked albedo contrast is not apparent due to direct Sun on the scarp. These layers may be transitional between the BU and Apl.

### 2.4.2. *Lower (BU/VBF) contact*

This contact is more difficult to recognize than the upper contact. Presumably, if Olympia Planitia consists entirely of the BU, then its contact with the surrounding plains should be an example of the BU/VBF contact (Fig. 3). Unfortunately, this contact is, for the most part, covered by dunes and does not represent a break in slope but instead a transition from dune material to Amazonian mantling material which is underlain by the knobby member of the VBF (Hvk) (Tanaka and Scott, 1987). The region just south of and concentric to the Olympia Planitia boundary may also have been reworked by glacial retreat activity (Fishbaugh and Head, 2000, 2001b, 2002a) obscuring any relationships between the BU and VBF at this contact. However, since Olympia Planitia has a convex topography, and a depression lies in the plains immediately to the south, we can assume that Olympia Planitia (and thus the BU) stratigraphically overlies the VBF with no gradual transition between the two (at least at this location). On the opposite side of the cap (0° W), there are no exposures of the BU so no lower contact is visible. Within Chasma Boreale, the bottom contact of the BU is not necessarily visible as the Chasma floor is not level with the surrounding plains but rather ends in a lobate structure. The contact of the lobate mouth deposits with the surrounding VBF plains is not visible in any MOC images due to frost and poor image quality and coverage at this location, though the lobate deposits are thought to stratigraphically overlie the polygonally grooved member (Hvg) of the VBF with no gradual transition (Dial and Dohm, 1994; Fishbaugh and Head, 2002b).

## 2.5. *Association with dunes*

Thomas and Weitz (1989) have proposed that the lower Apl is the north polar erg. Since at that time the existence of the BU was unknown, it may be that the authors were actually documenting erosion of dunes not from the lower Apl but from the BU lying below it. Byrne and Murray (2002) noted the close geographical association of dunes with exposures of the BU (see their Figs. 4 and 6). Dune fields exist within Chasma Boreale and in the nearby plains and completely cover Olympia Planitia, both being locations where

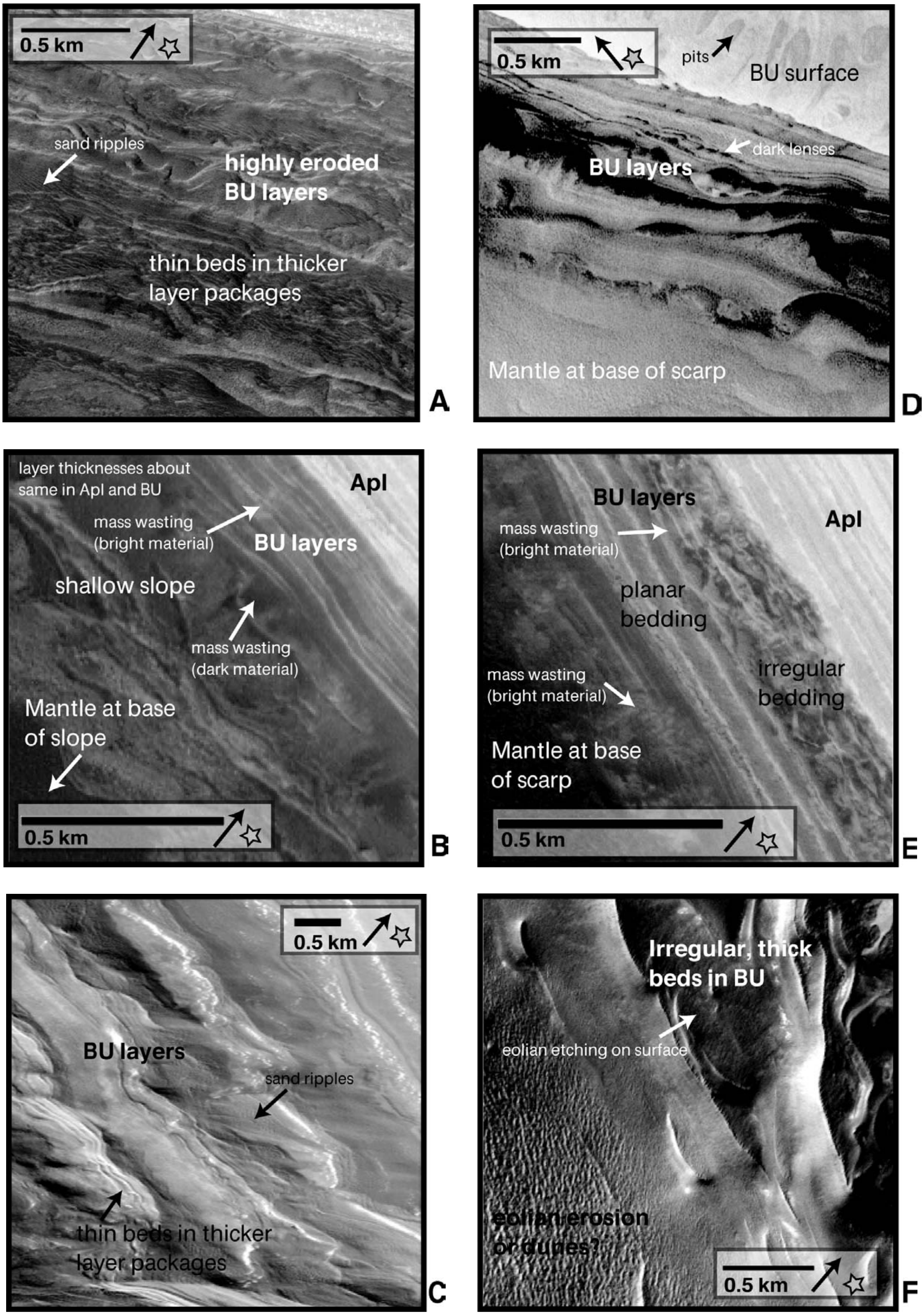



Fig. 12. Examples of layering within the basal unit. Features are described in the text. All examples are portions of narrow angle MOC images, and the image locations are shown in Fig. 4. (A) E02/00202. (B) E02/00023. (C) E02/01788. (D) E03/01735. (E) E03/01438. (F) E04/00539.

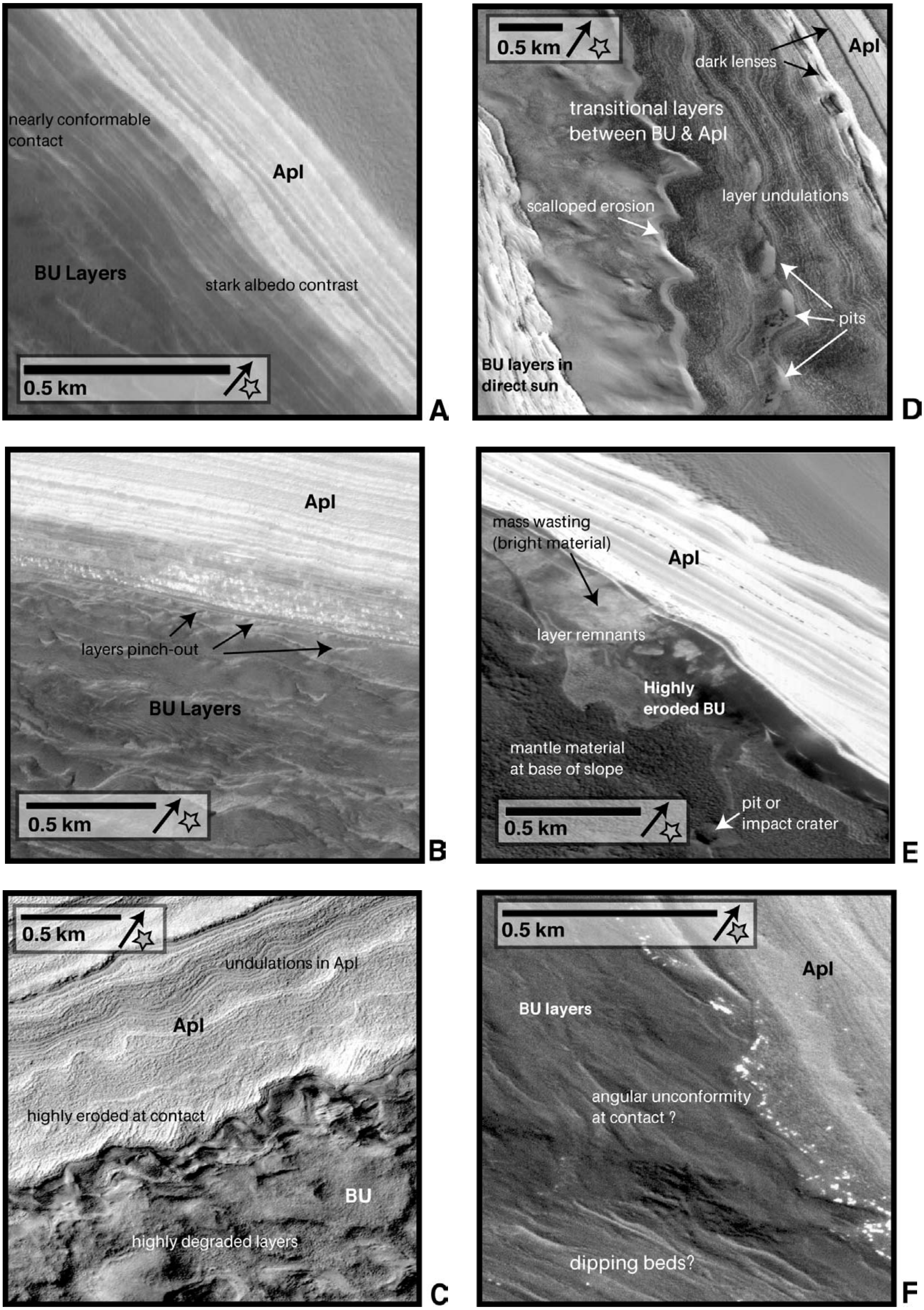



Fig. 13. Examples of the basal unit/Apl contact. Features are described in the text. All examples are portions of narrow angle MOC images, and the image locations are shown in Fig. 4. (A) E02/00023. (B) E02/00202. (C) E02/02634. (D) M20/01772. (E) E02/02811. (F) FHA/01488.

the BU is exposed as scarps and thus has probably been eroded by sublimation and wind, the two dominant erosion mechanisms in the north polar region. This close association has lead Byrne and Murray (2002) and Edgett et al. (2003) to conclude that the BU is an icy, sand-rich deposit and is the source for the large circumpolar ergs (Greeley et al., 1992).

According to our observations, dunes are visible in many but not all MOC images of the unit. We have mapped in Fig. 14 the places where dunes are clearly visible in the MOC images analyzed in detail for this study (map symbol = ⊙). Near the Olympia Planitia reentrant and the head of Chasma Boreale, there appear to be clusters of MOC images containing dunes. Where no dunes are present, there often exists a dark mantle of material which appears to have eroded from the BU (e.g., Figs. 2, 6).

Dune material appears to be eroding directly from the BU in many places (Fig. 15). While it seems that dune material is emanating from the BU/Apl contact in (c), other images show this dark material eroding from many stratigraphic levels within the BU ((a), (b), (d), (e), (f)). It is unsurprising that dune material comes not from one particular layer in the BU, since there are layers of similar appearance throughout the BU with no especially different marker layers.

Dunes may also be eroding from the dark Apl layers just above the BU/Apl contact. Edgett et al. (2003) noted dark lenses in one MOC image. In several other locations (e.g., Fig. 13d), we have observed lenses and discontinuous layers within the lower Apl that are as dark as the dunes in the north polar erg. These are not visible in every image which contains both the BU and dunes, and not all images with dark lenses have dunes (although the dunes may have migrated away); thus, the dark lenses cannot be the sole source of dune material. Dune material must also be eroding from the BU. Byrne and Murray (2002) approximated the basal unit volume, based on their cross-section (Fig. 16), to be $2.7 \times 10^5$ km$^3$, so only 0.4% of the BU would have needed to be eroded from the Olympia Lobe to supply the north polar sand seas which have a combined volume of about $1.2 \times 10^3$ km$^3$ (Greeley et al., 1992).

### 2.6. *Extent of the basal unit*

Byrne and Murray (2002) and Edgett et al. (2003) have found that the BU crops out in troughs near and extending into Olympia Planitia (elevation of BU/Apl contact at about $-4330 \pm 110$ m) and within the walls of Chasma Boreale (elevation at about $-4520 \pm 80$ m) (Figs. 3, 4, 16). Olympia Planitia may also be an exposure of the BU on the 180° W side (Byrne and Murray, 2002) (Figs. 3 and 16). While few quality MOC images of the cap in the longitude range 270°–70° W (Fig. 3) were available at the time of Byrne and Murray's (2002) study, further analysis of more recent MOC images shows that the BU does not appear in the troughs within this region. Thus, the layer pinches-out beneath the cap somewhere between Chasma Boreale and the cap margin or has been eroded back to this location. Figure 4 shows a map of the MOC images analyzed in detail for this study and the locations where the BU was identified. Note that while no images are mapped within the cap center or along the margin east of Chasma Boreale (270°–45° W),

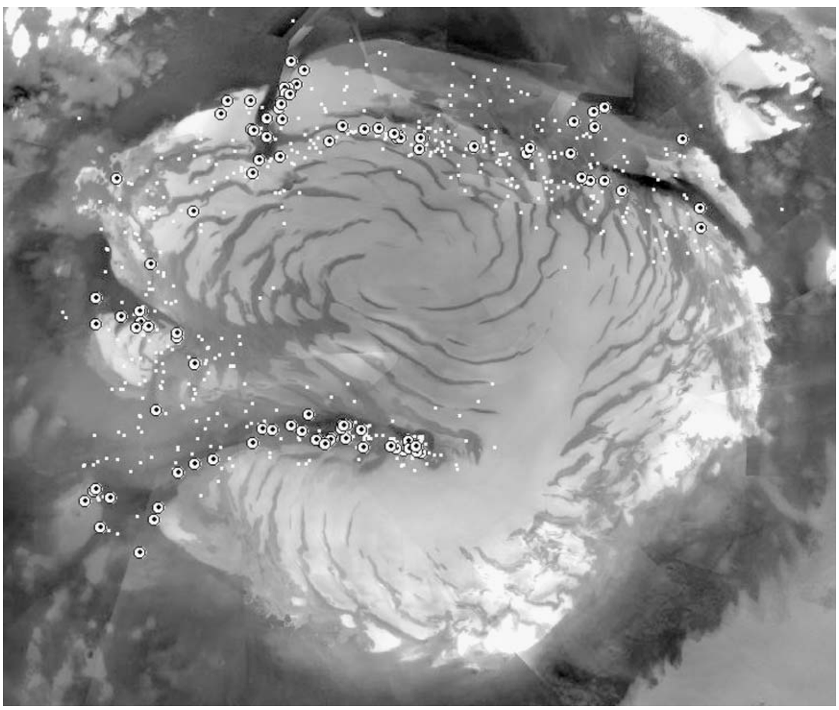


Fig. 14. Map of dune locations noted in MOC image survey. ⊙ = locations of MOC images wherein dunes were clearly visible. Small white squares = locations of most other MOC images analyzed in detail for this study. Note that there appears to be clustering of MOC images containing dunes near the Olympia Planitia reentrant and the head of Chasma Boreale.

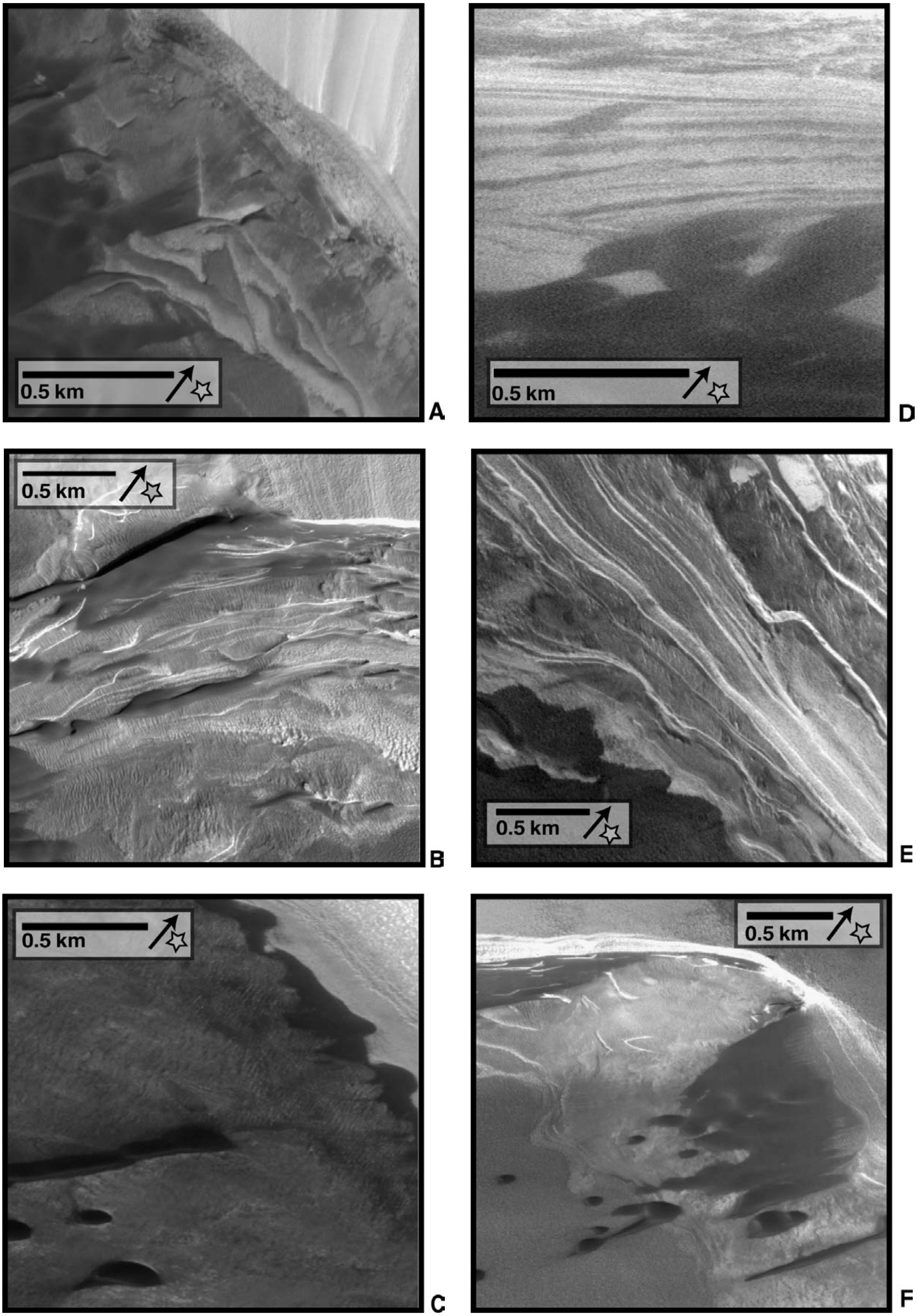



Fig. 15. Examples of dunes eroding from many stratigraphic levels within the BU. All examples are portions of narrow angle MOC images, and the image locations are shown in Fig. 4. (A) E01/01733. (B) E02/02460. (C) E02/01976. (D) E04/01247. (E) E04/01496. (F) FE02/00222.

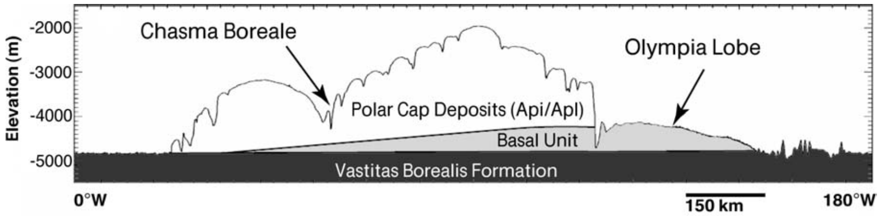



Fig. 16. Approximate profile of the basal unit (shaded grey) beneath the polar cap deposits (from Byrne and Murray, 2002, Fig. 11). The Olympia Lobe consists entirely of the basal unit. We would modify the profile to have the basal unit pinch-out near Chasma Boreale since it appears to outcrop within the northwestern wall, but not within the opposite wall.

we did examine many images within these regions and found no evidence of the BU. From the Fig. 4 it is evident that BU crops out in the scarps and troughs bordering and extending into Olympia Planitia and in the arcuate scarps at the head of Chasma Boreale. Possible examples of the BU are also found at the head of the smaller chasmae to the west. In places where the BU is expected to exist but has not been identified, such as within the walls of Chasma Boreale beyond the chasma head region, the walls of the smaller chasmata and along the cap margin west of the small chasmata, a combination of factors may be disguising its presence: low slopes, images which do not show the cap margin or chasma walls, frost covering and clouds, and poor image resolution. We have drawn in an approximate outline of the distribution of the BU in map view in Fig. 3.

We have continued the investigation into the occurrence of this unit at locations beyond the cap by examining MOC images of the features interpreted by Fishbaugh and Head (2000, 2001a, 2001b, 2002a) as polar material remnants and glacial retreat features (Fig. 3). While there are associated patches of dunes, there is no obvious exposure of the BU within these features. There are few MOC images of sufficient quality to do a thorough search for the BU in this area. Without images of the contact with Apl, it is difficult to recognize BU deposits.

Kolb and Tanaka (2001) have suggested that the sediment lobe extending from the mouth of Chasma Boreale is also a remnant of the BU, and Edgett et al. (2003) claim that the chasma floor consists of the BU. We have examined the few available images of this lobe and find no obvious evidence for it being an eroded BU, though we and Edgett et al. (2003) have noted mesas on the chasma floor which resemble BU deposits (e.g., Fig. 11). Since the lower contact of the BU is not visible in the walls of the chasma, it is possible that the chasma floor and lobe consist of lower BU layers modified by chasma formation, possibly with later deposits on top of them. Unfortunately, there are no images of the lobe scarp (through the E19-R02, February 2003 release) which would allow a cross-sectional view.

Beyond the mouth of Chasma Boreale are a large cratered mesa (Fig. 3) and smaller cratered mesas and cones which have been interpreted by Hodges and Moore (1994) and Sakimoto et al. (2000) as being volcanic in origin. Tanaka et al. (2003) believe that these may instead be isolated remnants of the basal unit (their $Apl_1$ unit, interpreted to be modified polar layered deposits or an ancient sand sea), implying that the unit was once much larger (see their Fig. 13). We find that the large size of the cratered mesa labeled in Fig. 3 precludes it being a remnant of upper Vastitas Borealis layers. Examination of images leads to inconclusive determination of the existence of the BU in these locations due to insufficient image coverage, quality and resolution. However, the fact that scarps have been eroded into the mesa just as they have in the BU and that dune fields lying near these features may have eroded from a BU source provides circumstantial evidence for the existence of the BU there. It is possible that the impact which formed the crater on top of the mesa armored some BU material while the surrounding BU was eroded away, yet there appear to be no major sediment accumulations in or near this location which could account for all of that missing material.

The vertical extent of the BU is more difficult to determine. The contact of the BU with the VBF has not been sufficiently imaged and in most places is hidden by ice and dunes, and the troughs toward the center of the ice cap do not cut deep enough to expose the BU/Apl contact there. Additionally, the presence of a potential polar cap root due to subsidence induced by the polar cap (Johnson et al., 2000) could mean that the BU/VBF contact will be difficult to find. We have seen no evidence thus far for another unit lying between the BU and VBF so that the BU was deposited directly on top of the VBF, any intervening material is hidden by dunes and ice, or the intervening unit has completely eroded away. Future data from the Mars Advanced Radar for Subsurface and Atmospheric Sounding (MARSIS) onboard Mars Express, new MOC images of the scarp at mouth of Chasma Boreale (which may expose the contact between lower BU layers and the VBF), and calculations of the dip angle of the exposed portions of the BU/Apl contact could help to elucidate further the three-dimensional extent of the BU. While it is possible is that the BU exists only at the edges of the polar deposits in the locations observed and not beneath the center, we can think of no likely mechanism to produce this effect (such as squeezing of sediment from beneath the cap or piling up of sediment against the cap) and preserve approximately horizontal layering.

## 3. Summary and interpretation of observations

The BU consists of alternating bright and dark, patchy, overlapping bands. In cross-section the brighter material accumulations appear thinner than darker layers, and in areas where the surfaces of the layers are exposed, brighter material appears to thinly drape the texture of darker layers, allowing buried dark layer features, such as yardang-like forms, to be clearly visible. Thus it can be assumed that the darker layers are thicker (tens of meters vs. meters or less) than the brighter. Streaks of bright material cross-cutting the layers indicate that the brighter material is more mobile than the darker. This is also evidenced by the fact that mass wasting of the darker material is uncommon. At present, we cannot determine whether the lighter material makes up layers or accumulates on flat portions of and terraces in the darker layers. The ubiquitous planetary dust or dirty frost may supply the brighter material, while the darker layers may be composed of the same material that makes up the dark north polar dunes. The steep slopes in parts of the basal unit (in places over 40°) indicate that the material is cohesive. Alaskan loess can maintain slopes in excess of the 35°–45° slopes typical of many BU exposures due to cohesion by "deposition of water soluble minerals at grain contacts" or by "the sublimation of mineral-rich ice at grain contacts" (Johnson and Lorenz, 2000). Yet one would expect mass wasting and shallowing of slopes within such material when exposed to wind. Since little mass wasting is evident in the BU, the cohesive agent must be a stronger than what is observed in the loess (e.g., the presence of magnesium sulfate or ice within pore spaces). Given the environment in which the BU is located (at the pole where ice would most likely collect at the surface), cementation by water ice seems most likely. The observation that many BU layers are exposed in a stair step fashion (e.g., at the bottom of Fig. 6) indicates that the layers of the BU have differing mechanical properties and erode at different rates, much more so than the overlying Apl which show less erosional variation between individual layers.

Since several outcrops lie within polar spiraling troughs, the sublimation and wind erosion forming the troughs (Howard, 2000) is able to erode both the Apl and BU. Edgett et al. (2003) note that the troughs deepen and widen when they encounter the BU. Since the high sediment content of the BU likely hinders sublimation, the unit must be easily erodeable by wind which frees some material and reworks it into dunes. Unless dunes are completely obscuring most trough-forms, few main-cap troughs extend into Olympia Planitia; thus, the wind appears to need an initial trough form (created by sublimation) in order to be concentrated enough for high erosive power. In addition, BU layers extend from beneath the Apl in most outcrops and therefore have not eroded back as far as the Apl. These characteristics indicate that the BU is perhaps eroding more slowly than the overlying Apl. While this suggests to Byrne and Murray (2002) that the Apl is more easily eroded than the BU, Edgett et al. (2003) propose the opposite since they have found places where the BU undercuts the Apl. However, we believe that focusing on which unit is more easily eroded may confuse the issue since each is primarily eroded by different mechanisms. The BU, with its higher sediment content and being a major source for the north polar dunes, is primarily eroded by wind and the Apl, with its higher ice content, by sublimation. Erosion of the BU is characterized by pitting, residual mesas, and dark mantles. The origin of this pitting is uncertain but eolian deflation and/or sublimation of entrained ice may have played a role. In some places ridges (e.g., Fig. 6) have the appearance of yardangs and so were probably created by eolian processes. The high likelihood of the BU being the source of the north polar dunes also indicates susceptibility to eolian erosion.

The Apl unit unconformably overlies the BU as illustrated by the fact that the top BU layers are eroded in many places (Figs. 8, 13). Erosion must have occurred for an unknown length of time between formation of the basal unit and formation of the Apl. Indirect evidence of this erosion may also lie within the dark lenses found in some lower Apl layers (Fig. 13d). Either the dark material of the BU was gradually being deposited in smaller quantities along with the newly forming Apl until it finally ceased to be deposited, or material may have eroded from the BU and migrated as dunes onto the young, still-forming Apl, leaving dark material within the lower layers of the Apl. We and others (Kreslavsky and Head, 2003) have observed dunes stratigraphically on top of thin Apl in the Chasma Boreale region and elsewhere, suggesting that if the Apl is thin enough, dunes may migrate on top of it and be trapped by later Apl deposition. Koerner (1989) has also hypothesized that when the Antarctic ice sheet was still small, wind-blown material was deposited on top of it, and as the ice sheet grew by ice accumulation, the material was trapped, forming the "silty ice" found in some ice cores; while the ice is termed "silty," a range of particle sizes are present, not just silt.

Dunes are geographically associated with the BU, have similar low albedos in MOC images, and appear to be emanating from several BU outcrops (Fig. 15), supporting the earlier conclusion that the BU is a major if not the sole source of the north polar dunes (Byrne and Murray, 2002; Edgett et al., 2003). Near the Olympia Planitia reentrant and within Chasma Boreale, the dunes are near enough to the BU to still lie within the same MOC image as a BU exposure (Fig. 14). Therefore, these may be the locations of the most recent dune formation from the BU scarps with the dunes not yet having had time to migrate a great distance from the BU. Thomas and Weitz (1989) also noted that most dunes are emanating from the Olympia Planitia reentrant. Since katabatic wind erosion is likely enhanced within topographic reentrants (Howard, 2000), the presence of relatively young dunes in those places is expected. It is possible that katabatic winds are being enhanced at the scarp as well.

We have found that the BU does not definitively exist anywhere except beneath the main polar cap, in the form of the Olympia Lobe, and possibly in remnants near Chasma Boreale. Since the BU is exposed in troughs which cut into the Olympia Lobe, we see that at least the upper portion of the Olympia Lobe consists of basal unit layers. This makes probable Byrne and Murray's (2002) hypothesis that the entire lobe consists of BU deposits. If the BU did exist elsewhere, it has since been significantly eroded and may be the source for isolated patches of dunes such as those within the rough topography south of Olympia Planitia (Fishbaugh and Head, 2001a, 2001b).

Kolb and Tanaka (2001) have suggested that the lobate deposits at the mouth of Chasma Boreale may consist of BU deposits. We have found no direct evidence of this, but that may be due to difficulty in recognizing the BU in planform view and a lack of images along the scarp in the lobate deposits. Additionally, the massive amounts of erosion which carved the chasma (Howard, 2000; Fishbaugh and Head, 2002b) might have obscured evidence for the presence of the BU. Indirect evidence comes in the form of the lack of a well-defined lower BU contact in the chasma walls. Since the BU exists on both sides of the head scarps, much of it must have been removed during chasma formation (at most 350 m in thickness), and remnants of the BU may lie on the Chasma Boreale floor (Edgett et al., 2003) (Figs. 11, 16). Our preliminary interpretation of these lobate deposits is that they represent a combination of deposition of eroded BU material and of lower BU layers exhumed by outflow (Fishbaugh and Head, 2002b) and katabatic winds (Howard, 2000). One problem with this scenario is that if the BU had indeed been protruding from beneath the cap in this area, then why would it be eroded from everywhere except the mouth of Chasma Boreale, the locus of the most intense erosion? One might expect instead that the erosional agent which carved the chasma would have eroded the BU in this area, leaving a missing section of the BU rather than leaving it as a prominent lobe.

We have found no definitive evidence thus far for deformation of the BU. The thinning of layers would be difficult to recognize. Possible folds and faults within the BU appear to be uncommon; however, we have found one example of faulting within the Apl (Fig. 17), suggesting that the Apl may be more brittle than the BU. We also mapped all candidate evidence of deformation and found no correlation with slope or topography, only a random scattering throughout the BU.

## 4. Possible origins

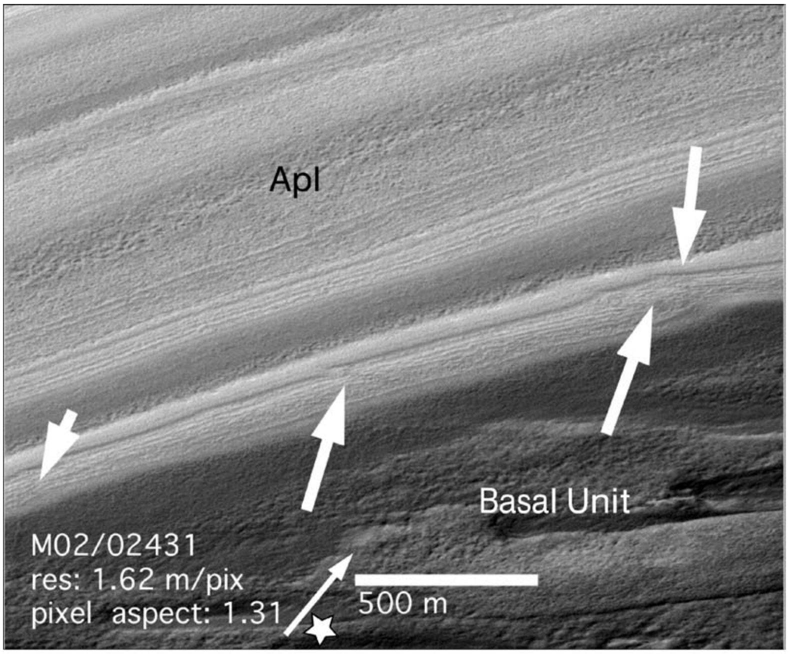
Any formation theory and post-formation modification must explain the following major characteristics of the BU: (1) Dark, patchy layers with varying apparent thicknesses and amounts and types of erosion, possibly with interbedded bright layers or accumulations of bright material on exposed flat portions of the dark layers, (2) likely significant erosion at the contact with the Apl, (3) likely major source of the polar dune seas, (4) differential erosion of layers, pitting, residual mesas, and eolian erosion of some layers, (5) erosion of the BU and deposition onto the early Apl, (6) confirmed existence only within bounds of the current polar deposits and Olympia Planitia, (7) a broad, mounded shape with a thickness of about 600 m, covering an area including the Olympia Lobe and stretching to Chasma Boreale, and (8) a volume of about $3 \times 10^5$ km$^3$ (Byrne and Murray, 2002), about 7% of the volume of the Vastitas Borealis Formation as estimated by Head et al. (2002).

### 4.1. *Outflow channel and oceanic deposits*

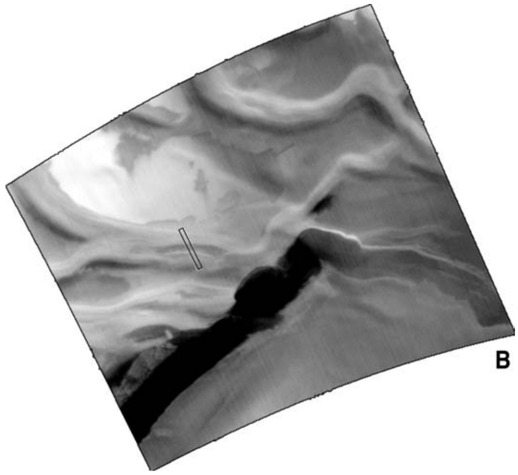
Sediment brought by outflow channels (Baker et al., 1992; Lucchitta et al., 1986) and a possible former standing body of water (Parker et al., 1993) could provide sand and ice-rich material for the basal unit (Fig. 18). Water from the outflow channels could have formed numerous standing bodies of water which froze. These ice and sand-rich deposits could then have sublimated, with the water being redeposited at the pole to form part of the north polar cap (Kargel et al., 1995). This cap would then preserve any sediments beneath it.

Kreslavsky and Head (2002) described sedimentation in an ocean created by outflow channel effluent. If the water was relatively warm, some of the outflow channel sediment load would be dropped soon after reaching the ocean, but turbulent circulation would keep some of the finer grained sediment suspended. As the ocean cooled and convection decreased, this suspended sediment would be deposited in a relatively even layer throughout the north polar basin, except in traps such as craters where it would be thicker. If the water temperature was below the freezing point or maximal density point, convection would decrease, and most of the sediment would be dropped close to the outflow channel mouths. As the ocean completely froze, it would form a sort of high latitude ice sheet which could trap near-pole sediment beneath it.

There are several major problems with forming the BU via this type of sedimentation, having to do with the placement of the BU and the details of its layering. (1) There is apparently little to no evidence for the existence of the BU within the rest of the North Polar Basin, especially the lowest area south of Chasma Boreale. The rough topography south of Olympia Planitia, interpreted by Fishbaugh and Head (2000, 2001a, 2001b, 2002a) to have been formed by retreat of the polar cap, may be a remnant of the BU as there are dunes nearby; but this connection is tenuous at best. Therefore, if the basal unit represents the approximate former thickness of outflow channel/oceanic sediments, then a large amount of material must have been removed from the rest of the north polar basin and deposited elsewhere. Where is that material now? One would also expect much of this material to have been reworked into dunes, yet the major dune seas are near the polar deposits. (2) According to Kreslavsky and Head (2002), as the ocean begins to freeze on top and loses heat to the cryosphere, convection cells modified by Coriolis forces would develop. This may lead to a circumpolar current, but by this time, there is little sediment left in suspension to be concentrated near the




A



B

Fig. 17. (A) Examples of possible faults in Apl (portion of MOC image M02/02431). Large arrows indicate fault locations. Image location is shown in Fig. 4. (B) MOC wide angle context image (M02/02432) showing location of *entire* narrow angle image.

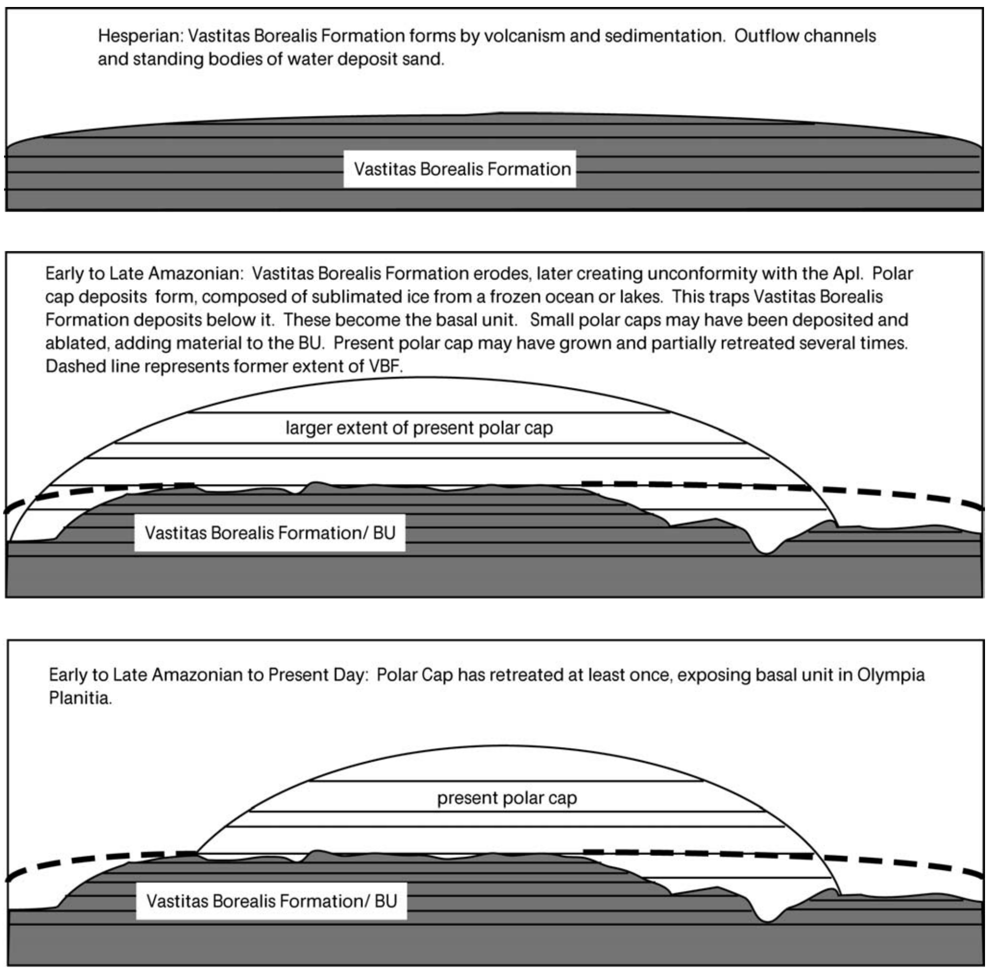



Fig. 18. Basal unit formation by sedimentation in an ocean and by outflow channels.

pole. (3) Following the scenario described by Kreslavsky and Head (2002), the heavier sediments would settle out of suspension first, and the lighter ones would be entrained in the turbulent convection, settling out later. Yet, we see dunes eroding from many stratigraphic levels within the BU and no evidence of a gradual progression in particle size downsection. Even dust storms would be prevented from supplying the brighter layers once ice covered the ocean surface. (4) While repeated outflow channel events could create layers, possibly with time between events for eolian erosion, they would deposit most of their sediment near the channel mouths, not the pole.

### 4.2. *Incorporation into basal ice*

#### 4.2.1. *Terrestrial basal ice*

On Earth, some glaciers and ice sheets have one or more layers of basal ice which can range in thickness from a few millimeters to tens of meters (Fig. 19). Basal ice can have up to greater than 50% sediment by volume and has structural, chemical, and isotopic characteristics distinct from the overlying less sediment-rich ice layers (Benn and Evans, 1998). The contact between the basal ice and overlying ice is often quite sharp. As illustrated by Fig. 19, terrestrial basal ice looks much like the BU.

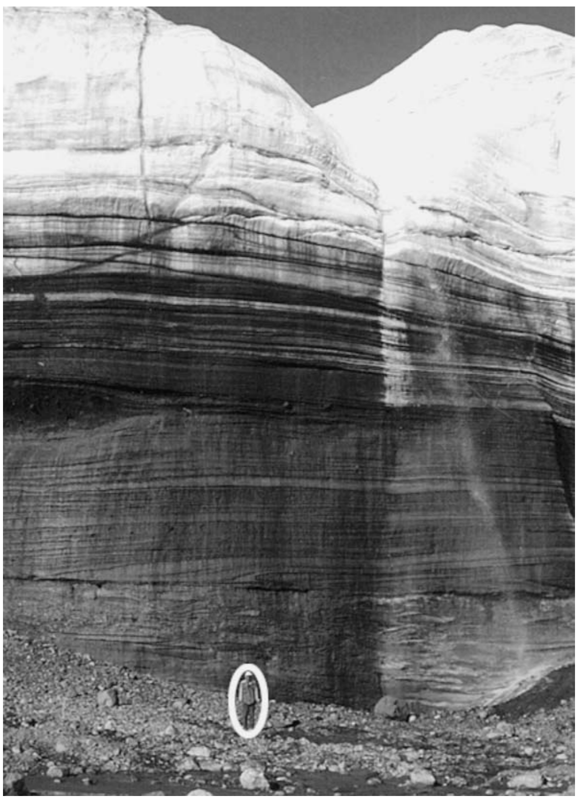


Fig. 19. Example of terrestrial basal ice on Ellesmere Island. Person in foreground is circled for scale. Note the sharp albedo contrast between the dark basal ice and the brighter glacial ice above it. Note also the layering within the basal ice. From Benn and Evans (1998, Fig. 5.19).

Basal ice can form by one or more of the following means (Benn and Evans, 1998; Knight, 1997): (1) regelation, (2) congelation, (3) entrainment of pre-existing ice, or (4) mixing and crevassing. Regelation is the process of localized melting and refreezing of glacial ice. Meltwater can drain vertically into the substrate, later freezing and including the underlying sediments. A steady-state thickness of a basal layer formed in this way is reached when the regelation rate equals the melting rate (Alley et al., 1997). For fine sediments (such as the dust on Mars) with a low permeability, this rate will be slow or even zero. Regelation can also occur around obstacles at the bed (e.g., impact craters on Mars), limiting the thickness of the basal layer to about the size of the obstacle (Knight, 1997). Acting by itself, regelation around obstacles on Earth may only create a basal layer a few millimeters thick and with a low sediment concentration ($\sim 5\%$) (Alley et al., 1997).

During congelation, non-glacial water is frozen onto the glacier and may bring sediment with it (Knight, 1997). This net-adfreezing may occur by conductive cooling or by glaciohydraulic supercooling (Alley et al., 1997). In conductive cooling, wet sediments are frozen on to the bed as the base cools. Climate change or a temporal or spatial fluctuation in the basal thermal conditions can lead to the necessary transformation from melting to freezing conditions. Short climate shifts on the order of centuries to millennia may only add 1–2 mm of basal ice on Earth (Alley et al., 1997). The thicker and more stable the ice sheet, the slower the rate of freeze-on will be as the surface changes in temperature are dampened throughout the ice thickness.

If subglacial meltwater at the pressure melting point flows uphill, following the hydraulic pressure gradient, and thus experiences a drop in overburden pressure, it will freeze and may bring sediment with it in a process termed glaciohydraulic supercooling (Alley et al., 1998). This process has been especially effective beneath Matanuska Glacier in Alaska where, over decades, up to 12 m of debris rich (25–35% by volume) basal ice have been accreted. Roberts et al. (2002) think that this process was active in Iceland beneath Skeidarárjökull during the 1996 jökulhlaup. An abundance of surface water (or a jökulhlaup event) which can also drain to the bed is needed for this process to be highly efficient (Alley et al., 1998). Glaciohydraulic supercooling may have been important beneath the outlet glaciers of the Laurentide Ice Sheet as well (Roberts et al., 2002).

Dirty basal ice can also be created when a glacier advances over buried glacial ice or debris aprons left by a previous retreat. Mixing of sediment and meltwater into the overlying glacier may take place by small-scale folding of the lower layers, and sediment may also be squeezed into basal crevasses (Alley et al., 1997). These mechanisms would be locally important but may not add large sections of basal ice. The role of large-scale folds and thrust faults in the formation of basal ice remains controversial.

Each of these processes requires some amount of meltwater which will later freeze, "gluing on" the basal ice. However, even cold-based glaciers and ice sheets are known to have dirty basal ice, visible in cores at the margins. Cuffey et al. (2000) suggest that this may occur by melting and refreezing near the margin before advance of the glacier, by advancing over permafrost which is sheared in the process, or by any of the processes described above at a time when the climate was warmer. Very small amounts of regelation can occur beneath cold-based glaciers due to the fact that a thin film of meltwater can exist at the interface between the ice and the substrate (Cuffey et al., 1999).

### 4.2.2. *Basal ice on Mars?*

Finnegan et al. (2003) have proposed that the BU may be similar to terrestrial basal ice based on the similar appearance and the stratigraphic location of the BU beneath the ice cap. Since the BU is so thick and widespread beneath the cap, net-adfreezing seems the most likely mechanism for creating such a basal ice sequence, rather than the other, more local processes. The distribution of the unit (being associated exclusively with the cap), the irregular layering, and the sharp appearance of the contact are consistent with this mode of origin.

However, there are several difficulties with creating basal ice in the Mars polar environment. (1) A large amount of meltwater is necessary since the BU is so much thicker than any terrestrial basal ice. Large-scale melting during the formation of Chasma Boreale (Benito et al., 1997; Fishbaugh and Head, 2002b) could provide the necessary amount of water. Yet the formation of the BU pre-dates the formation of Chasma Boreale since it can be seen within the chasma walls. (2) While slow regelation into the underlying sediment over tens of millions to billions of years could possibly account for the great thickness of this unit, there would have to be a means of providing small amounts of water continually or larger amounts of water cyclically for the regelation process. (3) Since the basal ice/Apl contact is a few hundred meters above the surrounding plains, sediment which was later frozen-on to the Apl would need to have already existed as a pile rather than being distributed more evenly. Formation of the BU as basal ice does not explain how the sediment got there in the first place.

Since its initial creation, the upper portion of the BU may have been frozen on to the polar cap as a basal ice sequence, and in this case would need to be taken into account when modeling the rheology of the cap. Interestingly, while basal ice on Earth is commonly deformed due to overburden stress and ice flow, we have seen no definite evidence for deformation in the BU; cold temperatures and high sediment content may prevent this.

## 4.3. *Paleopolar deposit*

While not expounded upon in detail, according to Kolb and Tanaka (2001), the BU represents a "an earlier phase of north polar deposits" (p. 30). We interpret this to mean that the unit consists of, an earlier phase of polar cap deposition (Fig. 20), and find three major observations lend support to this idea: (1) the fine-scale layering observed in the BU, (2) the exclusive association of the BU with current polar deposits (and possibly with remnants of former extents of polar deposits), and (3) the "pile" shape of the unit.

In this case, the BU/Apl contact represents an unconformity in polar cap deposition and an unknown amount of erosion. During the time which this unconformity spans, the environmental conditions changed such that one of two things happened. (1) The amount of atmospheric dust decreased (possibly suddenly) and polar deposition continued with a lower sediment/ice ratio (Fig. 20, Scenario 1). (2) Large-scale erosion of earlier polar cap deposits resulted in a gradually increasing sediment/ice ratio as the ice sublimated leaving sediment behind (Fig. 20, Scenario 2). It is important to note that if the BU is a paleopolar deposit, then the sediment within it would have to be transported atmospherically, making it dust or volcanic ash, or by impact ejecta processes (Wrobel and Schultz, 2004; Schultz and Mustard, 2004). Thus the dunes which have emanated from it must be made up of sand-sized aggregates of dust particles (filamentary sublimation residue) (Herkenhoff and Vasavada, 1999), rather than sand.

If the first scenario prevailed, a mechanism for deposition of greater amounts of sediment with ice needs to be invoked to explain the higher sediment/ice ratio of the BU as compared to the Apl. The atmosphere would need to be thick enough to support suspension of more sediment which could then be deposited in layers with the accumulating snow and frost. Yet the atmosphere could not be so thick as to increase the greenhouse effect to the point where the sublimation rate exceeds the deposition rate. Why so many layers pinch-out within a few kilometers is also puzzling. One might expect more laterally continuous layers in a paleopolar cap (like in the Apl). It is possible that the layers only appear to pinch out due to erosion and that they crop out again in nearby outcrops.

In the second scenario, each BU layer could represent the lag left behind by one phase of polar cap sublimation. Jakosky et al. (1995) calculate that at high obliquities (45–60°), the current cap, ignoring its particulate content, could completely sublimate in about $10^4$ years. If the BU was formed entirely by polar cap sublimation, then each time, the polar cap would have had to sublimate completely. If only partial sublimation occurred, then we would expect to find lenses of basal unit material spread stratigraphically throughout the Apl, and we have no evidence of such lenses thus far. A major question plaguing the hypothesis is: "Would a layered polar cap of a size similar to the current cap be able to completely sublimate?." As the ice sublimated, a lag from the admixed particulates would remain. Since very thin layers of dust can significantly hinder sublimation, the dust may have to be removed to allow more Apl layers to sublimate. Given the current polar wind regime

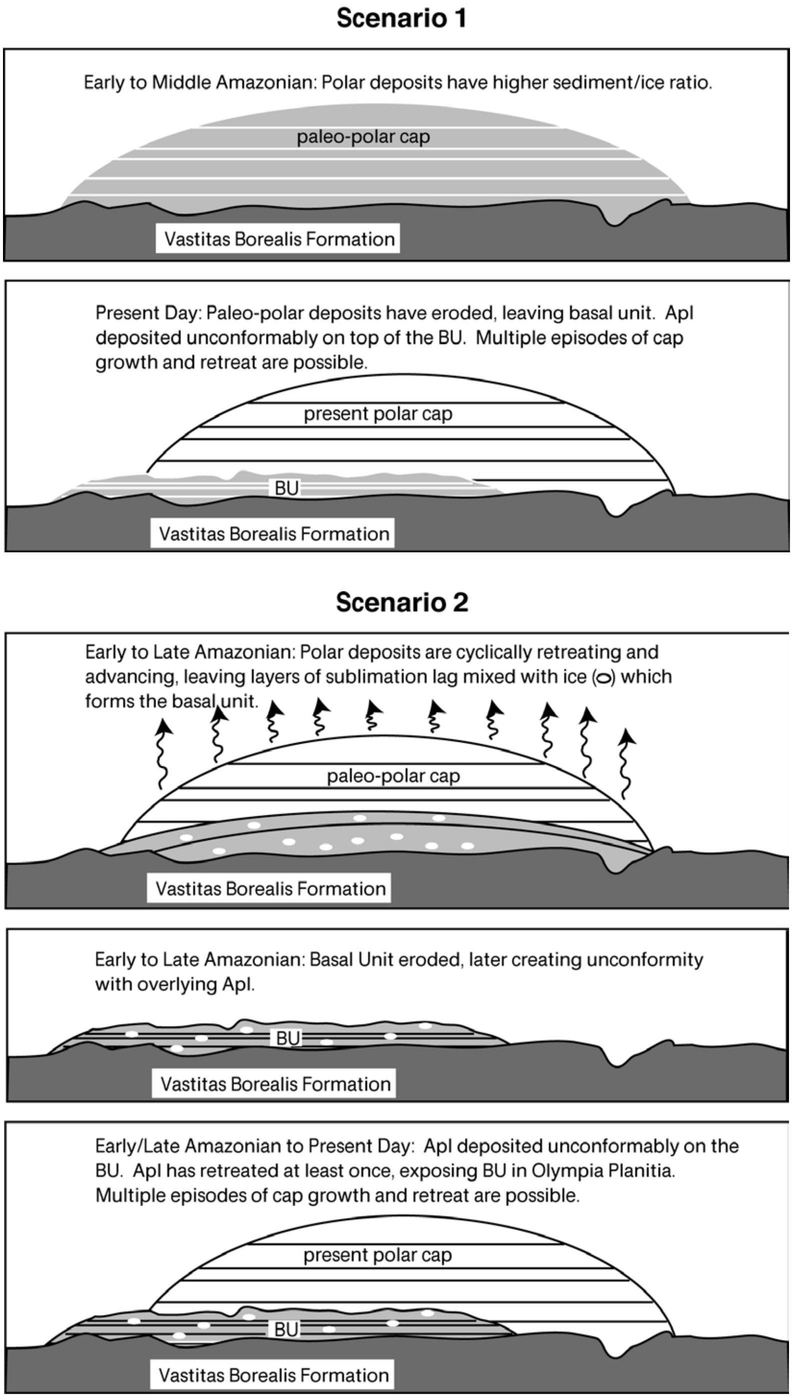



Fig. 20. Basal unit formation as a paleopolar cap deposit. Scenario 1: Basal unit represents polar cap with higher sediment/ice ratio, as explained in text. Scenario 2: Basal unit represents lag layers left by multiple cap growth and retreat cycles, as explained in text.

(Tsoar et al., 1979), the material would not be easily removed; for example, the north polar erg is confined to its present position. However, the wind regime may change with changing climate conditions, and katabatic winds may also contribute significantly to removal of material. One might expect to be left with a pile of sediment on top of preserved Apl. Perhaps Olympia Planitia represents just this case, rather than consisting entirely of the BU. Of course, the complete sublimation of much smaller caps than that which exists today may have added material to the basal unit, yet the apparently sandy nature of the unit precludes it having been formed entirely by that process.

Ancient polar caps could provide a limited amount of sandy material through diagenesis. For example, we have previously discussed evidence of the current polar cap having once been larger (Fishbaugh and Head, 2000, 2001a, 2001b). If the polar cap was ever thick enough to induce melting near the base or close enough to the melting temperature to allow thin films of water to develop around sediment grains (Clifford, 1987), then this water may have allowed solute transport. Together with pressure from the overlying polar cap, this process may have induced diagenesis of dust-sized grains into larger, sand-sized grains within the lower cap layers, thus contributing to some of the sandy material of the BU and north polar dunes. We find it unlikely that such a process could account for all of the sand-sized grains in the BU since the deposit is several hundred meters thick. To produce metamorphism of dust grains into sand grains throughout the entire deposit would probably require a quite thick polar cap lying above it. This process also would not easily produce a layered deposit.

As explained above, any northern polar caps existing during the time of basal unit formation were likely much smaller than the current cap due to the fact that they must have completely ablated. The lag left behind by such caps may have contributed to the BU, and this lag may also have been partially transformed into sand-grain sized particles. If any melting occurred within these polar caps due to such factors as high obliquity-driven temperature increases, the water could not only induce solute transport and removal of interstitial ice to create larger sediment grains, but also may be able to transport this sandy material towards the cap base. Again, it is improbable that the lag left behind by smaller caps could create layers of apparently several 10s of meters thickness and that a sufficient fraction of the lag was converted to sand-sized grains. Additionally, one would expect that BU formation by either of the scenarios described above could have been occurring at the south pole at the same time; however, no southern basal unit has yet been found.

### 4.4. *Eolian deposit*

According to Byrne and Murray (2002), the close geographical association of the BU with dunes indicates that it accumulated as an eolian deposit (Fig. 21). Anderson et al. (1999) have modeled the distribution of sand resulting from saltation, using a sand transport model of White (1979) and taking into account Pollack et al.'s (1990) Mars general circulation model (MGCM). They find that the northern midlatitudes undergo the most erosion and that the resulting sand would migrate mainly to the north polar region, Tempe Fossae, Chryse Planitia, Nilosyrtis Mensae, and towards the equator. The authors estimate that the north polar ergs could form within 50 Ky by these means. The north polar basin could provide a topographic trap to preserve the presence of eolian deposits here. Build up of sand at other locations could possibly contribute to isolated dune patches, but the lack of an enclosed basin may not allow the creation of deposits thick enough to rival that of the BU. The results of Anderson et al. (1999) are dependant upon the threshold stress required for transport, the trapping effects of topography, and the reliability of the MGCM. It is important to note that for the much of BU material to have been transported to the northern plains in a manner similar to that described by Anderson et al. (1999) one must assume a sufficient mid-latitude sand supply and a pattern of Mars general circulation during the time of BU formation which contributes to transport of sand towards the north pole. Even with these caveats in mind, our observations also support an eolian origin as being the simplest and primary means of forming the BU.

(1) Edgett et al. (2003) have found no evidence for the cross-bedding one may expect if the BU was a major dune sea. However, most classic examples of terrestrial cross-bedding in sandstone are exposed in near vertical outcrops (Fig. 22a). The BU, on the other hand, is exposed in shallower-sloped outcrops, making the classic cross-hatching pattern difficult to recognize. The BU layers appear to have been deposited in a patchy, overlapping fashion as expected from dune deposits and from shallower-sloped exposures of cross-bedding (Fig. 22b). In some locations, this "patchiness" is visible on a scale close to that of the size of individual dunes.

(2) The unit is likely the major (if not only) source for the northern circumpolar ergs. A sandy deposit supplying these sand seas is a much simpler explanation than requiring formation of filamentary sublimation residue (Herkenhoff and Vasavada, 1999) from the polar layered deposits to form the dunes, a hypothesis formed when the existence of the basal unit was unknown. While this hypothesis is predicated on the fact that the thermal inertia of the northern dunes is much lower than that of dark dunes at lower latitudes, suggesting that their bulk density is lower, Byrne and Murray (2002) warn that high local slopes and the high emission angle of the Viking observations could complicate interpretation of thermal inertia data.

(3) The unit has been found only in the northern region, the location where migrating sand dunes would be trapped as an erg (Anderson et al., 1999).

(4) Edgett et al. (2003) have also noted the presence of semi-crescentic patches of material being exhumed near the edges of the polar cap which may be paleodunes buried in the BU (see their Fig. 4).

(5) The other hypotheses of origin discussed above require much more complexity in overcoming problems than does the eolian hypothesis.

Assuming a sufficient sand supply, eolian migration could indeed explain an accumulation of sand at the pole, but may not explain the detailed layered structure unless the process occurred with some cyclicity. At the rate estimated by (Anderson et al., 1999), to create 10 layers, each with the volume of the current north polar erg, would take at least about 500 Ka, about 0.02% of the duration of the Amazonian period, allowing ample time for acquisition of sand layers. While the basal unit may be hundreds of meters thick, eolian sandstones of the Colorado Plateau are 3500 m thick in total ($1 \times 10^6$ km$^2$) (Kocurek, 1991), though some of this thickness is due to deposition in a topographic basin. It is also important to note that some of the original BU deposits may not be preserved in the strata. On Earth, often only

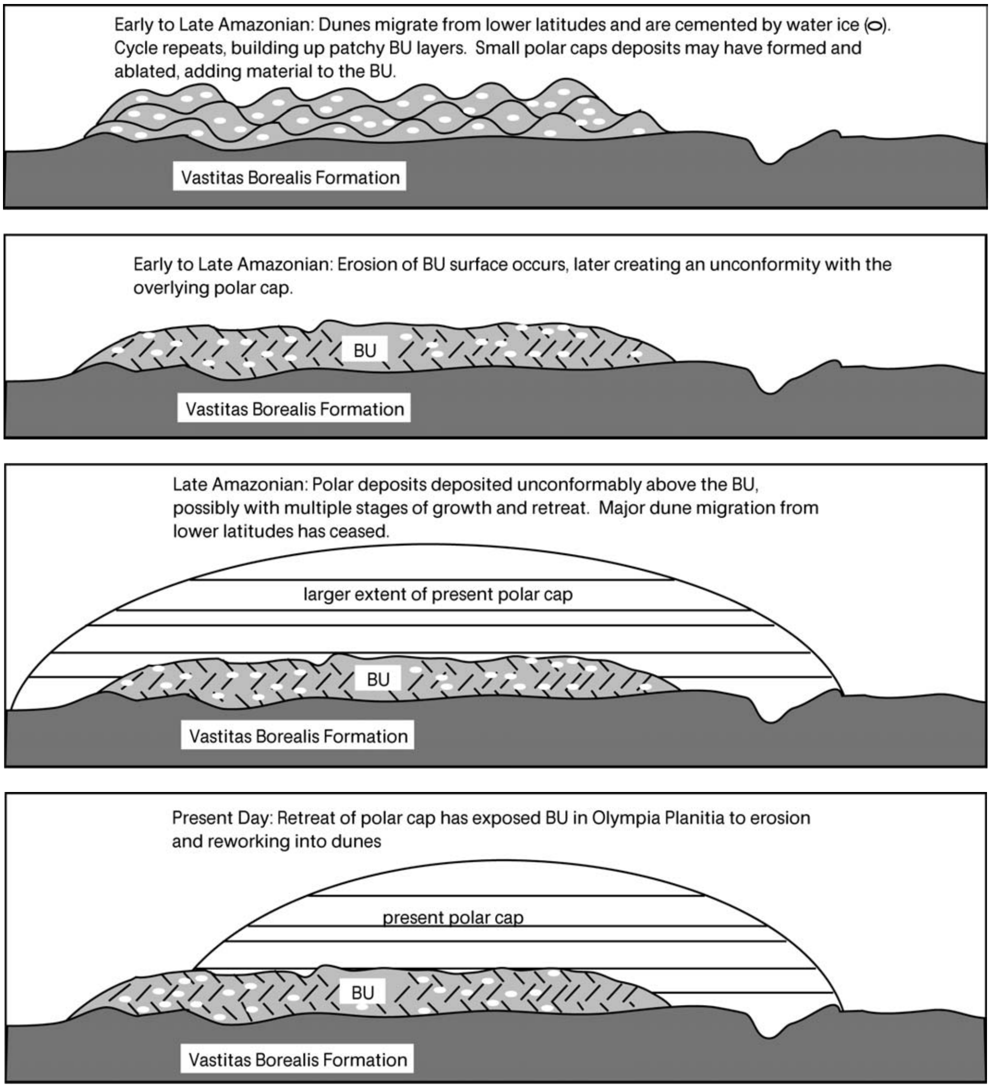



Fig. 21. Basal unit formation by periodic build up of eolian deposits and freezing of these deposits.

the lower portions of the deposit are preserved (Kocurek, 1991).

While one layer may be created quickly, there may be long periods of time between formation of each layer. What would cause the cyclicity remains an open question. On Earth, cyclicity can be caused by variations in sand supply related to such events as glacial cycles (Loope, 1985). Though this may not be applicable in the martian case, changes in orbital parameters and thus general circulation or changes in mid-latitude sand supply may provide cyclicity. In both the north and south 30°–60° latitude bands lies a mantle deposit interpreted to consist of ice-rich dust (Mustard et al., 2001; Head et al., 2003). This mantle undergoes periodic build-up and erosion corresponding to orbital cycles. Erosion may expose sand when the icy mantle disappears, allowing sand to migrate northward. When the mantle re-forms, sand migration could slow (or ceases), and dust may build-up on top of the frozen sand, creating the brighter, more mobile layers (or accumulations) within the basal unit. The seasonal $CO_2$/$H_2O$ cap contains dust and thus could also be a source of polar-concentrated dust which would build up on top of the frozen dunes.

Since the BU can have slopes near 40° and since more than 2.5 km of dirty ice rests on top of the BU, something would have to be cementing the sand. Chemical cementation would require the presence of water (which would freeze), as would physical cementation by ice. Therefore, as sand was accumulating it may have incorporated ice, eventually freezing the deposit in place and allowing dust to collect on the surface until the next episode of dune migration and deposition. Byrne and Murray (2002) speculate that changes in obliquity could ensure that differing amounts of ice would be incorporated into the different BU layers. The physical mechanism of incorporation of ice remains a question. Possibly, deposition of water as occurs on the current polar deposits also occurred in lesser amounts while the eolian deposits were forming. In this sense, the BU could be a "paleopolar deposit" which consisted of eolian deposited sand

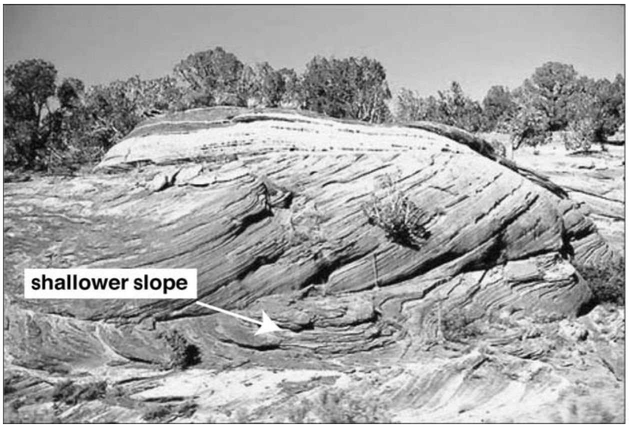




Fig. 22. Examples of terrestrial eolian cross-bedding. (A) Large scale cross-bedding exposed on a steep slope in the Navajo Sandstone. Note the classic cross-hatching pattern. Person is circled for scale. From: http://walrus.wr.usgs.gov/seds/fig68.html. (B) Smaller scale cross-bedding with a shallow-slope exposure near the bottom. Note that patchy layering is evident here, rather than the classic cross-hatched pattern. From: http://www.geosci.unc.edu/faculty/glazner/images/SedRocks/SedRocks.html.

and atmospherically deposited water and dust. Byrne and Murray (2002) find it unlikely that dust would be incorporated into the layers since saltation would preferentially remove it (Herkenhoff and Murray, 1990). Thus, the bright material (whether it make up layers or patches of accumulation) was probably deposited after saltation ceased and the dunes were frozen.

## 5. Age of the basal unit

Underlying the BU is the VBF which is about $3 \times 10^9$ years old (Tanaka and Scott, 1987; Kreslavsky and Head, 2002). The surface age of the polar cap deposits has been estimated to be at most $1 \times 10^5$ years old (Herkenhoff and Plaut, 2000). Together, this evidence gives the BU an age of Early to Mid/Late Amazonian ($1 \times 10^5$–$3 \times 10^9$ years old). It is impossible to absolutely date the BU itself since so little of its surface area is exposed. Also, any craters which may exist on the BU surface could merely date the exposure age of that surface. For example, even if the floor of Chasma Boreale is a basal unit surface, its craters may merely reflect when the Chasma was formed and the lower BU layers exposed rather than the original crater density at the time of BU formation. Also, even if the BU is very old (e.g., Early Amazonian), it may quickly have been covered by a polar cap and as a result may never have accumulated many craters. Thus crater counting does not constrain the age much better (if at all) than does stratigraphic position of the BU. We have counted all craters greater than 5 km in diameter in areas which we consider to be candidate exposed BU surfaces (Fig. 23) and find that there are fewer than 5 craters of $> 5$ km diameter per $1 \times 10^6$ km$^2$. This gives a crater age of Middle to Late Amazonian (Strom et al., 1992). There are craters, better highlighted by MOLA data than by images, near the edge of the polar cap which may be BU craters covered by a thin layer of Apl. However, they may instead be craters formed on top of a thin layer of Apl overlying the BU. We did not include these craters since the BU surface is not exposed here and could not be included in the total area estimate. Tanaka et al. (2003) include more outlying deposits within Vastitas Borealis as part of the basal unit—their Apl$_1$ unit—(deposits which we have not identified as part of the BU) and therefore obtain a higher crater count and thus an older age of Early Amazonian. It is important to note that we do not know how many BU craters are yet buried beneath the Apl. While the existence of more craters on the buried parts of the BU surface may increase the apparent age of the BU, this could be offset by the greater BU surface area involved.

## 6. Conclusions

We now know that the stratigraphy of the Mars north polar deposits reveals much more complexity than previously thought. Beneath the Apl lies a dark, complexly layered unit (Malin and Edgett, 2001; Kolb and Tanaka, 2001; Byrne and Murray, 2002; Edgett et al., 2003) that may preserve a record of the poorly understood north polar Amazonian history.

Our analyses have outlined the following major characteristics of the BU layers: (1) patchy distribution, (2) various manifestations of differential erosion, (3) differing amounts of eolian reworking, (4) relatively low albedo layers of varying thickness which, for the most part, do not exhibit appreciably different morphologies from one outcrop to the next, and (5) little to no deformation. The contact with the BU and the Apl exhibits strong evidence of an erosional unconformity. Part of the eroded BU may have been reworked into dunes which marched on top of the young, still-forming Apl and thus were trapped in the lower layers. So far, we have definitively identified the BU only within outcrops in the scarps and troughs near Olympia Planitia and part of the Chasma Boreale walls. The BU also appears to be the main (if not sole) source for the north polar ergs.

Given the observations that we and others (Byrne and Murray, 2002; Edgett et al., 2003) have made of BU characteristics, we feel that eolian deposition is the simplest way to

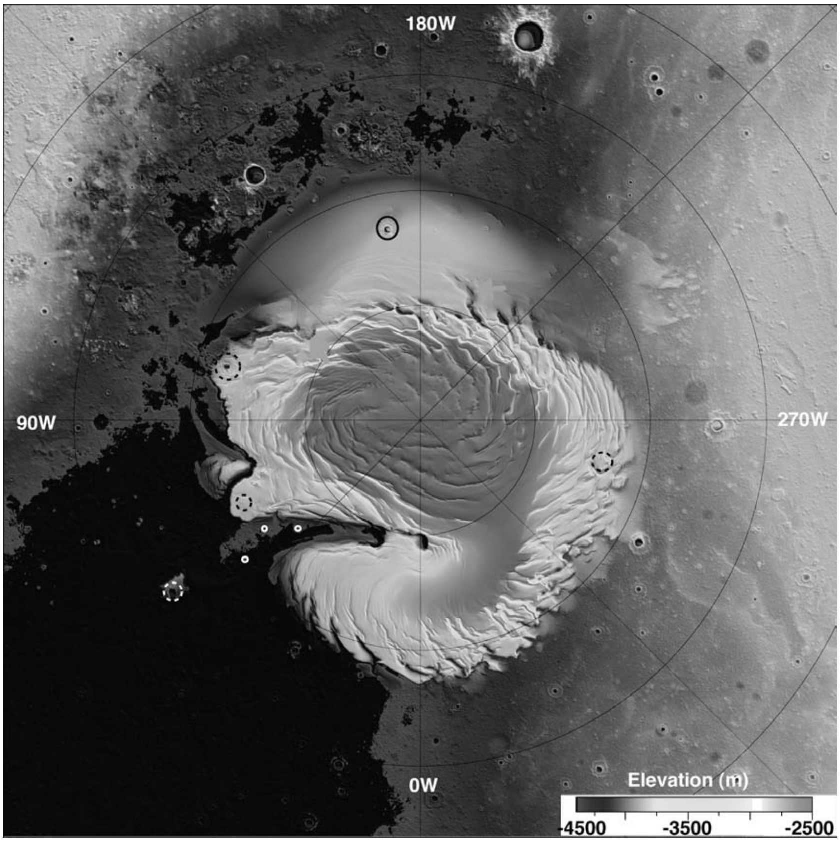



Fig. 23. MOLA shaded relief map showing locations of craters counted for this study (3 in Chasma Boreale, and 1 in Olympia Planitia). The craters circled with dotted lines may lie on the basal unit but are buried by Apl deposits. The mesa at the mouth of Chasma Boreale may be a basal unit remnant, in which case its crater would also lie on the BU.

form the north polar basal unit, supporting the earlier conclusion of Byrne and Murray (2002). It is plausible that during the Early to Mid Amazonian, sand and dunes migrated from lower latitudes (Anderson et al., 1999) and were trapped at the north pole (Fig. 21). Ice was deposited simultaneously, freezing the dunes and forming patchy layering. We do not know the ice/sand ratio in the basal unit, so how different this ice content was from that of the current polar deposits is unknown. This ice incorporation may have been the Early to Mid Amazonian manifestation of polar cap formation unless smaller polar caps were forming simultaneously.

Ubiquitous planetary dust may have collected on top, forming brighter toned layers, until another episode of dune migration took place, and the cycle repeated. The cause of cyclicity is unknown. It may have to do with changes in global wind patterns controlled by changes in insolation or with the periodic replenishment or exposure of the sand supply at lower latitudes due to build-up and erosion of the mid-latitude mantling layer (Mustard et al., 2001; Head et al., 2003). A link with orbital parameters and thus with changing wind patterns could be modeled to test the likelihood of forming the entire thickness of the BU and the total number of layers in this way. Since basal unit formation, agents such as wind and sublimation have eroded its surface. The cause of major erosion of the basal unit and subsequent deposition of the current polar layered deposits is not yet understood, but it may include a decrease in average obliquity. It is possible that a ready sand supply at lower latitudes has been exhausted. In addition, the time span covered by the unconformity between these two units is unknown.

Our observations and conclusions lead us to update the possible scenarios of north polar Amazonian history discussed in the Introduction and in Fishbaugh and Head (2001a). The erosional unconformity between the BU and Apl is consistent with Scenario 1 (Late Amazonian Apl formation). A predominantly eolian deposit in this region implies that there was a time (enough for accumulation of the entire BU) in the Early to Mid Amazonian when the polar cap deposits of the size we see today were not forming, thus deposition of the current polar cap is unlikely to have begun in the Early Amazonian (Scenario 2). If any ancient polar caps were as large as the current cap, then, as discussed above, sublimation of those deposits might be expected to halt when enough lag builds up to prevent it. When new polar deposits form on top of this lag, we may expect to find lenses of lag material within polar layered deposits. Indeed, Milkovich and Head (2004) have used Fourier Analysis of north polar cap layer sequences to find evidence of at least one possible lag layer. However, if any ancient polar caps were much smaller than the current cap, then complete sublimation of those deposits may be possi-

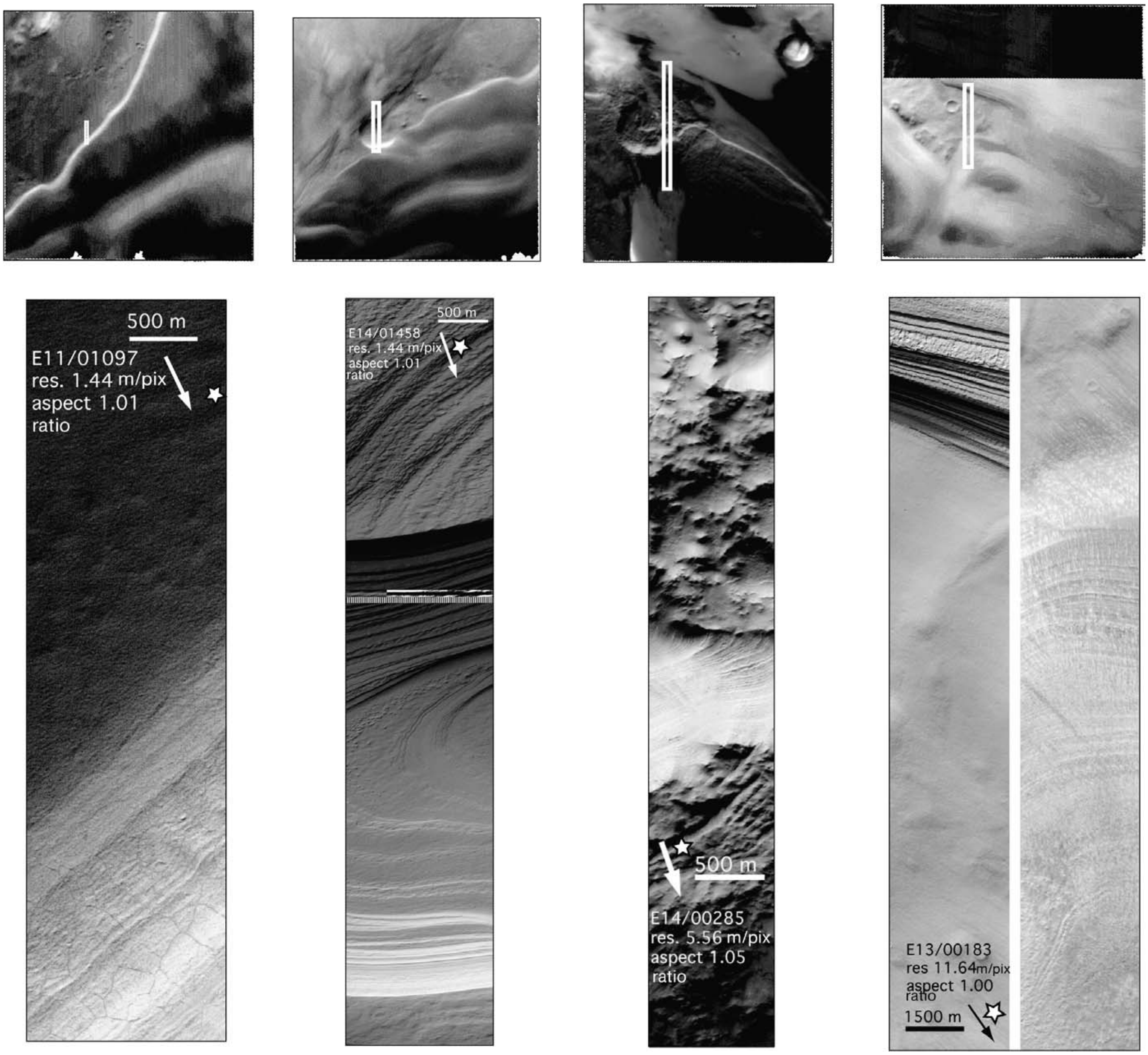



Fig. 24. Examples of portions of MOC images of the southern polar cap deposits and the associated MOC wide angle context images. The right two images show the walls of Chasma Australe and the left two images show other scarps in the Apl. All images lie within areas wherein a basal unit exposure would be expected by analogy with its exposures in the north. However, note that none of these images show deposits obviously similar to the northern BU.

ble, given a sufficiently small dust/ice ratio, favorable winds constantly removing dust, etc. (e.g., Jakosky et al., 1995); thus, Scenario 3 (fluctuating polar caps) remains a viable option. Sublimation of such caps could add material to the basal unit, yet the apparent sandy nature of the unit precludes it having been formed completely of sublimation residue, as the particulate material within the polar layered deposits is thought to consist primarily of atmospherically transported dust rather than sand. Thus, polar caps either were missing during the Early to Middle Amazonian while the basal unit was forming (Fig. 1b, Scenario 1), or were much smaller than the present cap and underwent growth and retreat (Fig. 1c, Scenario 3).

What factors may contribute to such scenarios? Solutions to orbital parameter variations before 20 Myr ago are chaotic (Laskar et al., 2004), so how polar cap growth and retreat are directly tied to obliquity variations during most of the Amazonian cannot be determined. However, the recent suggestion that Mars may have spent most of its time before 20 Myr ago at an obliquity of near 45° (Laskar et al., 2004) may explain this conspicuous polar cap absence or smaller size since Jakosky et al. (1995) find that pure, water ice polar cap deposits 3 km in thickness could sublimate in 10 Ky at high obliquities. Those authors assume no formation of dust lag or stabilizing effect of a permanent $CO_2$ ice cap covering, thus the actual sublimation time may be much longer. However, certainly the higher polar temperatures associated with such high obliquity would hinder formation of a large cap since the ablation/accumulation ratio may be much larger under those conditions than it is now. Additionally, under such obliquity conditions, most water would be stored within the regolith (e.g., Jakosky et al., 1993, 1995; Mellon et al., 2004), ice-rich mantling deposits (Mustard et al., 2001; Head et al., 2003), and the atmosphere, the three other martian water reservoirs in the absence of an ocean, thus leaving less ice available for deposition at the poles. If the obliquity was not much greater during the Early to Middle Amazonian, then the reason for the apparent lack of a large polar cap in the north remains an open question. Additionally, it is important to note that if the obliquity was much higher in the past, we would expect to find periods of absent or smaller southern caps. To test this idea, one might investigate the nature of the contact between the southern polar deposits and the underlying terrain; is it unconformable? Does there appear to be Early to Middle Amazonian deposition or erosion at the south pole which does not correspond to the presence of a polar cap of equivalent size to the one that exists today?

Since, in the absence of a large polar basin, migration of sand toward the southern pole results in trapping within craters along the way, the southern pole should not act as a trap for migrating sand (Anderson et al., 1999). Preliminary investigation of MOC images in the south polar region (particularly within the Chasma Australe walls and other locations where by analogy with the north we would expect to find it) has yielded no definite evidence for a similar southern basal unit (Fig. 24). It may be that the southern BU is quite small, either having formed as a smaller deposit or having been eroded back. A southern BU may also be manifested differently than in the north due to different environmental conditions. Murray et al. (2001) have found a massive, pitted unit underlying the southern Apl that may represent some type of old polar deposit or sublimation residue from such. The Dorsa Argentea Formation, part of a more extensive Hesperian southern cap (Head and Pratt, 2001) could be explored as a Hesperian manifestation of a southern BU. It should be noted that the detailed layering and relatively low albedo found in the northern BU is not evident in the pitted layer described by Murray et al. (2001) or within the Dorsa Argentea Formation. It is most probable that such a deposit never formed at the south pole. This would be consistent with formation of the northern BU by eolian deposition since Anderson et al.'s (1999) results show trapping of large eolian deposits only in the north.

We have built upon previous investigations of the basal unit (Malin and Edgett, 2001; Kolb and Tanaka, 2001; Byrne and Murray, 2002; Edgett et al., 2003) and have described its characteristics in detail and its possible eolian origin. The importance of this unit lies in the fact that it spans some of the nearly 3 byr of "missing" north polar history during the Amazonian and likely represents a phase of polar deposition different from what we see today. The upcoming MARSIS radar data from the Mars Express mission may provide more information about the vertical and lateral structure of the Apl/BU contact near the center of the polar cap. When THEMIS data are available, we may be able to use the infrared images to gain more information about ice versus sediment content in the BU.

## Acknowledgments

Thanks are extended to Shane Byrne (formerly at Caltech, now at MIT), Trent Hare (USGS), and Lynn Carlson (Brown U.) for enormous help in the GIS mapping portion of this project. We also appreciate valuable discussions with Ken Edgett (MSSS), Asmin Pathare (Caltech), Peter G. Knight (Keele University), and Jack Mustard (Brown U.). Steve Clifford gave us needed insight into the implications sediment diagenesis associated with ancient polar caps. Shane Byrne and Steve Clifford provided extremely helpful reviews of this paper which greatly increased its value and quality.

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