



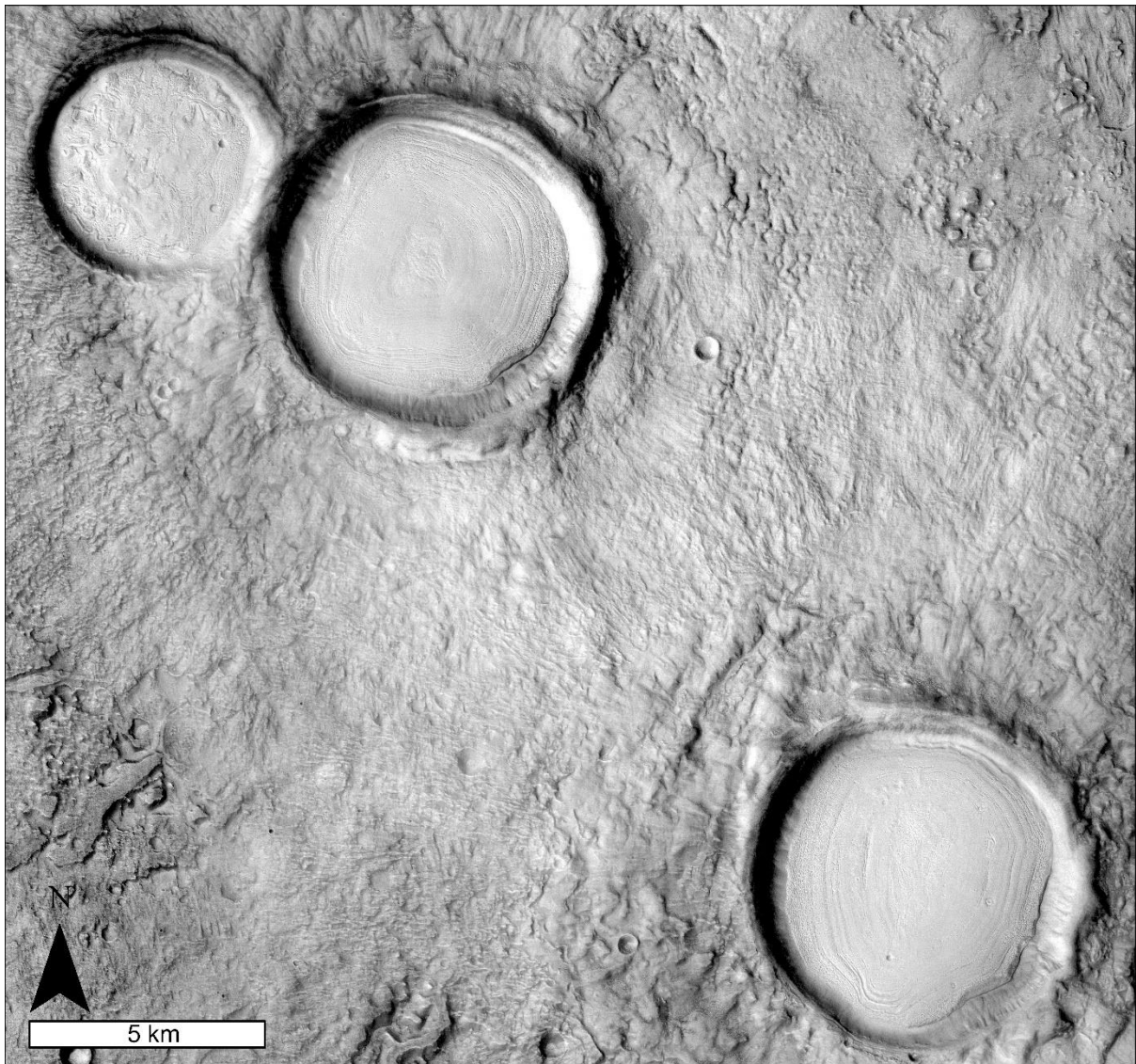
Concentric Crater Fills as Archives of Martian Mid-Latitude Glaciation

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A Thesis submitted in partial fulfilment of the requirements for the degree of
Doctor of Philosophy

The University of Sheffield
Faculty of Social Sciences
School of Geography and Planning

Submission Date
January 2026



Frontispiece. A trio of crater-confined glaciers in Acidalia Planitia. CTX image of three neighbouring concentric crater fills (CCFs) at 43.08° N, 51.01° W, providing a localised comparison of mid-latitude glacial preservation on Mars. Despite sharing a near-identical regional setting, the craters show marked contrasts in fill level and surface organisation, ranging from highly uniform concentric trachelines to more irregular and degraded expressions. The scene encapsulates a central theme of this thesis: that crater-scale geometry and local conditions exert a strong influence on the evolution and preservation of crater-confined ice (CTX image: P15_007099_2242).

Abstract

Mars is a cold, hyperarid planet, yet it retains a water ice inventory sequestered within the polar layered deposits, widespread ground-ice reservoirs, and mid-latitude debris-covered glaciers termed Viscous Flow Features (VFFs). Because surface ice is unstable at Mars' mid-latitudes today, VFFs are interpreted as relict landforms formed under a past climate regime. Concentric Crater Fills (CCFs) represent the most spatially extensive VFF subclass, confined within impact craters. CCF morphologies span well-preserved to degraded states, though it remains unclear whether their modern form reflects initial accumulation or subsequent post-glacial modification.

This Thesis isolates a global population of ~3,000 high-confidence CCFs and uses orbital images and elevation data to investigate the environmental gradients and mass-balance histories recorded in their current state. By treating impact craters as predictable geometric containers, my study of these CCFs separates crater-imposed topographic controls from external climatic signals. Analysis of the population reveals a decoupling that their ice volume generally increases with host-crater diameter, but their surface geometry (elevation deficit below crater rim or "apparent depth", surface slope, and slope aspect) exhibits coherent, latitude-dependent variations that are largely independent of host-crater capacity.

I argue that CCF surface geometry primarily records the effect of post-glacial ablation rather than initial accumulation. The poleward decrease in apparent depth and slope is consistent with insolation-mediated downwasting, where reduced solar forcing at higher latitudes favours the preservation of ice which fills craters closer to the rim level. A fundamental hemispheric asymmetry is identified: northern CCFs possess lower apparent depths, possess no surface aspectual bias and are more numerous, whereas southern CCFs exhibit lower volumes and stronger pole-facing aspect biases. This suggests an asymmetry in CCF evolution, where the northern mid-latitudes experienced more effective ice preservation. Collectively, these findings establish CCFs as systematic environmental archives.

Title

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Author: Ben Cornford, School of Geography and Planning, January 2026

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Declaration

I, the author, confirm that the Thesis is my own work. I am aware of the University's Guidance on the Use of Unfair Means (www.sheffield.ac.uk/ssid/unfair-means). This work has not been previously been presented for an award at this, or any other, university.

Acknowledgements

I wish to express my deepest gratitude to my supervisors, Felix, Frances, and Robbo. Thank you for your guidance and encouragement and for your faith in my ability—even when I had none—that made this work possible. I am especially grateful to Felix for his diligent feedback (including the dreaded red pen) and near-endless patience, and to "Mars Queen" Frances for being an encyclopaedic fountain of knowledge for all things Mars (and Lego). I hope this Thesis reflects the growth I have achieved under your mentorship.

To my colleagues and friends within the department thank you for making this experience—and coffee mornings—enjoyable: Emily, Jane, Becks, Winnie, Aishwarya, and Charlotte. Thank you to Dan for providing the mentorship and sanity checks I needed to stay the course.

To my wonderful friends, thank you for the many trips to the Peaks, introducing me to wild swimming, and the many adventures that made the last few years so much more than just a PhD. To Keeta and Chang, thank you for being on this journey with me from the very beginning and for never saying no to a hike in the Peaks (no matter how precarious). To Drew, thank you for being one of the most kind and considerate people I have ever met, and for keeping me sane throughout the last few years. To Bezo Otto, thank you for being an amazing person, a fellow troublemaker, and the Merry to my Pippin. To Hec, my fellow cave enthusiast, thank you teaching me to ride a bike (almost) and for being a sweet angel—and for saving me from that moth. To Libby, thank you for being your wonderfully chaotic self, never change. To Sims, Ruth, Jack, Jude, Emma, Lizzi, Alice, Alex, Steph, Harri, Ev, Greg, and Kai: thank you for your friendship, kindness and support and for making Sheffield such a wonderful place to call home. To Olive, meow. To Liam, thank you for proof reading. I am also deeply grateful to Peak Queer Adventures for the sense of community I found there.

To Mum, thank you for the lifelong support and encouragement that led me to this point. I am incredibly grateful for everything you've done to help me reach this point.

Finally, on a deeply personal note, I wish to thank the NHS Heartlands Hospital A&E staff, the University of Birmingham Neurology Department, and the Birmingham Complex Epilepsy Team. While the staff who saved my life after 42 one too many seizures and helped me overcome retrograde amnesia in 2018 may never read this, I owe them a debt I can never repay. I am quite literally here because of them.

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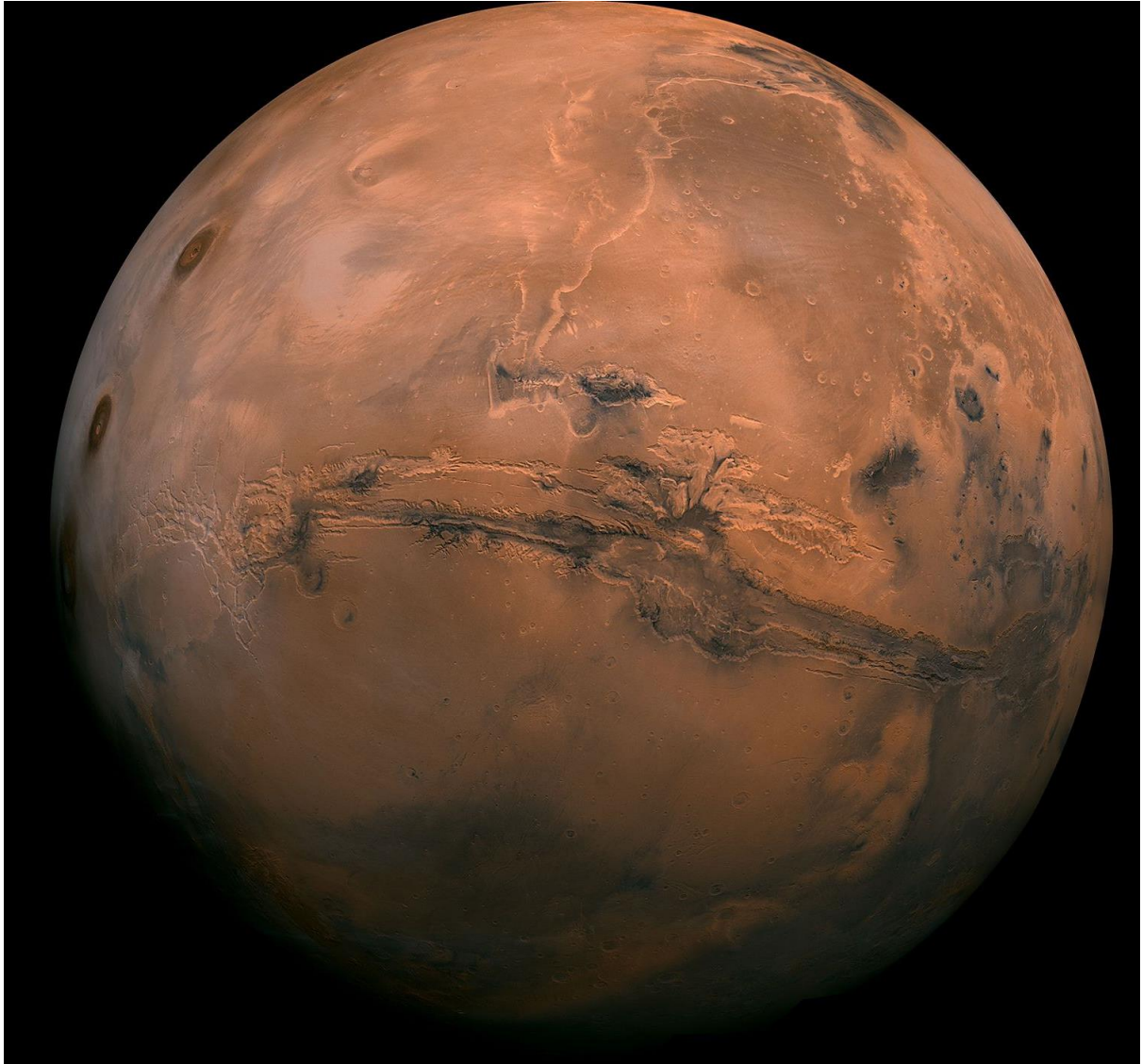
List of commonly used terms, their corresponding abbreviations and reference

Term	Abbreviations	Reference
Global equivalent layer	GEL	Jakosky, 2021
Concentric Crater Fill	CCF	Squyres (1979)
Viscous Flow Feature	VFF	Milliken et al. (2003)
Lobate Debris Apron	LDA	Squyres (1978)
Lineated Valley Fill	LVF	Squyres (1979)
Glacier Like Form	GLF	Souness et al., (2012)
Latitude Dependant Mantle	LDM	Kreslavsky and Head (2002)
Global climate model	GCM	-
Mars Orbiter Laser Altimeter	MOLA	Smith et al., 2001
High Resolution Stereo Camera	HiRISE	McEwen et al. (2007)
Context Camera	CTX	Malin et al. (2007)
Shallow Subsurface Radar	SHARAD	Zurek and Smrekar (2007)
Thermal Emission Imaging System	THEMIS	Christensen et al., 2004
Digital elevation model	DEM	-
Gigayear	Gyr	-
Megayear	Myr	-
kiloyear	kyr	-
Apparent depth	d_a	Chapter 5
CCF surface elevation	Z_{CCF}	Chapter 5
Host crater rim	Z_{rim}	Chapter 5
Host crater diameter	D	Chapter 5
Lambert Conformal Conic	LCC	Snyder (1987)
CCF volume	V_{CCF}	Chapter 6
Host crater volume	V_{host}	Chapter 6

"Yes, that's what happens when water gets cold."

Megamind (2010)

Chapter 1: Introduction



Mars mosaic comprising 102 Viking Orbiter images centred on Valles Marineris

Image credit: JPL-Caltech/NASA

1.1. Introduction

Present day Mars (Figure 1.1) is a hyperarid desert planet, but water ice remains a dominant component of its near-surface environment at mid-to-high latitudes and a key archive of its recent climate history. The modern Martian surface is characterised by a mean annual temperature of $\approx -60^\circ\text{C}$ (e.g., Piqueux et al., 2024), low surface gravity (3.71 m s^{-2} , e.g. de Pater and Lissauer, 2015) and a mean surface pressure of only $\approx 6\text{ mbar}$ (e.g., Jakosky, 2019) within a thin, CO_2 -dominated atmosphere containing only trace amounts of water vapour (H_2O partial pressure $\approx 1\text{ }\mu\text{bar}$; De Vincenzi et al., 1999; Read and

Lewis, 2004) such that perennial surface ice is only stable in the polar regions (Leighton & Murray, 1966) and shallow ground ice is thermally stable only at high latitudes ($\gtrsim 60^\circ$; Mellon and Jakosky, 1995). Mars experiences large, chaotic variations in its orbital forcing (e.g., Ward, 1974; Laskar and Robutel, 1993; Laskar et al., 2004), which modulate the spatial and seasonal distribution of insolation (e.g., Ward, 1974), and are therefore theorised to exert first-order controls on the stability, migration, and storage of near-surface ice (e.g., Mellon and Jakosky, 1995; Montmessin et al., 2007; Mischna et al., 2013). At present, Mars has an obliquity of $\sim 25.1^\circ$ and an orbital eccentricity of ~ 0.093 , and reaches perihelion at a solar longitude of $L_s \approx 250^\circ$ (e.g., Ward, 1973; Murry et al., 1973). These parameters are transient, with dynamical solutions indicating that obliquity oscillates by $\sim \pm 10^\circ$ on ~ 120 kyr timescales, eccentricity oscillates between ~ 0 and ~ 0.12 on ~ 2.5 Myr timescales, and precession operates on ~ 51 kyr cycles (e.g., Murry et al., 1973; Laskar et al., 2002, 2004; Levrard et al., 2007). These orbital variations can be reconstructed robustly only over the last ~ 20 Myr; beyond this, solutions become non-unique as small uncertainties in precession amplify through chaotic spin-orbit dynamics (Laskar and Robutel, 1993; Laskar et al., 2004). Such large shifts in obliquity are interpreted to have permitted substantial ice accumulation at latitudes where surface ice is not stable today (e.g., Mellon and Jakosky, 1995; Levrard et al., 2004; Madeleine et al., 2009). Thus, while non-polar ice may reflect inheritance from different orbital states, its modern distribution is contingent upon long-term preservation via protective overburden or sublimation lag.

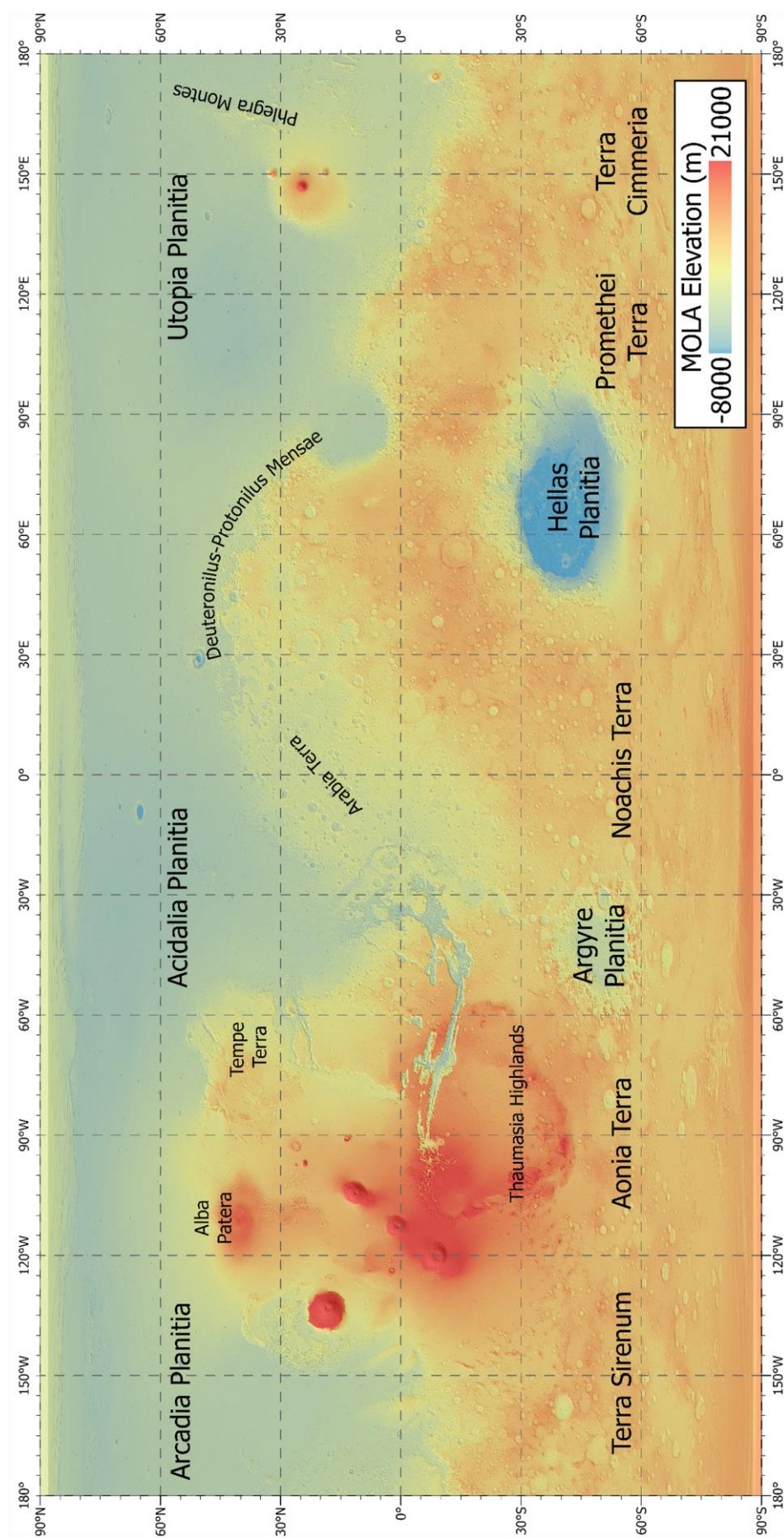


Figure 1.1. Global elevation map of Mars and major physiographic regions. Mars Orbiter Laser Altimeter (MOLA; Smith et al., 2001) gridded elevation data displayed in an equirectangular projection and overlain on a MOLA-derived hillshade basemap. Major physiographic regions within the mid-latitude belts ($\sim\pm 30^\circ$ – 60° N/S) discussed in this thesis are annotated for geographic reference. The map also illustrates the hemispheric topographic dichotomy between the high-standing southern highlands and the low-lying northern lowlands, which dominates Mars large-scale physiography.

Since the mid-1960s, more than 30 successful missions (1965–January 2026), launched by multiple space agencies (e.g. NASA, ESA, ISRO, ISAS, Roscosmos, MBRSC, CNSA), have progressively revealed a vast, latitudinally organised distribution of ice deposits. While early flybys established the harshness of the Martian environment, it was the orbital imaging from Mariner 9 in 1971 that first identified landforms interpreted as ice-related, providing the initial evidence for non-polar ice reservoirs (Masursky, 1973; Squyres, 1979). In the subsequent decades, high-resolution imaging, spectroscopy, topographic data and radar sounding have demonstrated that vast subsurface ice reservoirs persist equatorward of these limits, buried or shielded from sublimation.

Current estimates place the total volume of exchangeable water—the reservoir available for active cycling between the surface and the atmosphere—on Mars at ~30–50 m global equivalent layer (GEL) liquid water (Carr and Head, 2015, 2019; Alsaeed and Jakosky, 2019; Jakosky and Hallis, 2024; Jakosky et al., 2026), with only a negligible fraction (~10 μm GEL) present in the atmosphere as water vapour (Titov, 2002; Jakosky, 2021). This inventory is partitioned latitudinally among the polar ice caps (e.g. Smith et al., 1999; Smith et al., 2001), shallow ground ice at mid- to high- latitudes (e.g., Byrne et al., 2009; Piqueux et al., 2019; Dundas et al., 2021), and thick buried ice bodies preserved within debris-covered glacial landforms in the mid-latitudes (e.g., Squyres, 1979; Milliken et al., 2003; Souness et al., 2012; Levy et al., 2014). Hydrated minerals retain additional structurally bound water (Scheller et al., 2021), but this reservoir is effectively isolated from active exchange and is not considered part of the modern mobile water inventory.

The polar ice caps constitute Mars' largest individual ice reservoir, with a combined volume of $\sim 2.5 \times 10^6 \text{ km}^3$ (equivalent to ~22–33 m global equivalent layer, GEL) (Smith et al., 1999; Smith et al., 2001; Plaut et al., 2007; Byrne, 2009). Their layered stratigraphy is widely interpreted to record orbitally modulated climate variability, although the detailed climatic significance of this record remains actively debated (e.g., Smith et al., 2016; Becerra et al., 2019). Beyond the poles, multiple lines of evidence indicate that substantial volumes of shallow (<10 m) subsurface ice are also present at mid to high latitudes. A component of this ice is thought to be pore-filling, occupying the pore space of unconsolidated regolith and forming in situ by vapour diffusion or freezing in equilibrium with the atmosphere (Mellon and Jakosky, 1993; Schorghofer and Aharonson, 2005; Hudson et al., 2007; Bryson et al., 2008). However, radar and observations increasingly evidence the presence of excess ice, whose volume exceeds available pore space and therefore demands additional emplacement mechanisms, such as burial and preservation of atmospherically deposited snow or ice during past orbital states (e.g., Head et al., 2003; Madeleine et al., 2009), development of subsurface ice lenses (Sizemore et al., 2015), or other processes that locally concentrate ice beyond simple vapour-diffusive filling (Fisher, 2005). The presence of both pore-filling and excess ice is supported by converging geomorphic and geophysical observations, including neutron spectroscopy (Boynton et al., 2002; Feldman et al., 2004), thermal inertia measurements (Bandfield, 2007; Mellon et al., 2008), fresh impact exposures (Dundas et al.,

2021), excess ice-exposing scarps (Harish et al., 2020; Dundas et al., 2021), clean ice exposing impact craters at decimetre-to-metre depths (Byrne et al., 2009; Dundas et al., 2021, 2023), widespread polygonal terrain (e.g., Costard and Kargel 1995; Mangold et al., 2005; Mellon et al., 2008) and in situ measurements by the Phoenix lander of both pore and excess ice (Mellon et al., 2009; Smith et al., 2009), indicating that non-polar subsurface ice represents a significant component of the Amazonian cryosphere. Using MARSIS radargrams with probing depths of 60–80 m, Mouginot et al. (2010) inferred the latitudinal extent of ice-bearing regolith and estimated a total ground-ice volume of $\sim 10^6$ km³, equivalent to a polar cap (~ 7 m GEL), potentially rising to ~ 10 m GEL if ice-rich material extends quasi-continuously to depths of ~ 100 m.

A major surficial expression of this same subsurface excess ice system is termed as the latitude-dependent mantle (LDM), a metres-thick, ice-rich deposit widely interpreted as the product of atmospherically deposited ice–dust mixtures emplaced during high-obliquity intervals and subsequently modified by sublimation and degradation (Kreslavsky and Head, 2002; Head et al., 2003; Levy et al., 2009). The LDM covers $\sim 23\%$ of Mars’ surface at mid to high latitudes (Mustard et al., 2001; Kreslavsky and Head, 2002; Levy et al., 2009) and has a volume equivalent to several metres GEL (e.g., Feldman et al., 2004; Lasue et al., 2013).

Although present climatic conditions in the Martian mid-latitudes (~ 30 – 60°) lie outside the latitudinal zone of present surface-ice stability, this zone hosts an extensive population of putative debris-covered glacier collectively termed Viscous Flow Features (VFFs; Milliken et al., 2003). First recognised in Viking images (Carr and Schaber, 1977; Squyres, 1978, 1979), these landforms include Lobate Debris Aprons (LDAs; e.g., Squyres, 1978, 1979; Pierce and Crown, 2003; Mangold, 2003), Lineated Valley Fills (LVFs; e.g., Squyres, 1978, 1979; Levy et al., 2007), Concentric Crater Fills (CCFs; e.g., Squyres, 1978, 1979, 1989; Levy et al., 2009, 2010), and Glacier-Like Forms (GLFs; e.g., Arfstrom and Hartmann, 2005; Souness et al., 2012). Among these, CCFs (Figure 1.2) form the most abundant and globally distributed subset (Levy et al., 2014) and constitute the focus of this Thesis. All VFF subtypes display morphological evidence consistent with ice-rich creep, including convex surface profiles, lobate termini, surface lineations (here termed tracelines), compressional and extensional ridges, and flow around topographic obstacles (e.g., Mangold, 2003; Head et al., 2005). Early estimates of their ice content ranged from debris-rich mixtures to near-pure ice, but orbital radar sounding has since inferred that many LDAs contain very high ice fractions ($> 90\%$) beneath a thin debris cover (Holt et al., 2008; Plaut et al., 2009; Petersen et al., 2018; Gallagher et al., 2021; Perry et al., 2024). Given their morphological similarity to terrestrial glaciers, and their spatial co-existence and physical transitions between forms, these landforms are commonly interpreted as components of an integrated glacial landscape (e.g., Head et al., 2006, 2010; Souness et al., 2012), an association that is consistent with a broadly shared formation history and predominantly ice-rich composition. Understanding how these

glaciers developed requires placing it in the context of Mars' long-term climatic and hydrological evolution.

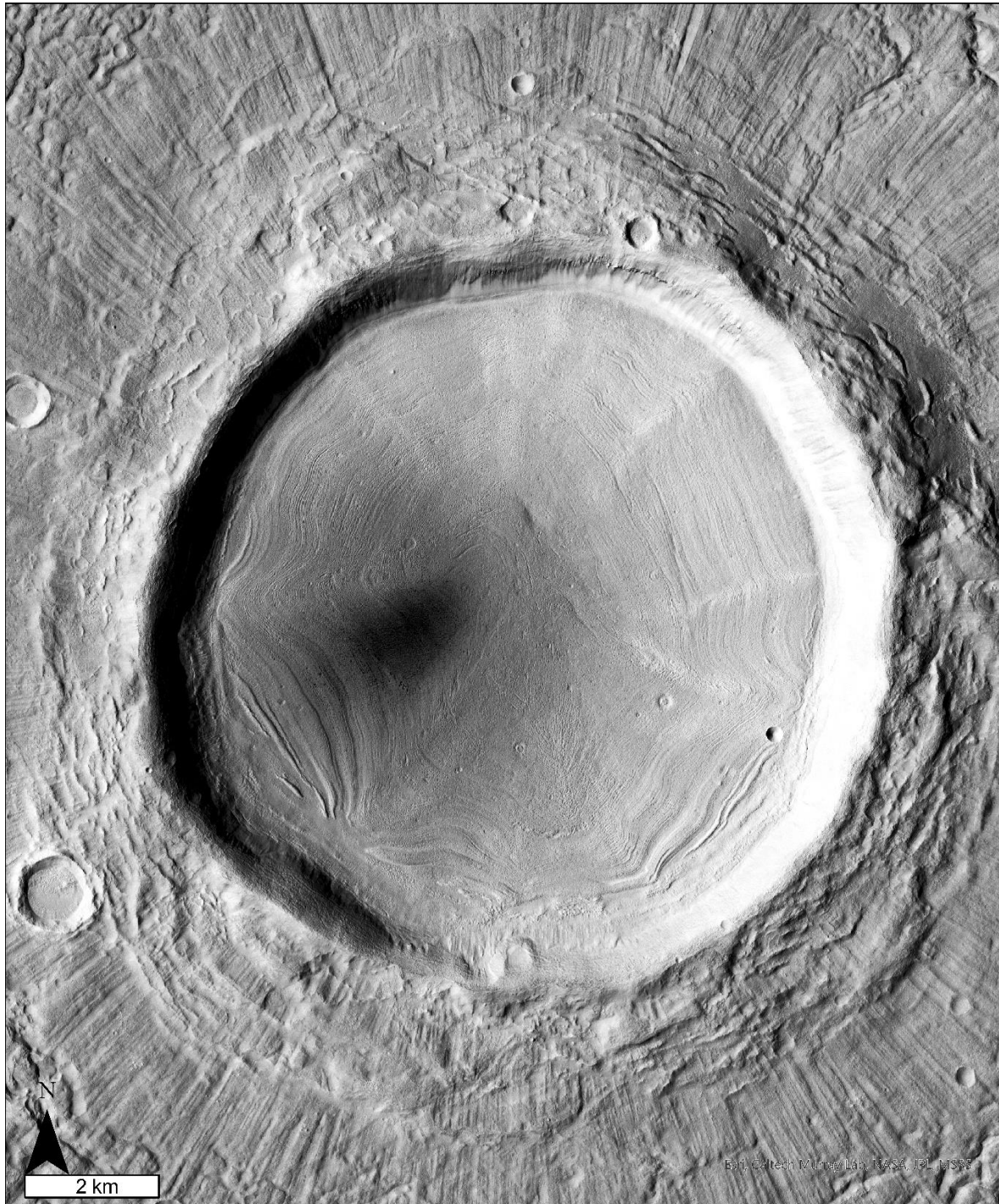


Figure 1.2. Example of Concentric Crater Fill in Utopia Planitia. CCF located at 36.07N, 91.45E. CTX image: B21_017827_2151.

Early in Martian history, climate conditions were markedly different from those of today, with abundant geomorphological evidence indicating widespread surface liquid water and substantially larger inventories of volatiles such as water and CO₂ than are currently preserved (Clifford and Parker, 2001;

Carr and Head, 2010; Di Achille and Hynek, 2010; Kurokawa et al., 2014). Estimates of early exchangeable water inventories range from several hundred metres (Cardenas and Lamb, 2022) to several kilometres GEL (Luo et al., 2017), implying that the current ~30–50 m GEL known water ice reservoir represents only a small residual fraction of the planet’s original water budget (Carr and Head, 2010; Alsaeed and Jakosky, 2019; Jakosky and Hallis, 2024; Jakosky et al., 2026). A substantial fraction of Mars’ early water inventory is thought to have been lost to space during progressive atmospheric escape (e.g., Carr and Head, 2015; Jakosky et al., 2018). As atmospheric pressure and surface temperatures are widely inferred to have declined through time (e.g., Carr and Head, 2015, 2019), liquid water is thought to have become progressively less stable, and a growing fraction of the exchangeable component of Mars’ water inventory is therefore interpreted to have been sequestered into increasingly long-lived ice reservoirs at the surface, and in the subsurface (e.g., Carr and Head, 2015, 2019; Alsaeed and Jakosky, 2019; Jakosky and Hallis, 2024; Jakosky et al., 2026).

At present, atmospheric H₂O on Mars is a minor trace constituent (Titov, 2002), transported primarily on diurnal and seasonal timescales, with net exchange dominated by seasonal frost deposition and sublimation in the polar regions (e.g., Appéré et al., 2011), and is continually depleted by ongoing escape processes (Jakosky, 2021). Under current pressure–temperature conditions, surface and near-surface ice outside the polar regions is thermodynamically unstable unless insulated by regolith or debris, such that the contemporary climate system primarily acts to sublimate and redistribute exposed non-polar ice, as observed at the Phoenix lander site (Mellon et al., 2009; Smith et al., 2009), in fresh impact exposures (Dundas et al., 2021), and at ice-exposing scarps (Dundas et al., 2018).

The persistence of massive debris-covered ice deposits in the Martian mid-latitudes is difficult to reconcile with the modern climate, because surface non-polar ice is unstable (Leighton and Murray, 1966) and long-term stability of shallow pore ice is restricted to high latitudes. Thermal models of shallow ground-ice stability (e.g., Mellon and Jakosky, 1995) address this instability by predicting where pore-filling ice emplaced by vapour diffusion can persist in the regolith, suggesting that the latitudinal limit of stable pore ice migrates equatorward as obliquity increases because the depth to stability progressively decreases until intersecting the surface; decreasing obliquity reverses this behaviour by driving loss at the lowest latitudes first and deepening the stable ice table at higher latitudes. However, thermal models of pore-ice cannot account for the decametre-to-kilometre thicknesses of VFFs. Instead, General Circulation Models (GCMs) are able to model atmospherically deposited snow and ice accumulation at the surface during different obliquity states (e.g., Forget et al., 2006; Madeleine et al., 2009). These GCM predict that intermediate obliquities (~35°) favour net snow and ice accumulation in the mid-latitudes (Mischna et al., 2003; Levrard et al., 2004, 2007; Madeleine et al., 2009), and high obliquities (~>45°) permit equatorial ice stability and even the formation of tropical mountain glaciers (Mischna et al., 2003; Forget et al., 2006). The mid-latitude glaciation hypothesis therefore interprets present-day VFFs as the remnants of thick, atmospherically emplaced

ice bodies formed during these orbital states, subsequently buried by dust or debris and preserved out of equilibrium with the modern climate while undergoing extremely slow, incomplete net loss (e.g., Head et al., 2003, 2005, 2006).

VFFs form the principal suite of landforms recording Amazonian mid-latitude glaciation. Within this broader assemblage, CCFs represent the most numerous and spatially widespread (Levy et al., 2014). Their confinement within impact craters creates a glacier type with no direct counterpart on Earth, where wholly impact-crater-confined glaciers are absent and only partial similarities exist in sub glacial impact crater, such as the crater beneath the Hiawatha Glacier in Greenland (Kjær et al., 2018) and the proposed—but controversial—Wilkes Land impact basin (diameter 510 km) beneath the Antarctic ice sheet (Von Frese et al., 2013; Weihaupt et al., 2015). Within Mars’ mid-latitudes CCFs are described as near-ubiquitous (Levy et al., 2014)—in contrast to the more spatially restricted LDAs, LVFs, and GLFs (Milliken et al., 2003; Souness et al., 2012; Levy et al., 2014)—and their widespread spatial distribution creates an exceptional opportunity for systematic, global population-scale analysis. Because most CCFs are wholly confined within impact craters, whose size and shape scale predictably with diameter (e.g., Garvin et al., 2003), these deposits have evolved within a quantifiable geometric container. This setting means that in a population study of CCFs, the influence of the host crater—namely maximum accommodation space, depth–diameter scaling, and volumetric capacity of the host craters— can be separated from signals of glacial processes. Addressing this requires separating crater-imposed controls from environmental signals, which I do in this research.

This geometric constraint is critical because it allows a fundamental question to be posed: to what extent—if any—do the present-day properties of CCFs record signals of their palaeo–mass balance during active glaciation, or do they instead primarily record post-glacial ablation and degradation, and to what extent are they dominated by the host-crater geometry? The CCF surface depth below the crater rim provides a first-order measure of how full a crater is relative to its geometric capacity, and therefore a way to test whether variations in CCF thickness are influenced primarily by crater size or by external forcing. The preserved (i.e. observed) CCF volume offers a complementary measure on how much ice remains relative to that capacity, while surface slope and aspect offer a means of assessing whether surface geometry reflects accumulation, asymmetric ablation, or purely topographic steering by the host crater.

At population scale, comparisons of CCF surface depth, preserved volume, slope, and aspect across latitude, elevation, and longitude, after accounting for host-crater geometry, provide a spatially distributed test of environmental controls on Martian mid-latitude glaciation. CCFs, and VFFs more generally, are interpreted as debris covered glaciers formed under one or more past climate regimes (e.g., Squyres, 1979; Milliken et al., 2003; Souness et al., 2012; Levy et al., 2014). As such, their modern distribution and morphologies may record a continuum between two end-member histories. If post-

glacial mass loss has been negligible, the present population should approximate the initial distribution, volume, and surface geometry of crater-confined accumulation during CCF formation. If substantial post glacial ablation occurred, the present population instead records where deposits survived and net mass loss became sufficiently slow for long-term preservation under less favourable climate conditions, rather than the full extent of the original population. Distinguishing between these alternative scenarios is necessary to constrain the former spatial extent and magnitude of mid-latitude ice accumulation, providing empirical benchmarks for testing whether General Circulation Model simulations generate sufficient ice across the appropriate regions under plausible past climate states and orbital configurations. Reconstructing CCF evolution consequently informs both Amazonian climate history and the redistribution of water between different latitudes through time.

The broader climatic, orbital, and glaciological context for Martian mid-latitude ice, and the current knowledge about CCF glacial history are reviewed in depth in Chapter 2. Here, having explained why CCFs may provide a unique avenue for studying mid-latitude glaciation, my focus narrows down to the specific questions addressed by this Thesis. The following section therefore sets out the research aim and objectives that structure the analysis in this Thesis.

1.2. Research aim and questions

The overall aim of this Thesis is to expand current understanding of Martian CCFs, their evolution, and their present-day state, and to establish to what extent—if any—their geometry and surface morphology record signals of Amazonian (3 Gyr–present) mid-latitude glaciation as opposed to the geometric constraints and preservation effects imposed by their host craters.

Because CCFs are near-ubiquitous in the Martian mid-latitudes, they represent the most spatially extensive surviving record of glacier-related activity in this region, and their spatial distribution, surface forms, and volumes can be used to test for large-scale environmental controls. Their configuration and distribution therefore represent a particular stage within a longer evolutionary history that is, at least in its current phase, geometrically confined by the host crater. This constraint allows CCFs to be used to test whether observed variations in surface depth below the crater rim, volume, and distribution with latitude, elevation, and regional setting primarily reflects (i) environmental gradients governing ice accumulation and ablation, (ii) the geometric and topographic constraints of the host craters that filter or overprint those climatic signals, (iii) neither, implying a dominant role for local processes and/or post-glacial modification, or (iv) signals of both (i) and (ii).

In this context, the spatial configuration of today's CCFs may preserve signatures of initial ice accumulation, subsequent post-glacial ablation, and longer-term degradation, but it is not presently known which of these processes—if any—dominates their modern morphology, nor how this balance varies with latitude, elevation, and regional setting. This Thesis therefore uses the global distribution,

surface geometry, volumes, and surface slopes of CCFs to attempt to disentangle climatic controls from geometric and preservation controls, and to assess which aspects of Martian mid-latitude glaciation can—and cannot—be robustly reconstructed from these deposits.

These aims are addressed through a chapter sequence that first introduces the scientific context, datasets, and controlled CCF inventory, before moving to the quantitative analyses in Chapters 4–7 and ending with a synthesis in Chapter 8. The remainder of Chapter 1 introduces the datasets, terminology, and thesis structure used throughout the work. Chapter 2 reviews the wider scientific context for Amazonian mid-latitude glaciation, VFFs, and CCF formation and preservation, and identifies the unresolved scientific issues that motivate the thesis. Chapter 3 establishes the controlled, high-confidence CCF inventory required for quantitative comparison. The empirical chapters, Chapters 4–7, then use this basis to test four chapter-level research questions, introduced here in outline and developed in greater detail after the literature review:

Chapter 4: Does the global distribution of high-confidence CCFs reflect environmental controls or the availability of craters suitable for glaciation?

It is well established within the literature that VFFs cluster strongly within the mid-latitudes (e.g., Squyres, 1978, 1979; Milliken et al., 2003; Souness et al., 2012; Levy et al., 2014), yet it remains unresolved whether the spatial distribution of CCFs—the most abundant VFF—reflects glacial–environmental controls or simply the availability of the antecedent topography required to form a CCF i.e. a host crater. In Chapter 4 I determine whether CCF occurrence patterns arise from background crater abundance or whether they record additional environmental gradients such as latitude, elevation, and/or regional relief. To achieve this, I construct a matched control population of non-CCF-hosting craters from the Robbins and Hynek (2012) crater database, filtered to the same diameter range and latitudinal domain as the high-confidence CCF population defined in Chapter 3. This control dataset permits a direct statistical comparison between CCF-hosting and non-hosting craters in the mid-latitudes, allowing environmental signals to be distinguished from the background availability of host craters.

Chapter 5: Why do CCFs have variable surface depths below their rim?

Pedersen and Head (2010) and Levy et al. (2014) documented that CCF surfaces sit at markedly different elevations relative to their crater rims—even between craters of similar size—implying that the present depth of CCF surfaces below the rim reflects a combination of host-crater geometry and glacial accumulation–ablation history. In Chapter 5 I assess whether the variations of these depth deficits, which I term ‘apparent depths’, arise solely from differences in crater depth and volume (which scales with diameter) or whether latitude, longitude, elevation, and regional climatic setting impose additional controls.

Chapter 6: Do preserved CCF volumes record glacial and/or climate controls, or only crater capacity?

Crater depth and overall shape scale predictably with diameter on Mars, transitioning from simple bowl-shaped to more complex, flatter-floored profiles as crater diameter increases (e.g., Garvin et al., 2003). This scaling implies that larger craters have higher volumetric capacity for crater-confined ice. In Chapter 6, I use this geometric framework to test whether the volumes of preserved CCFs are primarily controlled by host-crater size and shape alone, or whether they record additional influences such as spatially variable accumulation, post-glacial ice loss, and/or long-term differences in preservation efficacy. In other words, the analysis asks whether present-day CCF volumes simply reflect what each crater can hold, or whether they retain a more complex record of environmental and glaciological history.

Chapter 7: Do CCF surface slopes and aspects record asymmetric accumulation or ablation?

Global VFF studies consistently show strong pole-facing and secondary east-facing orientation biases (Milliken et al., 2003; Souness et al., 2012), yet no study has examined CCF surface slopes or aspects in isolation. Dickson et al. (2012) inferred flow directions for mid-latitude crater-confined ice features by qualitatively identifying surface tracelines and lobes. Furthermore, their sample combined three morphologically distinct populations (classic CCFs, low-definition CCFs, and pole-facing intracrater flows—essentially crater-confined LDA), therefore preventing assessment of CCFs as a discrete glacial class. Their analysis additionally assumed an orientation-neutral, concentric geometry for each CCF when inferring its flow direction, without quantifying its surface topographic gradient. In Chapter 7, I move beyond this qualitative assumption by measuring the magnitude and direction of the first-order surface slope of high-confidence CCFs. This allows me to investigate whether CCF surfaces exhibit systematic slope–aspect patterns and evaluate whether—if found—these patterns record asymmetric accumulation, differential ablation, or long-term preservation.

1.3. Images and topographic data sources

Table 1.1 lists the available orbital imagery and digital elevation datasets for Mars, including their spatial resolution, footprint size, and approximate global coverage. Datasets used for analysis in this Thesis are shown in bold. Although the datasets and methods used in each analysis are described in the relevant chapter methods sections, the principal datasets are introduced in greater detail below.

Table 1.1. Key freely available visible, infrared, and digital elevation datasets for Mars, with data-access links. Datasets used for analysis in this Thesis are shown in bold. Table adapted from Butcher et al. (2026)					
Acronym	Resolution	Footprint width	Planetary coverage in 2026	Data access link and references	Period of operation
Visible Image Products					
Viking Imagery	232 m	40 km	97%	[1]	1976–1980
MOC	3 to 12 m	~3 km	5.45% at better than 12 m/pixel; ~0.5% at better than 3 m/pixel	[2] Malin et al., 1992	Sep 1997 to Nov 2006
CTX	6 m	~26.5 km	Global	[3], Malin et al., 2007; Dickinson et al., 2024	Mar 2006 to Present
HiRISE	25–50 cm	~5 km	~3.5%	[4], McEwen et al., 2007; 2024	Mar 2006 to Present
HRSC	~15–30 m	~80 km	~Global	[5], Neukum et al., 2004	Jan 2004 to Present
Cassis	5 m	~10 km	~6%	[6], Thomas et al., 2017	Nov 2016– Present
Thermal Image Products					
THEMIS	100 m	Day-IR: Global Mosaic Night IR: 60°N–60°S	Global	[7, 8] Christensen et al., 2004	Mar 2002 to Present

Table 1.1 continued

Table 1.1 continued					
Acronym	Resolution	Footprint width	Planetary Coverage in 2026	Reference and data access link	Period of operation
Digital Elevation Model (DEM) Products					
MOLA	463 m	Global gridded	Global	[9], Smith et al., 2001	Nov 1996 to Nov 2006
HRSC	50-100 m	80 km	~41%	[3], Jaumann et al., 2007	Jan 2004 to Present
-	200 m	Global	Global	[10], Fergason et al., 2018	-
CTX	24 m	26.5 m	? >5000 pairs	[11], Malin et al., 2007; Laura et al., 2021	Mar 2006 to Present
HiRISE	1-2 m	5 km	~0.05% of planet	[12], McEwen et al., 2007; Kirk et al., 2008	Mar 2006 to Present
CaSSIS	4-5 m	9.4 by 47 km	-	[6] Thomas et al., 2017; La Grassa et al., 2026	Nov 2016- Present
[1] https://astrogeology.usgs.gov/search/map/mars_viking_colorized_global_mosaic_232m [2] https://asimov.msss.com/moc_gallery/ [3] https://viewer.mars.asu.edu/viewer/ctx or global mosaic: https://murray-lab.caltech.edu/CTX/ [4] https://www.uahirise.org/hiwish/browse [5] https://maps.planet.fu-berlin.de/#map=3/2074498.35/0 [6] https://observations.cassini.unibe.ch/ [7] https://astrogeology.usgs.gov/search/map/Mars/Odyssey/THEMIS-IR-Mosaic-ASU/Mars_MO_THEMIS-IR-Day_mosaic_global_100m_v12 [8] https://astrogeology.usgs.gov/search/map/Mars/Odyssey/THEMIS-IR-Mosaic-ASU/Mars_MO_THEMIS-IR-Night_mosaic_60N60S_100m_v14 [9] https://astrogeology.usgs.gov/search/map/mars_mgs_mola_dem_463m [10] https://astrogeology.usgs.gov/search/map/mars_mgs_mola_mex_hrs blended_dem_global_200m [11] https://stac.astrogeology.usgs.gov/docs/data/mars/ctxdtnms/ and https://registry.opendata.aws/nasa-usgs-controlled-mro-ctx-dtms/ [12] https://www.uahirise.org/hiwish/maps/dtms.jsp					
Instrument			Spacecraft		
Visual Imaging Subsystem			Viking 1 and Viking 2 Orbiters		
Mars Orbiter Camera			Mars Global Surveyor		
Context Camera			Mars Reconnaissance Orbiter		
High Resolution Imaging Science Experiment			Mars Reconnaissance Orbiter		
High Resolution Stereo Camera			Mars Express		
Colour and Stereo Surface Imaging System			ExoMars Trace Gas Orbiter		
Thermal Emission Imaging System			Mars Odyssey		

Instrument	Space Craft
Mars Orbiter Laser Altimeter	Mars Global Surveyor
High Resolution Stereo Camera	Mars Express
HRSC and MOLA Blend	Mars Global Surveyor & Mars Express
Context Camera	Mars Reconnaissance Orbiter
High Resolution Imaging Science Experiment	Mars Reconnaissance Orbiter
Colour and Stereo Surface Imaging System	ExoMars Trace Gas Orbiter

The following data sources from Table 1.1 are used in this Thesis for analysis.

Mars Orbiter Laser Altimeter (MOLA; Smith et al., 2001).

MOLA on Mars Global Surveyor acquired laser altimetry profiles at sub-kilometre spacing, with footprint scales of 120 m, absolute vertical precision of 30 m, and horizontal precision of 400 m (Smith et al., 2001). I use the standard $1/128^\circ$ gridded product generated from the full mission dataset, which incorporates orbital cross-over corrections and internal consistency adjustments (Neumann et al., 2001; Smith et al., 2001). The gridded MOLA product is derived from irregularly spaced laser-altimetry tracks rather than from uniformly distributed point measurements. As a result, the number of DEM pixels within a feature reflects the raster resolution of the gridded product and not the number of original MOLA shot points used to constrain the elevation surface. MOLA shot spacing varies with latitude, with generally sparser coverage towards the equator and denser coverage at higher latitudes. This limitation is reduced here because the analyses focus on mid-latitude bands rather than equator-to-pole comparisons, but interpolation errors may still affect local topographic measurements, particularly for smaller craters. This dataset is suitable for CCF analysis because it offers a globally consistent geodetic framework that ensures regional elevation comparisons remain statistically valid across the planet. To maintain accuracy, the sample was occasionally subset to overcome resolution limitations for smaller features.

A recently released global topographic model that merges MOLA with High Resolution Stereo Camera (HRSC; Gwinner et al., 2009, 2016) elevations was considered but not adopted. Although HRSC stereo data provide higher native spatial resolution, the blended MOLA-HRSC (Ferguson et al., 2018) uses 200 m sized pixels as a middle ground pixel size between MOLA and HRSC. Note that this means where there is only MOLA coverage, the pixel size is not the true spatial resolution of the data, which remains at 463 m. This is particularly relevant for this Thesis, focusing on mid-latitude glaciation, where

HRSC coverage is incomplete, with many CCF-hosting craters intersecting areas of partial or absent HRSC data. Incorporating the blended product would therefore introduce heterogeneous elevation sources and spatially variable uncertainties across the CCF sample. To preserve a uniform topographic baseline and ensure consistent comparability between features, all analyses in this Thesis use the internally consistent MOLA-derived grid alone.

ConTeXt camera images (CTX; Malin et al., 2007):

The CTX instrument on Mars Reconnaissance Orbiter has been imaging the surface of Mars since 2006, acquiring 5–6 m/pixel coverage in swaths typically sized ~ 30 by < 40 km. Its long operational duration has produced near-global, often repeat, imaging of Mars at 5–6 m/pixel resolution (Dickson et al., 2018), providing the primary basemap for CCF mapping and morphological analysis in this Thesis.

High Resolution Imaging Science Experiment (HiRISE; McEwen et al., 2007):

HiRISE on Mars Reconnaissance Orbiter provides the highest-resolution orbital images available for Mars, typically at ~ 0.25 – 0.5 m/pixel and, from stereo pairs, digital terrain models (DTMs) with horizontal resolutions of ~ 1 – 2 m/pixel. Although coverage is targeted rather than global, HiRISE frames are used throughout this Thesis—where available—to verify CTX-scale interpretations and to examine small-scale CCF surface morphology in greater detail. In this Thesis, HiRISE DTMs are used in Chapter 3 for a small number of case-study CCFs to extract detailed elevation profiles along their surfaces and to examine fine-scale surface geometry that is unresolved in lower-resolution datasets.

Thermal Emission Imaging System (THEMIS); Christensen et al., 2004):

THEMIS aboard Mars Odyssey provides infrared day/night imaging, with the thermal IR imager acquiring multispectral data at ~ 100 m/pixel across nine wavelengths between ~ 6.8 and 14.9 μm . Although designed primarily for compositional and thermophysical analyses, its broad, uniform coverage makes it a useful supplementary dataset for geomorphological mapping. In this Thesis, THEMIS daytime infrared images are used in Chapter 5 to support CTX interpretations.

1.4. Thesis structure

This Thesis provides an exploratory global assessment of CCFs, examining their form, distribution, and morphometric characteristics to understand how these properties encode the environmental and palaeoglaciological conditions that shaped CCF accumulation, modification, and preservation within Mars' mid-latitudes.

The exposition begins with this introductory chapter (Chapter 1), covering data sources and outlining the thesis structure. It is followed by a contextualisation of the study in Chapter 2, which reviews

scientific background in more detail; that material presents an overview of mid-latitude glaciation, an introduction to CCFs focusing on their scientific importance, the outstanding knowledge gaps and finally the key research questions and motivations of this work. Chapter 3 develops and applies my CCF classification framework to the Levy et al. (2014) global inventory, producing the high-confidence CCF dataset analysed throughout the rest of the thesis. In Chapters 4 to 7, I conduct original data analysis, addressing the objectives outlined in Section 1.2. The final chapter (Chapter 8) synthesises the findings and discusses their implications for CCF evolution and Mars' mid-latitude glacial history. Supporting datasets and figures are given in the appendices where needed.

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Chapter 2: Mid-Latitude Glaciation and CCFs

2.1 Mid-latitude glacial history

2.1.1. Geologic and climate history of Mars

Understanding present mid-latitude glaciation on Mars requires placing it within the planet's long-term geological and climatic evolution. Mars formed ~4.5 Ga ago (Carr and Head, 2010a) and its surface history is divided into three broad periods—the Noachian (4.1–3.7 Gyr), Hesperian (3.7–3.0 Gyr), and Amazonian (3.0 Gyr–present)—defined through stratigraphy, crater densities, and surface morphology (Figure 2.1; Scott and Carr, 1978; Tanaka, 1986; Hartmann and Neukum, 2001; Head, 2007; Carr and Head, 2010a). Each epoch reflects distinct geological and environmental conditions, including changes in impact flux, volcanic activity, atmospheric pressure, and the varying stability and distribution of surface and near-surface volatiles. The following sections briefly outline the key geological and climatic characteristics of these epochs (Section 2.1.1.1), then outline the role of orbital forcing (Section 2.1.1.2) and the evolution of the Martian H₂O inventory from the Noachian to the beginning of the Amazonian (Section 2.1.2). This provides the necessary context for the remainder of this chapter by establishing how the long-term evolution of the volatile inventory created the antecedent conditions and water-availability baselines that governed the subsequent development of the Amazonian cryosphere.

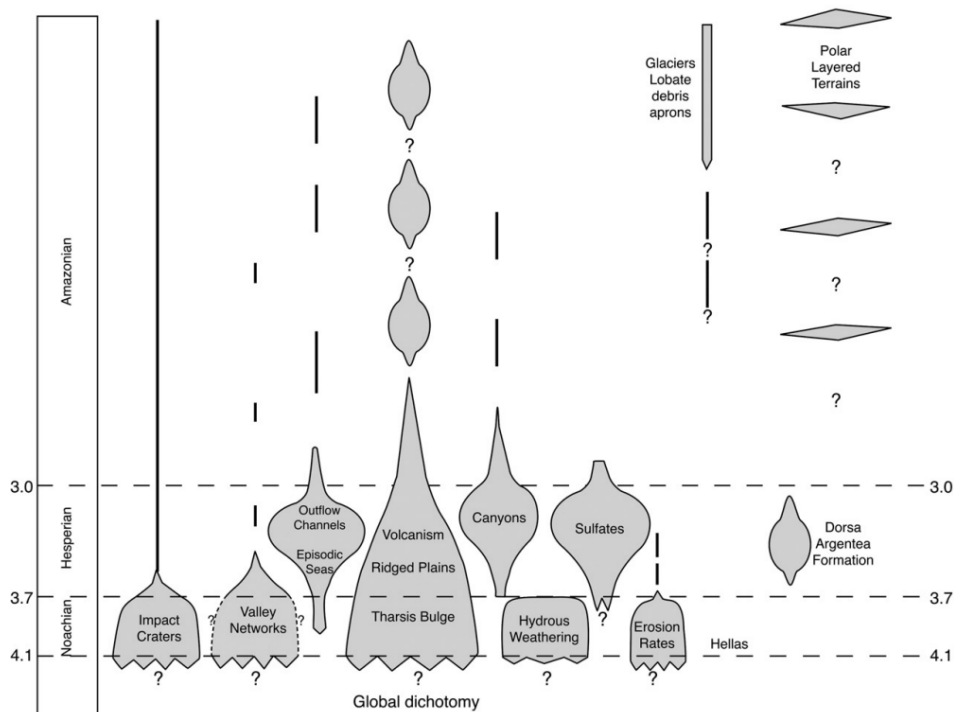


Figure 2.1. Geologic evolution of Mars. Summary of the relative timing and intensity of major geological processes through Martian history, showing the Noachian, Hesperian, and Amazonian epochs and their approximate boundaries. Y-axis denotes time in Gyrs, bandwidths indicate the relative importance of processes over time. Redrawn from Carr and Head 2010a, with dates from Neukum et al. (2001).

2.1.1.1 Mars Geologic Eras

Pre-Noachian (~4.5–4.1 Gyr ago)

Pre-Noachian Mars covers ~4.5 Gyr to the formation of Hellas Basin (~4.1 Gyr) (e.g., Figure 2.1; Carr and Head, 2010a). Much of this surface record has been erased or buried by impacts and resurfacing, limiting direct constraints on climate and hydrological activity (e.g., Carr and Head, 2010a). The main geomorphic legacy is the crustal dichotomy—~5.5 km elevation difference between the northern lowlands and southern highlands (Figure 1.1)—likely formed before major basin formation through impacts or mantle dynamics (McGill and Squyres, 1991; Nimmo and Tanaka, 2005).

The Noachian Epoch (~4.1–3.7 Gyr ago)

The Noachian Epoch is characterised by intense impact cratering, forming the Hellas, Isidis, and Argyre basins (e.g., Strom et al., 1992; Ivanov and Head, 2001). Erosion rates were 2–5 orders of magnitude higher than in the subsequent Hesperian and Amazonian epochs (Scott and Tanaka, 1986; Malin and Edgett, 2001; Craddock and Howard, 2002; Carr and Head, 2010a; Figure 2.1). Substantial volcanic activity, associated with early Tharsis construction, also occurred throughout this epoch (e.g., Phillips et al., 2001; Carr and Head, 2010a). Dense valley network systems and open-basin lakes demonstrate that surface runoff and fluvial erosion were widespread (Craddock and Howard, 2002; Irwin and Howard, 2002; Fassett and Head, 2008). However, whether this fluvial activity reflects a persistently warm and wet climate or instead a predominantly cold climate punctuated by transient melting episodes remains debated (Carr and Head, 2019; Kite and Conway, 2021).

Early climate modelling and geological interpretation framed the Noachian climate in terms of two endmember states: a persistently warm, wet regime capable of sustaining surface liquid water, and a predominantly cold, dry regime in which fluvial activity was limited to episodic melting of ice or groundwater release (e.g., Gaidos and Marion, 2003; Di Achille and Hynek, 2010). More recent syntheses instead view the Noachian as occupying a continuum between these extremes, most plausibly characterised by generally cold background conditions punctuated by transient warmer and wetter intervals driven by volcanism, impacts, and/or orbital forcing (Wordsworth, 2016).

The mineralogical record provides particularly strong constraints on this continuum. The global distribution of phyllosilicates indicates extensive aqueous alteration during the Noachian (Bibring et al., 2006), while early interpretations of widespread olivine preservation were taken to imply spatially limited alteration (Mustard et al., 2005). However, subsequent work has shown that large fractions of Mars' primordial water inventory may have been sequestered into the crust through hydration, implying far more pervasive early alteration than previously recognised (Scheller et al., 2021). In this view, the Noachian likely involved a large, dynamic exchangeable water reservoir that declined rapidly through atmospheric escape and crustal hydration (Scheller et al., 2021), permitting both widespread early

alteration and a progressive transition toward the colder, drier conditions that dominate the Hesperian and Amazonian.

The Hesperian Epoch (~3.7–3.0 Gyr ago)

The Hesperian Epoch (Figure 2.1) is marked by declining impact flux (e.g., Hartmann and Neukum, 2001); reduced erosion rates (e.g., Fassett and Head, 2008); increased volcanism producing extensive flood lavas and ridged plains that resurfaced up to ~30 % of the planet (Greeley and Schneid, 1991; Head et al., 2002) and large-scale tectonism, including the formation of Valles Marineris (e.g., Phillips et al., 2001). A key geochemical transition during the Hesperian is the dominance of sulphates over Noachian phyllosilicates, interpreted as a shift toward acidic, saline, and increasingly arid surface conditions where transient aqueous activity was driven by volcanic outgassing and groundwater rather than sustained surface runoff (Bibring et al., 2006). This transition was also characterised by the accumulation of ice deposits in the subsurface—and perhaps on the surface—as the climate shifted toward global cooling and a thinning atmosphere, with transient warming events insufficient to melt through the subsurface cryosphere (e.g., Clifford, 1993; Carr and Head, 2015; Halevy and Head, 2014).

Hydrologically, the Hesperian transitioned from regionally integrated surface runoff to spatially restricted, episodic water release. Although precipitation-driven activity largely ceased, fluvial activity persisted through catastrophic outflow channels sourced from chaos terrains and graben, including systems associated with the Dorsa Argentea Formation that have been interpreted as recording Hesperian glacial meltwater production (Carr, 1979; Tanaka et al., 2005). These systems record the rapid discharge of vast water volumes, likely released from pressurised aquifers or the melting of ice-rich substrates as the cryosphere thickened in response to the cooling climate (Lucchitta et al., 1992; Ghatan et al., 2005; Carr and Head, 2010a).

The Amazonian Epoch (~3.0 Gyr ago to present)

The Amazonian is the longest of the Epochs but is characterised by persistently low erosion and weathering rates (Golombek et al., 2006), restricted volcanism (Greeley and Schneid, 1991; Hartmann and Neukum, 2001), spatially restricted fluvial activity (e.g., Hynek et al. 2010; Carr and Head, 2010a; Conway et al., 2015) and a cryosphere that has been a primary agent of landscape modification (e.g., Clifford, 1993; Mustard et al., 2001; Milliken et al., 2003; Kreslavsky and Head, 2006; Souness et al., 2012; Levy et al., 2014; Bramson et al., 2017). Climatically, the Amazonian is commonly treated as broadly comparable to present-day Mars—a hyper-arid, cold-desert state with a mean surface temperature of ~-60 °C and a thin, CO₂-dominated atmosphere (~6 mbar) containing only trace amounts of water vapour (H₂O partial pressure ≈ 1 μbar; Clark, 1997)—but geological and glaciological evidence demonstrates that this long interval was punctuated by substantial climate variability and repeated departures from modern conditions.

This long-term climate variability is widely thought to be recorded in the Amazonian cryosphere, which is partitioned among several distinct ice reservoirs, including the polar layered deposits (e.g. Smith et al., 1999; Smith et al., 2001), widespread shallow ground ice at mid- to high- latitudes (e.g., Byrne et al., 2009; Piqueux et al., 2019; Dundas et al., 2021), mid-latitude debris-covered glaciers (e.g., Squyres, 1979; Milliken et al., 2003; Souness et al., 2012; Levy et al., 2014), and perhaps relict equatorial ice deposits (e.g., Shean et al., 2010), though the survival of such ice to present remains speculative. Under present orbital conditions, perennial surface ice is only stable in the polar regions (Leighton & Murray, 1966) and shallow ground ice is thermally stable only at high latitudes ($\geq 60^\circ$; Mellon and Jakosky, 1995). Consequently, the existence of extensive ice bodies at lower latitudes is fundamentally inconsistent with modern climate equilibrium. The distribution of these non-polar ice reservoirs is therefore widely interpreted to reflect climate-driven redistribution of ice and latitudinal migration of stable ice reservoirs over Amazonian timescales. It is theorised that these climate-driven ice redistributions are driven by orbital forcings (e.g., Mustard et al., 2001; Head et al., 2003; Levrard et al., 2004; Fastook et al., 2008), the mechanisms of which are discussed in Section 2.1.1.2.

2.1.1.2 Orbital forcings through Mars' history

Following the transition from an early Mars characterised by endogenic and impact-driven surface modification to the low-erosion, ice-dominated Amazonian surface environment, variations in Mars' orbital parameters provide the principal mechanism linking long-term climate evolution to the redistribution of surface and near-surface ice. Mars' obliquity, eccentricity, and precession (Figure 2.2A) undergo periodic oscillations (e.g., Ward, 1974; Laskar and Robutel, 1993; Laskar et al., 2004). These oscillations regulate the spatial and seasonal distribution of insolation, exerting first-order controls on the stability and migration of ice (e.g., Mellon and Jakosky, 1995; Laskar et al., 2002; Montmessin et al., 2007; Mischna et al., 2013; Figure 2.2B).

The mid-latitude glaciation hypothesis provides a theoretical framework for the deposition of non-polar ice, positing that—unlike on Earth—Martian glacial periods are initiated during intervals of intermediate obliquity ($\sim 35^\circ$) when enhanced polar insolation destabilises perennial ice reservoirs and promotes the equatorward transport of atmospheric water vapour (e.g., Mellon and Jakosky, 1995; Levrard et al., 2004; Montmessin et al., 2007). Under this model, mid-latitude debris-covered glaciers are interpreted as relic, debris-covered remnants of these past climate regimes (Head et al., 2003; 2005). The physical plausibility of this hypothesis has been tested through GCMs, which find that under certain orbital parameters (eccentricity of 0.1) significant mid-latitude ice accumulation can occur during intervals of intermediate obliquity ($\sim 35^\circ$) (Forget et al., 2006; Madeleine et al., 2009).

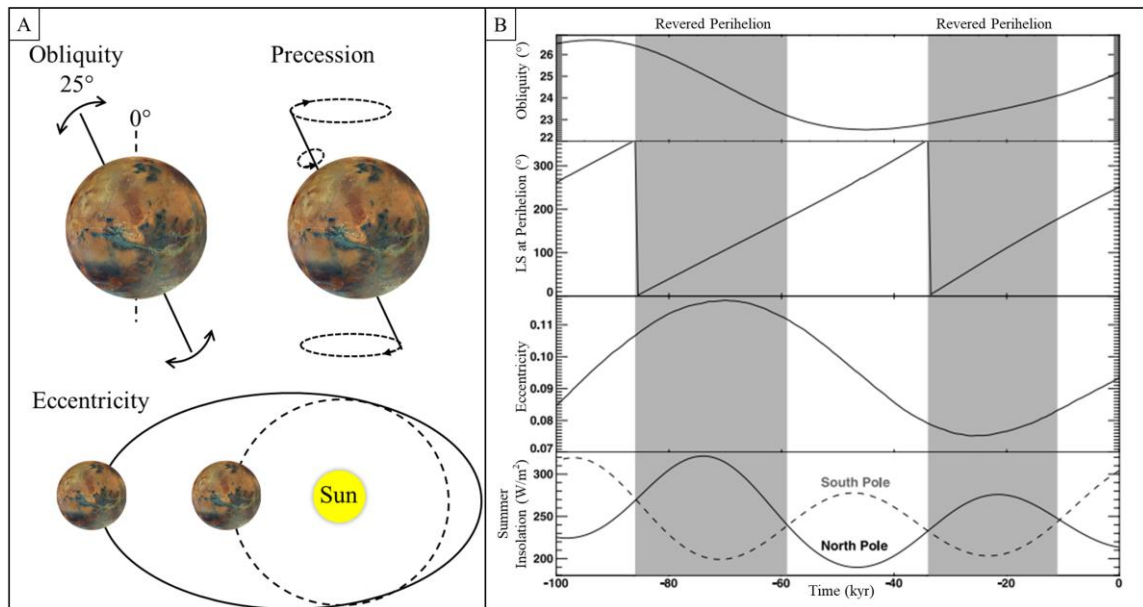


Figure 2.2. Martian orbital parameters and their climatic forcing. (A) Schematic illustrations of Mars' three primary orbital parameters: obliquity, precession, and eccentricity. (B) Time series of Mars' orbital variations over the last 100 kyr, with obliquity shown in the upper panel, precession in the second panel, and eccentricity in the third panel. The lower panel illustrates the combined effect of these parameters on summer solar insolation at the north and south poles. (B) redrawn and modified from Montmessin et al. (2007).

At low obliquity ($<20^\circ$), GCMs predict preferential cold trapping and net ice accumulation at the poles, with limited mid-latitude incursions (e.g., Levrard et al., 2004). At high obliquity ($>45^\circ$), increased polar insolation destabilises polar ice, and numerical modelling instead predicts net accumulation shifting toward lower latitudes or even the equator (e.g., Jakosky et al., 1995). At intermediate obliquity ($\sim 35^\circ$), GCMs similarly predict unstable polar ice but favour net accumulation in the mid-latitudes (e.g., Madeleine et al., 2009). Precession, modulated by orbital eccentricity, further controls the seasonal timing of perihelion and therefore the hemispheric asymmetry of peak summer insolation (e.g., Montmessin et al., 2007), a mechanism that GCMs find dictates hemispheric net ice accumulation during a given orbital configuration (e.g., Madeleine et al., 2009).

Mars' current obliquity ($\sim 25^\circ$) is thought to be a transient state within a historically highly variable range (e.g., Laskar and Robutel, 1993; Laskar et al., 2004). While dynamical solutions resolve obliquity over the last ~ 20 Myr (Figure 2.3A), the system is mathematically chaotic beyond this limit, as minor perturbations lead to divergent solutions that prevent precise orbital reconstruction for deeper time (Laskar et al., 2002, 2004). Within the resolvable 20 Myr window, the record shows a transition around 5 Ma from an intermediate regime ($\sim 35^\circ \pm 10^\circ$) to the current lower regime ($\sim 25^\circ \pm 10^\circ$), with both states characterised by a ~ 120 kyr oscillation (Figure 2.3A; Laskar et al., 2002, 2004). Superimposed on these 120 kyr cycles, precession operates on a ~ 51 kyr cycle, while eccentricity oscillates between ~ 0 and 0.12 over a ~ 2.5 Myr cycle (Figure 2.3B; Laskar et al., 2002, 2004).

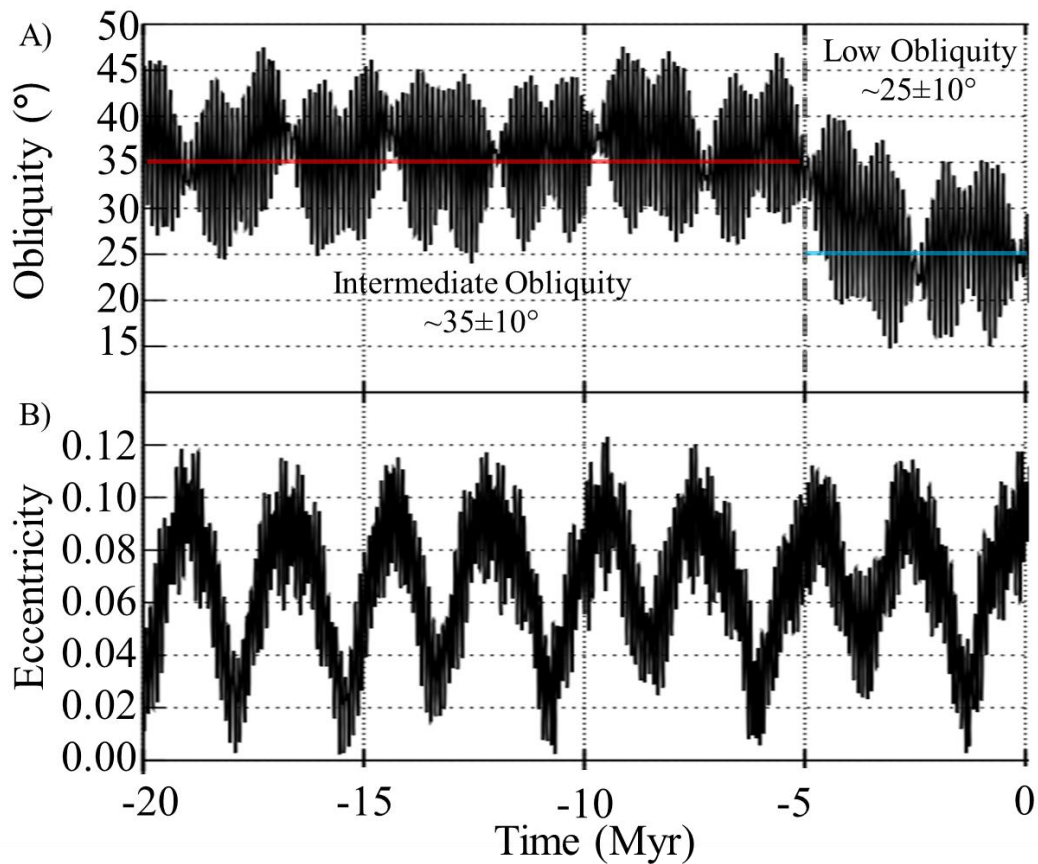


Figure 2.3. Long-term variations in Martian obliquity and eccentricity over the past 20 Myr. (A) Mars' obliquity over the past 20 Myr, showing the marked shift in mean obliquity at ~ 5 Myr ago and the superimposed ~ 120 kyr obliquity oscillations. (B) Mars' orbital eccentricity over the same interval, highlighting the dominant ~ 2.1 Myr eccentricity cycle that modulates the amplitude of seasonal insolation variations. Redrawn and modified Laskar et al. (2004).

Beyond ~ 20 Myr, orbital solutions become non-unique as minor uncertainties in precession amplify through chaotic spin-orbit dynamics (Laskar and Robutel, 1993; Laskar et al., 2004). Probabilistic reconstructions of orbital solutions extending to ~ 250 Myr nonetheless indicate repeated cycle between low-, intermediate-, and high-obliquity states, with the planet oscillating between a minimum of 0° and a maximum of $\sim 65^\circ$ around a long term mean obliquity of $\sim 35\text{--}40^\circ$ (Figure 2.4; Laskar et al., 2004). Mathematical modelling further suggests that although extreme obliquities ($>60^\circ$) were unlikely to persist for long durations, they probably occurred episodically within the past 1 Gyr (Ward, 1973; Touma and Wisdom, 1993; Laskar et al., 2004) with approximately 95% of solutions reaching at least 60° , and $\sim 63\%$ reaching $\geq 60^\circ$ at least once within the past 1 Gyr.

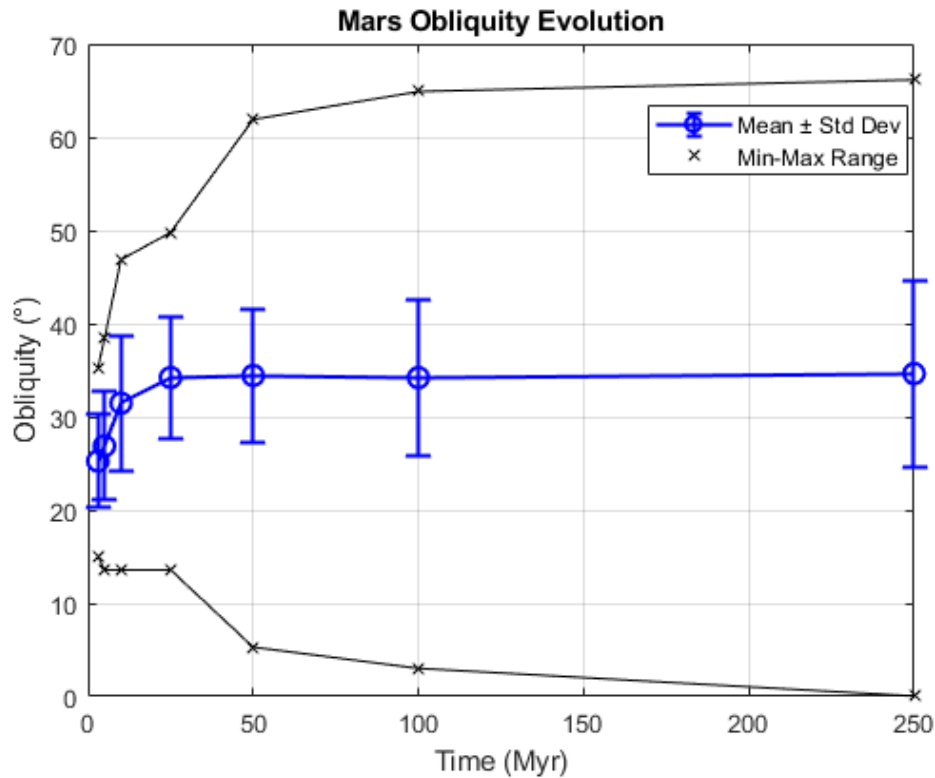


Figure 2.4. Statistical envelope of Martian obliquity variability over the last 250 Myr. Summary of chaotic orbital solutions from Laskar et al. (2004), showing the mean obliquity and $\pm 1\sigma$ dispersion (blue) together with the full minimum–maximum range spanned by the ensemble of solutions (black) as a function of time before present. This figure does not represent a unique or deterministic obliquity history for Mars. Instead, it illustrates the range of obliquity states that are dynamically permissible at each time, given the chaotic nature of the Solar System. Excursions to very high ($> 60^\circ$) and low ($< 10^\circ$) obliquity are therefore possible within this envelope, but their exact timing, duration, and frequency cannot be uniquely reconstructed. Data compiled from Laskar et al. (2004).

Despite the uncertainty, attempts have been made to model and reconstruct the long-term evolution of Mars' orbital parameters beyond 250 Myr. Reconstructions by Laskar et al. (2004) further suggest the probability of obliquity exceeding 80° over 5 Gyr is very low ($\sim 0.015\%$), and over the last ~ 4 Gyr, the mean obliquity was $\sim 37.6^\circ$ with a standard deviation of $\sim 13.8^\circ$. Independent forward modelling of crater ellipticity by Holo et al. (2018) suggests a history dominated by a small number of major obliquity transitions since the Hesperian; specifically, they find that mean obliquity was likely low—between $\sim 10^\circ$ and $\sim 30^\circ$ —and that a low fraction of time was spent above 40° . Their findings remain consistent with probability reconstructions from Laskar et al. (2004), implying that Mars spent most of the Amazonian in relatively low-to-intermediate states, with these periods punctuated by intermediate-obliquity pulses capable of driving significant glaciation.

Ultimately, while these orbital cycles provided the mechanical pacemaker for ice flow, the magnitude of ice accumulation was also constrained by the total volume of water available for transport. To understand the scale and limits of Amazonian glaciation, these orbital drivers must be considered

alongside the long-term depletion of the Martian water budget, examined in the following section (Section 2.1.2).

2.1.2. Evolution of Mars' water inventory through time

The long-term evolution of Mars' climate and cryosphere is governed by the size of its *exchangeable* water inventory, defined as the fraction of planetary H₂O that can be transferred between the atmosphere, surface, and near-surface reservoirs on geological timescales (i.e., excluding water permanently sequestered in hydrated minerals or the deep interior; Carr and Head, 2015; Jakosky and Hallis, 2024). This exchangeable reservoir is reduced primarily by atmospheric escape and by progressive incorporation into crustal hydrated phases, while volcanic outgassing provides only a minor long-term source (Craddock and Greeley, 2009; Carr and Head, 2015, 2019; Scheller et al., 2021). By utilising Global Equivalent Layer (GEL) estimates to quantify changes in the volume, state, and distribution of H₂O, it is possible to reconstruct the path from early Mars' abundance to the restricted budget available during the Mid-Late Amazonian for CCF formation (Figure 2.5).

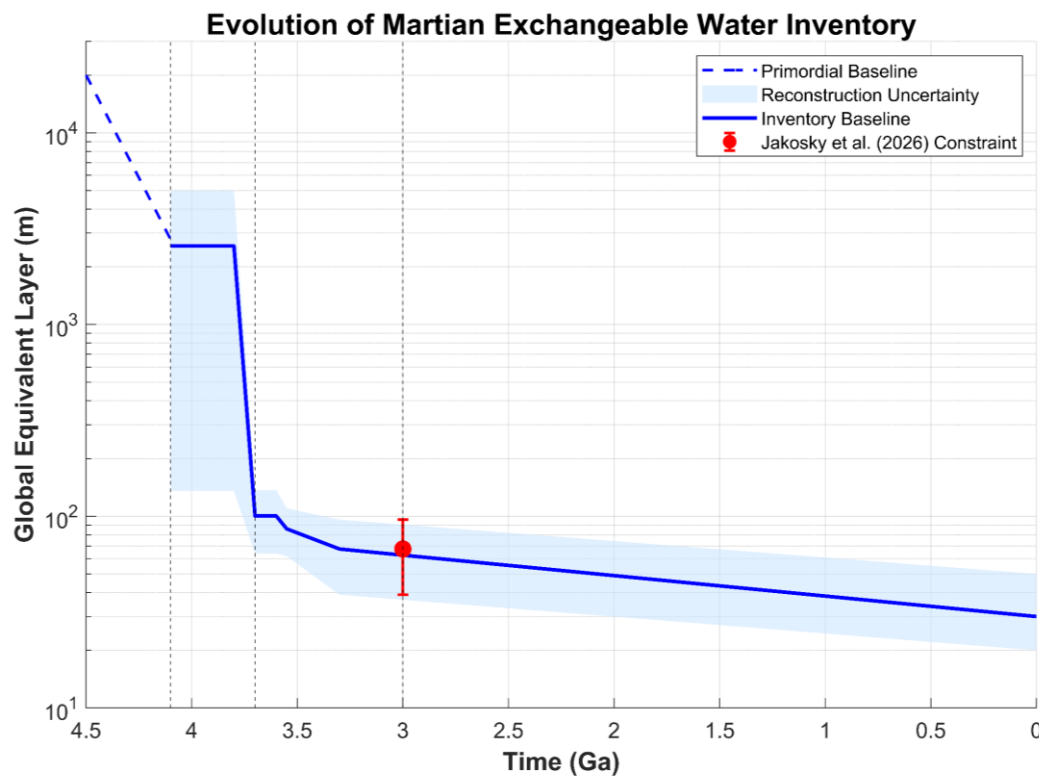


Figure 2.5. Reconstructed evolution of the Martian exchangeable water inventory (GEL) from the Pre-Noachian (4.5 Ga) to the present. The dashed region (4.5 – 4.1 Ga) represents the theoretical primordial value. The blue solid line (4.1 – 0 Ga) represents the baseline inventory, with the shaded region indicating the uncertainty range. The red marker at 3 Ga indicates the isotopic constraint from Jakosky et al. (2026), representing a restricted late-Hesperian/early-Amazonian budget 1.3–3.2 times the current inventory (30 m GEL). Y-axis plotted on a logarithmic scale. Data obtained from Carr and Head, 2010a, 2015, 2019; Luo et al., 2017; Scanlon et al., 2018; Alsaeed and Jakosky, 2019; Scheller et al., 2021; Jakosky and Hallis, 2024; Jakosky et al., 2026.

The initial hydration of Mars, like that of the other terrestrial planets, is thought to have been dominated by exogenous delivery from asteroids, comets, and planetary embryos (e.g., Morbidelli et al., 2000; Lunine et al., 2003; Brasser, 2012). Owing to its proximity to the asteroid belt, Mars is inferred to have acquired much of its water from bodies forming at $\sim 2\text{--}3$ Astronomical Unit (AU) (e.g., Lunine et al., 2003; Brasser, 2012; Lammer et al., 2013). Formation models suggest an initial water content of $\sim 0.1\text{--}0.2$ wt.% (weight percent) (e.g., Lammer et al., 2013; Brasser, 2012), exceeding that of Earth and Venus and implying a pre-Noachian inventory equivalent to tens of kilometres GEL (Carr and Head, 2010a; Figure 2.5). However, this primordial reservoir is poorly constrained, and quantitative reconstructions therefore typically begin in the Noachian, when the exchangeable inventory becomes meaningfully recorded in the geological record (Jakosky and Hallis, 2024).

The Noachian is widely regarded as the interval during which Mars hosted its largest geologically constrained exchangeable water inventory (Carr and Head, 2015). Geological and mineralogical evidence indicates both abundant surface liquid water and extensive cryospheric storage during this period, expressed by valley network incision, basin-scale drainage systems, and widespread phyllosilicate formation (Carr and Clow, 1981; Fassett and Head, 2008; Andrews-Hanna and Lewis, 2011; Bibring et al., 2006). Proposed Noachian inventories span a wide range, from 136 m to 5 km GEL, depending on whether climate states are interpreted as predominantly cold and icy or intermittently warm and wet (e.g., Scanlon et al., 2018; Rosenberg et al., 2019; Kamada et al., 2022; Cardenas and Lamb, 2022; Luo et al., 2017; Figure 2.5). Isotopic constraints and crustal hydration budgets further indicate that a large fraction of this inventory was subsequently removed by atmospheric escape and by sequestration into hydrated minerals during and after the Noachian (Sun and Milliken, 2015; Wernicke and Jakosky, 2021; Scheller et al., 2021).

By the Hesperian, Mars had transitioned into a dominantly cold climate state in which ice, rather than liquid water, became the primary form of surface and near-surface storage (Carr, 1979; Clifford, 1993; Carr and Head, 2015, 2019). Although episodic, high-magnitude outflow events continued to mobilise substantial volumes of water to the surface (Carr, 1979; Manga, 2004), this water was short-lived and rapidly redistributed between the atmosphere and cold-trap ice reservoirs through freezing, sublimation, and evaporation, with additional losses to space and sequestration into the subsurface (Clifford, 1993; Carr and Head, 2015, 2019). By the end of the Hesperian, multiple independent constraints indicate that the exchangeable reservoir had declined to at most a few tens of metres GEL, likely only $\sim 1.3\text{--}3.2$ times the present inventory (Carr and Head, 2015; Jakosky et al., 2026; Figure 2.5).

2.2. The Amazonian cryosphere

During the Amazonian, nearly all of Mars' contemporary exchangeable water is thought to reside within the cryosphere (Clifford, 1991; Clifford et al., 2010; Carr and Head, 2015; Weiss and Head, 2017). Current estimates place the cryosphere inventory at $\sim 20\text{--}50$ m GEL of liquid water (Carr and Head,

2015, 2019; Alsaeed and Jakosky, 2019; Jakosky and Hallis, 2024; Jakosky et al., 2026), with a negligible contribution ($\sim 10 \mu\text{m GEL}$) stored as atmospheric water vapour (Titov, 2002; Jakosky, 2021), which represents only a small fraction of Mars' earlier Noachian water budget and reflects a diminished remainder of the inventory carried over from the end of the Hesperian (Figure 2.5). This inventory is distributed among polar ice deposits, ground ice at mid- to high- latitudes, relict glaciers preserved within debris-covered landforms across the mid-latitudes and, in some cases, equatorial regions. Hydrated minerals retain additional structurally bound water (Scheller et al., 2021), although this reservoir is effectively isolated from active exchange.

The following sections examine the Amazonian cryosphere, outlining first its non-mid-latitude components (Section 2.2.1), then mid-latitude ice deposits ($\sim \pm 30\text{--}60^\circ$) excluding viscous flow features (Section 2.2.2), before focusing specifically on mid-latitude viscous flow features (Section 2.2.3).

2.2.1. Polar and equatorial ice

2.2.1.1. Polar layered deposits

The polar layered deposits (PLDs) of Mars, first observed in detail by the Mariner missions (Figure 2.6), consist of stratified icy layers with varying dust-to-water ice volume mixing ratios that are thought to preserve a record of astronomically driven climate variations over Mars' recent orbital history (e.g., Murray et al., 1973; Laskar et al., 2002; Byrne, 2009; Smith et al., 2016; Becerra et al., 2019). These deposits possess a combined volume of ~ 2.5 million km^3 (22–33 m GEL) and represent the planet's largest permanent water reservoirs that remain fully exchangeable with the atmosphere under present climatic conditions (Smith et al., 1999; Smith et al., 2001; Plaut et al., 2007; Byrne, 2009), a volume comparable to that of Earth's Greenland ice sheet (~ 2.6 million km^3 ; Byrne, 2009). The northern PLD consists of $>95\%$ water ice, whereas the southern PLD contains $\sim 90\%$ water ice, with the remaining $\sim 5\text{--}10\%$ in each case comprising lithic material and atmospherically-derived dust (Smith, 2022); both are seasonally mantled by a thin veneer of CO_2 ice in winter (e.g., Forget et al., 1998; Appéré et al., 2011). The longevity of this CO_2 veneer differs between the hemispheres, with a perennial cover in the south that insulates the underlying water ice from atmospheric exchange under present conditions, and a purely seasonal winter cover in the north that is lost each summer (e.g., Appéré et al., 2011; Portyankina et al., 2012).

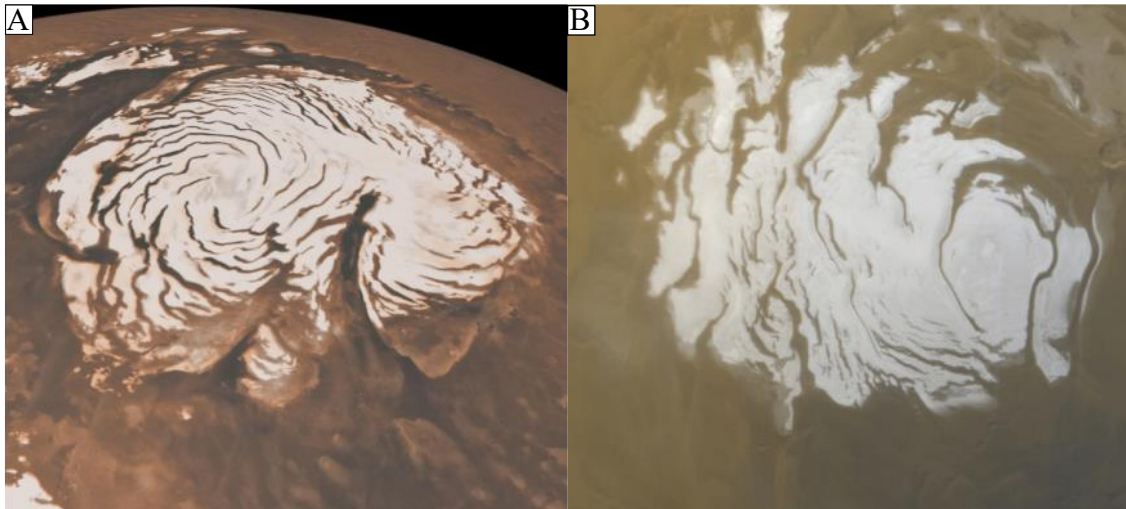


Figure 2.6. The polar layered deposits by NASA's Mars Global Surveyor. A) North polar cap. B) The south polar cap taken in summer. Photo credit: NASA /JPL-Caltech/MSSS

2.2.1.2. Equatorial ice

Although water ground ice is thought to be unstable at any depth near the Martian equator under present climatic conditions (e.g., Schorghofer and Aharonson, 2005; Schorghofer, 2007), multiple lines of evidence indicate that the Amazonian cryosphere extended beyond the polar and mid-latitude regions and that ice was intermittently emplaced at equatorial latitudes during periods of high obliquity (e.g., Forget et al., 2006). Lobate-ejecta craters in equatorial terrains record excavation into ice-rich substrates (e.g., Squyres et al., 1992), while fan-shaped deposits on the western and north-western flanks of the Tharsis Montes (Figure 2.7) and Olympus Mons are widely interpreted as remnants of cold-based mountain glaciers (e.g., Zimbelman and Edgett, 1992; Head and Marchant, 2003; Shean et al., 2005; Shean et al., 2007; Fastook and Head, 2008). GCM simulations under high obliquity ($\sim 45^\circ$) predict surface ice accumulation in equatorial zones, sourced from the northern polar cap (e.g., Forget et al., 2006). These GCM results suggest these deposits may have formed through repeated episodes of high obliquity ($>45^\circ$) during the Late Amazonian (e.g., Shean et al., 2005; Forget et al., 2006; Fastook and Head, 2008). The presence of ring-mould craters superposed on the Arsia and Pavonis Mons tropical glacier deposits has been interpreted as evidence for excavation into buried remnant ice beneath a sublimation-lag or tephra mantle, with crater-size distributions suggesting minimum burial depths of ~ 10 – 20 m and possibly much greater thicknesses (Head and Weiss, 2014). The absence of SHARAD radar detections (Campbell et al., 2013) would require that any such ice—if present at all—occur as a tephra–ice mixture or in a form that does not produce a diagnostic subsurface reflector (Head and Weiss, 2014).

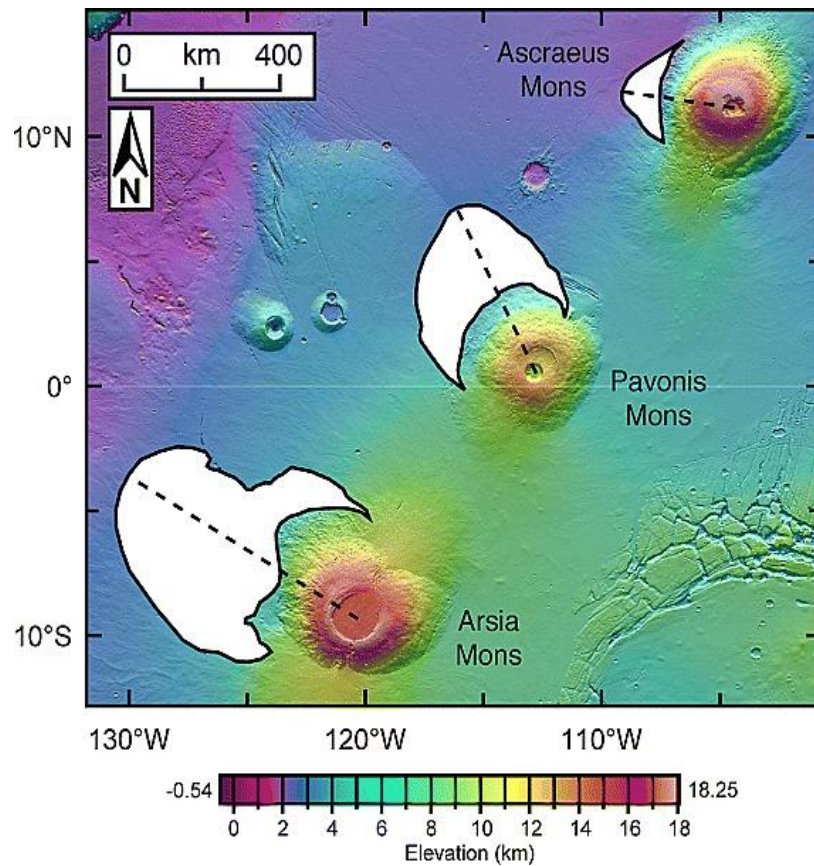


Figure 2.7. MOLA topography of the Tharsis Montes showing the distribution of fan-shaped deposits. Shaded-relief map derived from MOLA data at 128 pixels per degree in simple cylindrical projection. Fan-shaped deposits are mapped as white areas with solid outlines on the western to north-western flanks of Arsia Mons, Pavonis Mons, and Ascraeus Mons. Reproduced from Shean et al., 2005.

Additional observations have also been proposed to support the former presence and/or long-term preservation of low-latitude ice reservoirs beyond the large tropical glacier deposits themselves. Radar sounding has been proposed to indicate the possible presence of deeply buried (~300–600 m) ice-rich units beneath equatorial deposits—most notably within the Medusae Fossae Formation—consistent with a two-layer structure in which any preserved ice would be isolated beneath a thick, dry cover rather than occurring as near-surface ice (Watters et al., 2007; Wilson et al., 2018; Watters et al., 2024). Unlike the ring-mould crater evidence, which pertains to decametre-scale burial depths and remains undetectable to orbital radar, these radar-based interpretations refer to much deeper (>300 m) subsurface units and, if correct, would imply a fundamentally different, deeply buried and isolated mode of low-latitude ice preservation rather than shallow or near-surface ice.

Comparable evidence for former equatorward ice-rich crater fills, and their subsequent degradation, has been suggested by Shean (2010) within a small number of craters in the Sinus Sabaeus region (Figure 2.8). There, crater infills display concentric and curvilinear ridges and furrows, and central depressions, perhaps indicating ice deformation during emplacement and/or subsequent evolution. These features appear to be relatively young, as they cut talus on crater walls and are superposed only by dunes, and

they may delineate former fill margins prior to or during removal, implying hundreds of metres of lateral retreat associated with surface lowering. The implied removal has been attributed to a combination of surface erosion and subsurface volatile loss (e.g., sublimation), with collapse pits and depressions supporting the latter mechanism. However, it remains unclear from available data whether any relict ice is currently preserved within these crater fills (Shean, 2010).

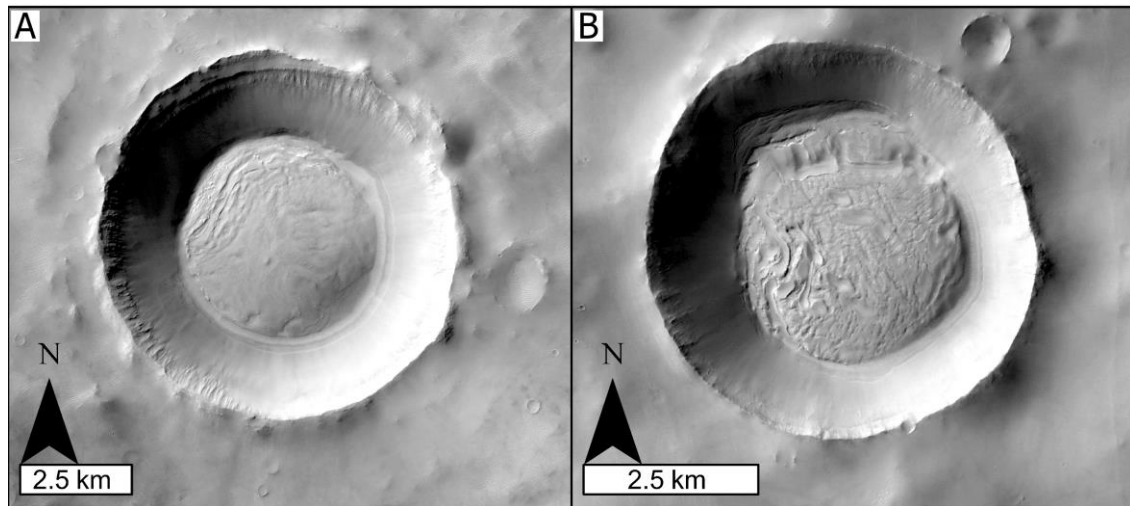


Figure 2.8. Representative examples of equatorial crater fill in the Sinus Sabaeus region. (A) 8.60°S, 347.69°W; CTX image: B19_017184_1711, (B) 8.66°S, 343.93°W; CTX image: G01_018542_1713. Redrawn and modified from Shean, 2010

2.2.2. Mid- and high- latitude subsurface ice and mantling deposits

Beyond the polar layered deposits and equatorial relicts, a substantial fraction of the Amazonian cryosphere is thought to be stored at mid to high latitudes within the subsurface (e.g., Mouginot et al., 2010; Viola et al., 2015; Bramson et al., 2015). In this context, ground ice refers to ice contained within the regolith or shallow crust, and it occurs in several physically distinct forms that imply different emplacement mechanisms and climatic significance. The dominant component is generally inferred to be pore-filling ice, which occupies the pore space of unconsolidated material and forms in situ, primarily by vapour diffusion or freezing in equilibrium with the atmosphere (Mellon and Jakosky, 1993; Schorghofer and Aharonson, 2005; Hudson et al., 2007; Bryson et al., 2008). This pore ice is thought to be geologically young (~0.4 to 2.5 Myr; e.g., Dundas et al., 2023) and highly responsive to orbital and climatic variations (e.g., Head et al., 2003; Bryson et al., 2008;), accumulating and sublimating on geologically short timescales as atmospheric conditions change.

Thermal models of vapour-diffusive ice accumulation in the regolith predict a strong latitudinal control on near-surface pore-ice stability, with shallow (<1 m) stable ice at high latitudes and progressively deeper, or absent, stable ice toward lower latitudes (e.g., Leighton and Murray, 1966; Mellon and Jakosky, 1995; Mellon et al., 2004; Schorghofer and Aharonson, 2005; Piqueux et al., 2019; Figure 2.9).

In the northern hemisphere, modelled near-surface stability extends regionally to $\sim 35^\circ\text{N}$, whereas in the southern hemisphere it is more restricted, poleward of $\sim 55^\circ\text{S}$ (Piqueux et al., 2019; Figure 2.9). These models, however, address only pore ice formed by vapour diffusion and do not treat snow accumulation or the emplacement and stability of thick ice bodies at decametre to hectometre depths. They therefore constrain where shallow pore ice can be stable today, not the full distribution of subsurface ice.

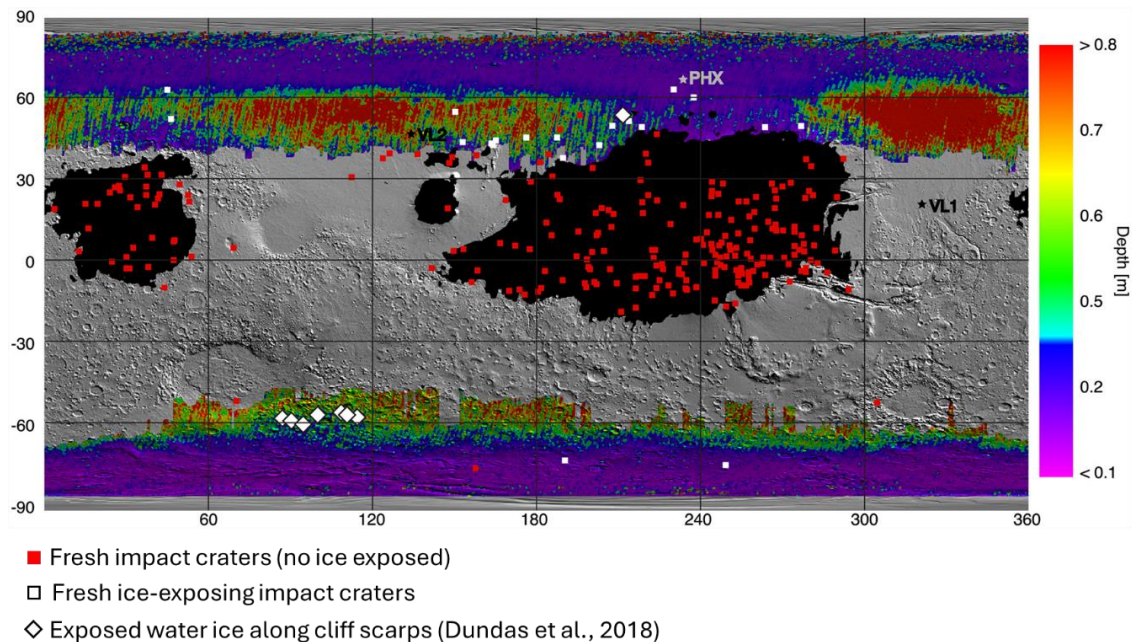


Figure 2.9. Depth to the top of the near-surface ground-ice table derived from MCS thermal data.

Map showing the modelled depth to the top of the water-ice table, overlain on MOLA shaded relief. Regions of low thermal inertia considered unsuitable for landing are masked. The locations of the Phoenix (PHX; Figure 2.10), Viking Lander 1 (VL1), and Viking Lander 2 (VL2) landing sites are indicated. White squares mark fresh impact craters that expose near-surface ice (e.g., Figure 2.11A, B), red squares indicate craters where ice was not identified, and white diamonds denote exposures of water ice in scarp faces (Dundas et al., 2018). Reproduced from Piqueux et al. (2019).

Orbital observations indicate that near-surface ice is in fact widespread at high latitudes and regionally present at mid-latitudes. Neutron and gamma-ray spectroscopy reveal extensive hydrogen-rich material in the upper ~ 1 m of the surface, interpreted as either hydrated minerals or water pore ice, poleward of $\sim 55\text{--}60^\circ$ in both hemispheres, with regional extensions to lower latitudes (Boynton et al., 2002; Feldman et al., 2004, 2011; Wilson et al., 2018; Pathare et al., 2018). Thermal infrared observations further suggest that this near-surface ice is locally pervasive across high-latitude terrains (Bandfield, 2007; Mellon et al., 2008). The extent and apparent abundance of this ice exceed what is expected from pore-ice stability considerations alone.

There is therefore increasing evidence for the presence of excess ice, defined as ice whose volume exceeds available pore space and thus requires an additional external supply (e.g., Fisher, 2005; Viola et al., 2015; Bramson et al., 2015). Proposed formation mechanisms include burial and preservation of

snowpack or glacier ice from past orbital periods (e.g., Head et al., 2003; Levrard et al., 2004; Schorghofer, 2007; Holt et al., 2008; Schorghofer and Forget, 2012; Madeleine et al., 2009, 2014), development of subsurface ice lenses (Sizemore et al., 2015), or other processes that locally concentrate ice beyond simple vapour-diffusive filling (e.g., Fisher, 2005). Pore ice and excess ice forms are not mutually exclusive, as demonstrated by the Phoenix lander at 68°N, which exposed both pore-filling ice and relatively pure excess ice within centimetres of the surface (Mellon et al., 2009; Smith et al., 2009; Figure 2.10).



Figure 2.10. Ground ice exposed by trenching at the Phoenix lander site. Shallow subsurface ice exposed at ~68°N during Phoenix trenching operations. The trench is ~22 cm wide, ~35 cm long, and ~7–8 cm deep at its deepest point, and exposes bright, near-surface water ice. Image credit: NASA/JPL-Caltech/University of Arizona/Texas A&M University.

Further evidence for excess ice comes from fresh impact craters, which have repeatedly exposed clean ice at decimetre-to-metre depths at latitudes >40° in both hemispheres (Byrne et al., 2009; Dundas et al., 2021, 2023; Figure 2.11A, B). When exposed, this ice sublimates rapidly and leaves behind a protective lag deposit that strongly suppresses further loss (Dundas et al., 2021). These sublimation lags permit long-term ice survival even where shallow pore ice is not predicted to be stable today, indicating that thermal-model stability limits describe where pore ice can form under current conditions, not the full distribution of inherited subsurface ice.

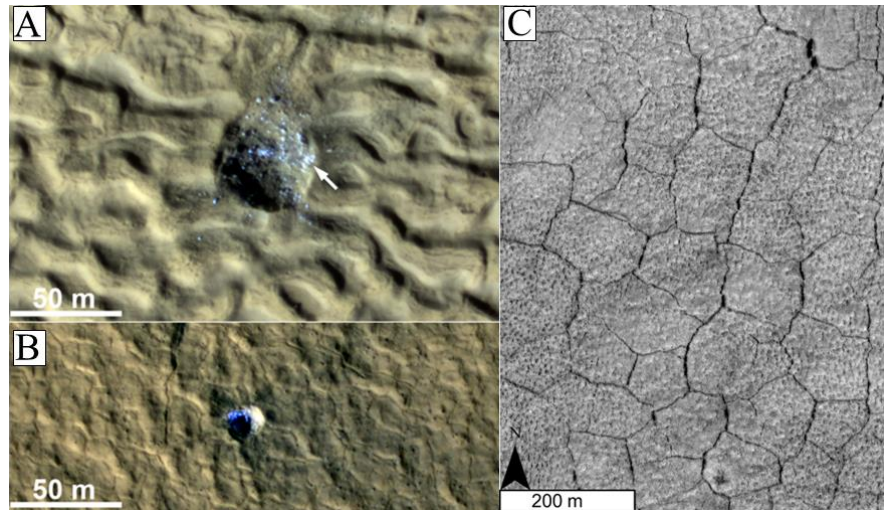


Figure 2.11. Geomorphologic indicators of shallow ground ice in the Martian mid-latitudes. (a–b) Fresh impact craters exposing bright, clean subsurface ice in the Protonilus Mensae region, reproduced from Dundas et al. (2021): (A) HiRISE image ESP_046707_2220 (41.4° N, 48.8° E; Ls = 185°) and (B) HiRISE image ESP_046747_2220 (41.9° N, 36.4° E) redrawn from Dundas et al., 2021. (C) Polygonised terrain in Utopia Planitia. 42.17° N 89.99° E; HiRISE image: ESP_017260_2225.

An important geomorphologic indicator of the shallow subsurface ice system is the widespread occurrence of polygonal terrain (e.g., Costard and Kargel 1995; Mangold et al., 2005; Mellon et al., 2008; Figure 2.11C). On Earth, polygonal patterned ground forms in periglacial environments primarily by thermal contraction cracking of ice-rich substrates and is widely used as a diagnostic indicator of shallow ground ice (e.g., Plug and Werner, 2002). On Mars, morphologically similar polygonal terrain is distributed across the mid- to high-latitudes and covers tens of millions of square kilometres of the surface (e.g., Costard and Kargel, 1995; Mangold, 2005; Mellon et al., 2008; Figure 2.11C). The global distribution of small-scale polygons closely correlates with independently inferred near-surface ice abundance from neutron spectroscopy (Mangold et al., 2005), linking their occurrence directly to the presence of ground ice rather than to purely superficial surface processes. Although polygon morphologies indicate formation by a combination of mechanisms including thermal contraction cracking, frost heave, and sublimation-driven modification (e.g., Mellon, 1997, 2008; Seibert and Kargel, 2001; Marchant and Head, 2007; Morgenstern et al., 2007; Balme et al., 2008; Lefort et al., 2009), their sheer spatial extent, terrestrial analogues, and correspondence with independent ice detections together indicate that much of the Martian mid- to high-latitude surface is underlain by ice-rich ground.

A major surficial expression of this same shallow subsurface ice system is the latitude-dependent mantle (LDM), a widespread, ice-rich deposit containing characteristic polygonal patterns that smooths pre-existing topography and covers approximately ~23% of Mars' surface at mid to high latitudes (Mustard et al., 2001; Kreslavsky and Head, 2002; Levy et al., 2009; Figure 2.12). The LDM is most continuous poleward of ~60°, becomes increasingly patchy between ~30° and 60°, and is largely absent equatorward of ~30° latitude (Kreslavsky and Head, 1999, 2000, 2002; Head et al., 2003; Milliken et

al., 2003). Thicknesses are typically metres to tens of metres, and the unit contains substantial ice, equivalent to several metres GEL, making it a major component of Mars' shallow subsurface ice inventory (Feldman et al., 2004; Lasue et al., 2013). Crater retention ages suggest that high-latitude mantle deposits are generally young ($\lesssim 1$ Ma), while discontinuous mid-latitude occurrences may locally reach ages of up to ~ 8 Ma (Head et al., 2003; Levy et al., 2009; Kreslavsky et al., 2011; Schön et al., 2012).

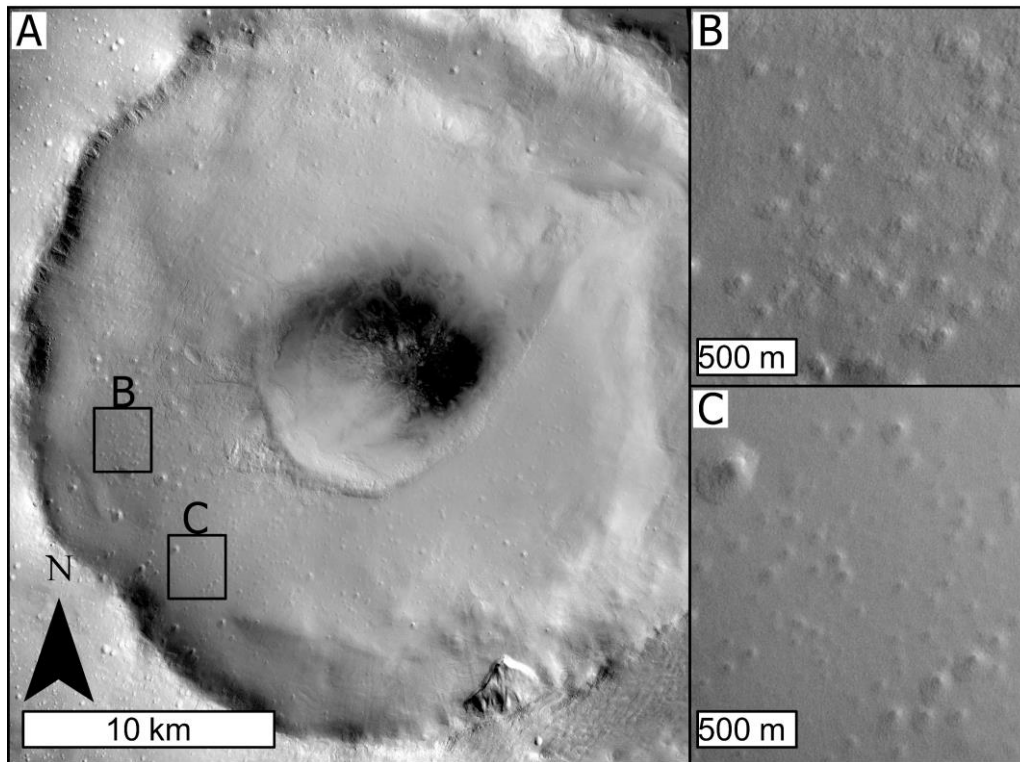


Figure 2.12. LDM superimposed on crater fill in southeastern Arcadia Planitia. (A) Crater interior smoothed and mantled by the LDM. (B, C) Close-ups showing pits and degraded textures within the mantling deposit (51.4° N, 139.2E°; CTX image: P18_007880_2314).

The LDM is widely interpreted as the product of geologically recent, climate-driven ice deposition from the poles toward the mid-latitudes during periods of high obliquity (e.g., Head et al., 2003; Forget et al., 2006; Schorghofer, 2007). Although vapour diffusion into the regolith was initially proposed as a formation mechanism (Mellon and Jakosky, 1995), the observed ice contents and thicknesses are more consistent with direct atmospheric deposition of snow or frost, subsequently stabilised by a thin dust mantle or sublimation lag (Mustard et al., 2001; Milliken et al., 2003; Conway and Balme, 2014). Mantled terrains display pits, knobs, remnant mesas, and scalloped depressions indicative of ongoing ice loss and surface lowering (Mustard et al., 2001; Head et al., 2003; Séjourné et al., 2011), indicating that the LDM is an evolving component of the shallow cryosphere rather than a static deposit.

Orbital and modelling studies suggest that the shallow subsurface ice system has been repeatedly reorganised by obliquity-driven climate variability, emplaced through a combination of snowfall and

subsurface ice directly deposited as pore ice and preserved through a sublimation lag formed during glacial retreat (Schorghofer and Aharonson, 2005; Mellon et al., 2008; Schorghofer and Forget, 2012; Piqueux et al., 2019). By integrating the observed latitudinal extent of ground ice, Mouginot et al. (2010) estimated a total ice volume of $\sim 10^6$ km³ (~ 7 m GEL), potentially rising to ~ 10 m GEL if ice-rich material extends quasi-continuously to depths of ~ 100 m. However, despite these first-order volumetric constraints, the internal vertical structure of this reservoir and its physical partitioning between pore ice, excess ice, and buried ice bodies remain poorly constrained owing to a fundamental observational gap at metre–decametre depths (Lavagna et al., 2022).

2.2.3. Viscous flow features

Mars' mid-latitudes ($\sim \pm 30^\circ$ – 60°) host a diverse suite of ice-related landforms whose morphology and surface textures closely resemble terrestrial debris-covered glaciers (Squyres, 1979; Milliken et al., 2003; Dickson et al., 2012; Souness et al., 2012; Levy et al., 2014). First recognised in Mariner and Viking images as anomalously smooth, lobate, and lineated deposits suggestive of viscous deformation (Sharp, 1973; Squyres, 1978, 1979), these buried ice deposits are now collectively termed Viscous Flow Features (VFFs; Milliken et al., 2003). The mid-latitude distribution of VFFs is at odds with thermal model predictions of shallow pore ice stability under present climatic conditions (e.g., Mellon and Jakosky, 1995; Mellon et al., 2004; Schorghofer and Aharonson, 2005), indicating that they are not in equilibrium with the modern climate and resulting in their interpretation as debris-covered glaciers (e.g., Head et al., 2005; Marchant and Head, 2007).

VFFs are subdivided into four principal classes—lobate debris aprons (Section 2.2.3.1; Figure 2.13A), lineated valley fill (Section 2.2.3.2; Figure 2.13B), concentric crater fill (Section 2.2.3.3; Figure 2.13C), and glacier-like forms (Section 2.2.3.4; Figure 2.13D) (Squyres, 1978, 1979; Milliken et al., 2003; Arfstrom and Hartmann, 2005; Souness et al., 2012)—which together store an estimated 0.9–2.7 m GEL of water ice (Levy et al., 2014; Brough et al., 2019). The following sections review VFF subclasses before synthesising their spatial distribution (Section 2.2.3.5), ice content (Section 2.2.3.6), and interconnected evolutionary histories (Section 2.2.3.7).

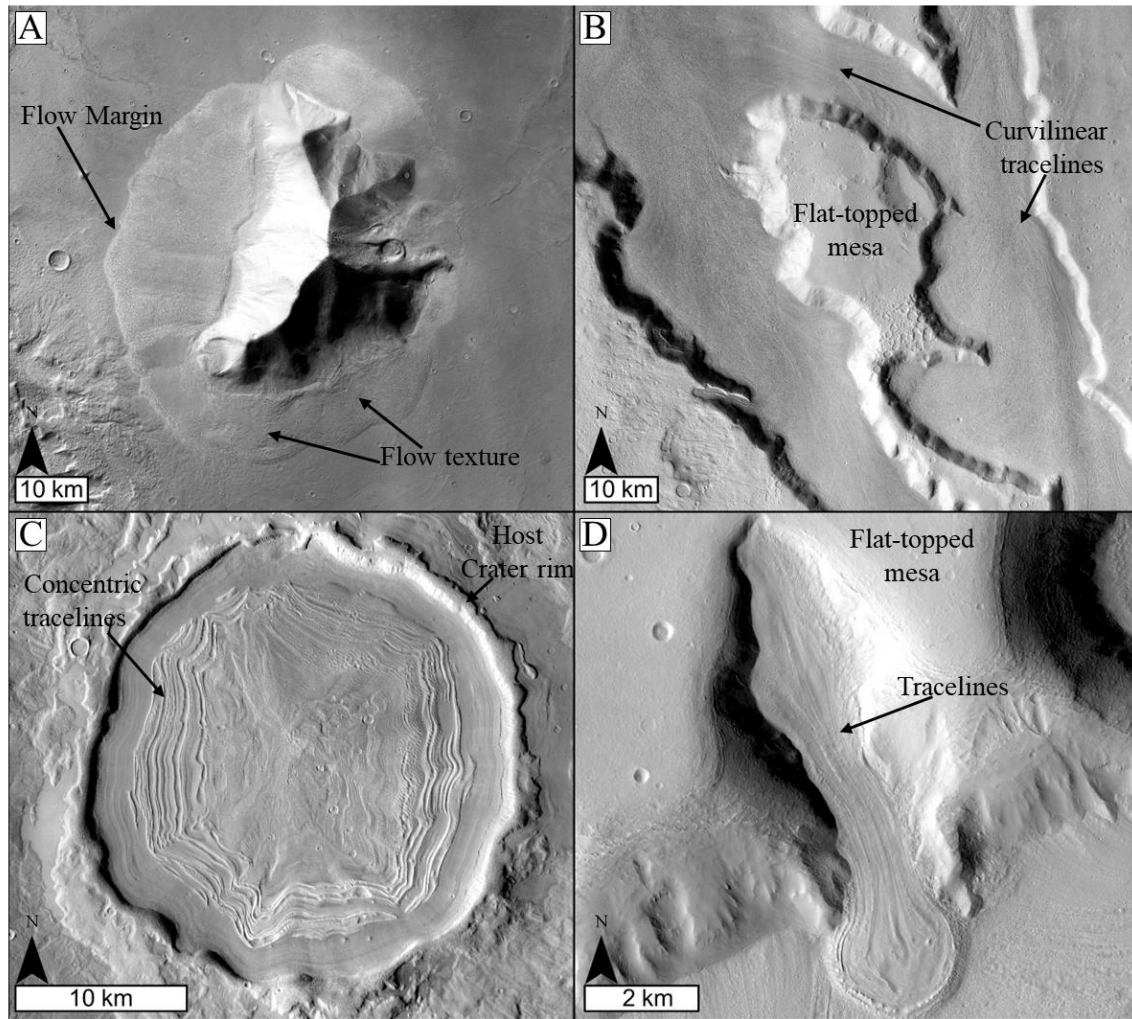


Figure 2.13. Viscous flow features (VFFs) and their principal morphological subtypes. CTX images showing representative examples of the four main VFF classes: (A) Lobate Debris Apron (LDA) in eastern Hellas (40.81° S, 105.88° E; CTX images D13_032266_1393, F13_040877_1391, D10_031066_1393). (B) Lineated Valley Fill (LVF) in northern Arabia Terra (40.24° N, 37.53° E; CTX images F18_042725_2212, F23_044690_2213, P16_007175_2192). (C) Concentric Crater Fill (CCF) in southwest Utopia Planitia (37.89° N, 75.67° E; CTX images P17_007714_2184, B22_018144_2182). (D) Glacier-Like Form (GLF) in Protonilus Mensae (42.23° N, 50.53° E; CTX image G02_018857_2226).

2.2.3.1. Lobate Debris Aprons

Lobate Debris Aprons (LDAs) are wide, apron-like flows that spread from massifs, mesas, crater rims, or escarpments (e.g., Squyres, 1978, 1979; Pierce and Crown, 2003; Mangold, 2003; Levy et al., 2007; Berman et al., 2015), typically terminate at lobate margins (Head et al., 2005), and may form both linear aprons along valley walls and crater interiors and circumferential aprons around isolated mesas, where flow is interpreted to be directed radially outward from massif walls (Head et al., 2010; Figure 2.13A). LDAs cover thousands of square kilometres, extending up to several 10s km over plains and valley floors, and are typically wider than they are long (e.g., Lucchita, 1984; Morgon et al., 2009). The

morphology and surface characteristics of LDAs—including apron geometry, surface tracelines, and convex topographic profiles—imply viscous flow and internal deformation of ice-rich material (e.g., Head et al., 2010). These flow configurations and surface expressions have led to comparisons with terrestrial ice caps or piedmont glaciers (e.g., Mangold, 2003; Milkovich and Head, 2006; Levy et al., 2007; Berman et al., 2015).

2.2.3.2. *Lineated Valley Fill*

Lineated Valley Fills (LVFs) (Figure 2.13B) resemble terrestrial valley glaciers, occupying valley floors with surfaces marked by parallel to curvilinear ridges and troughs formed by ice flow and deformation. These ice-rich deposits extend for hundreds of kilometres and reach widths of several tens of kilometres, with surface tracelines (e.g., Lucchitta, 1984; Squyres, 1978, 1979; Levy et al., 2007; van Gasselt et al., 2010; Baker et al., 2010; Souness and Hubbard, 2012; Wueller et al., 2025). While small-scale alcoves along valley walls can locally feed ice into the system, the primary source regions for the massive volumes of LVF ice remain poorly defined and do not always show clear topographic catchments. Rather than appearing as isolated features, LVFs typically merge into integrated flow networks within trunk valleys, terminating in lobate accumulations or emerging into broader LDA complexes at valley outlets (Head et al., 2006; Levy et al., 2007; Morgan et al., 2009; Wueller et al., 2025). Although their lineations usually parallel the valley axis, more complex lobate or transverse patterns are also observed (Levy et al., 2007).

2.2.3.3. *Concentric Crater Fill*

Concentric Crater Fills (CCFs) (Figure 2.13C) are classified as a crater-interior deposit characterised by repeating concentric rings or loops of surface tracelines, often forming distinctive ridge-and-trough patterns (Squyres, 1978, 1979, 1989; Mangold, 2003; Levy et al., 2009, 2010). These deposits are confined to mid-latitude craters ranging from ~2 to 72 km in diameter (Dickson et al., 2012; Levy et al., 2014), with estimated thicknesses of up to 1.7 km based on modelled crater depth-to-diameter relationships (Levy et al., 2014).

CCF surfaces display a range of features indicative of glacial modification, including lobate flow forms, ring-shaped depressions linked to sublimation, and parallel ridges (Soare et al., 2009; Levy et al., 2010; Pearce et al., 2011; Weitz et al., 2018). Following the descriptive classification of Levy et al. (2010), two morphological subtypes are recognised within the literature: ‘classic’ and ‘low-definition’ CCFs (Figure 2.14). Classic CCFs (Figure 2.14A) are distinguished by well-defined, concentric ridges ~100 m wide, aligned with the crater rim, and a characteristic “brain terrain” surface texture (Levy et al., 2009, 2010). In contrast, Levy et al. (2010) define ‘low-definition’ CCFs (e.g., Figure 2.14B) as crater-filling deposits that lack clearly defined concentric lineations and/or well-developed brain-terrain cells.

Instead, these surfaces may display widened fractures or linear depressions oriented both concentrically and radially to the crater rim, or a muted surface texture characterised by subtle stippling or shallow undulations. Levy et al. (2010) interpret these forms as degraded or mantled expressions of CCFs, in which the original concentric surface morphology has been subdued or obscured, particularly at higher latitudes where latitude-dependent mantle deposits are thicker and more continuous (Head et al., 2003; Levy et al., 2009). These different forms of CCF are elaborated on further in Chapter 3.

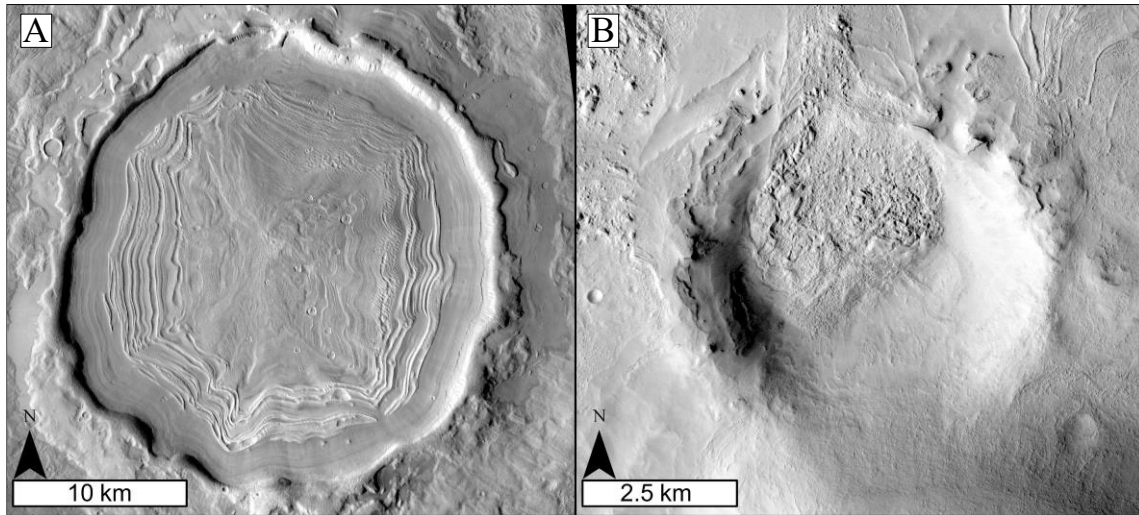


Figure 2.14. Examples of classic and low-definition CCF surface morphologies. (A) Classic CCF in southwest Utopia Planitia (37.89° N, 75.67° E) exhibiting well-developed concentric surface textures, CTX images: P17_007714_2184, B22_018144_2182. (B) Low-definition CCF in the Nilosyrtis Mensae region (33.15° N, 60.39° E), characterised by subdued or poorly expressed concentric tracelines (Levy et al., 2010), CTX image: P16_007385_2149.

2.2.3.4. Glacier-Like Forms

Glacier-like forms (GLFs) (Figure 2.13D) are the smallest subtype of VFFs, resembling terrestrial debris-covered or valley glaciers in planform, originating from confined alcoves on hillslopes unlike LVF (Arfstrom and Hartmann, 2005; Hubbard et al., 2011, 2014; Souness et al., 2012; Brough et al., 2019). Initially termed glacier-like flows (Arfstrom and Hartmann, 2005), they were later renamed as glacier-like forms by Hubbard et al. (2011).

GLFs typically originate in high-relief bedrock alcoves or cirque-like depressions before extending downslope into valleys (e.g., Souness et al., 2012; Li et al., 2024). They often transition from broad upper basins into elongated, narrow tongues confined by raised latero-terminal ridges (Souness et al., 2012; Brough et al., 2016, 2019). These arcuate ridges, bounding their termini, resemble terrestrial moraine ridges composed of debris transported by glacial ice (Milliken et al., 2003; Souness et al., 2012). Morphologically, GLFs share key characteristics with terrestrial valley glaciers, including

surface lineations, crevasses, and moraine-like ridges, all indicative of past or ongoing ice-related deformation (Hubbard et al., 2011, 2014).

2.2.3.5. Spatial distribution of VFFs

VFFs are consistently documented as mid-latitude landforms concentrated within the $\sim 30\text{--}50^\circ$ latitude belts of both hemispheres (Squyres, 1978, 1979; Milliken et al., 2003; Levy et al., 2007, 2014; Dickson et al., 2012; Souness et al., 2012; Figure 2.15). LDAs, LVFs, and GLFs are spatially clustered in the northern hemisphere, concentrated along the steep slopes of the crustal dichotomy boundary, where kilometre-scale relief coincides with the highest mid-latitude slope values (Milliken et al., 2003; Levy et al., 2007; Dickson et al., 2008; Souness et al., 2012; Brough et al., 2019). In contrast, southern hemisphere LDAs and LVFs are more spatially restricted despite widespread steep terrain, a pattern interpreted to reflect the combined influence of slope availability and colder thermal conditions that limit formation and/or preservation to the steepest, pole-facing slopes (Milliken et al., 2003). GLFs display a similar dependence on relief in both hemispheres. Still, southern examples typically initiate partway down crater walls rather than at plateau edges, highlighting hemispheric asymmetries in topographic and thermal controls (Souness et al., 2012). By contrast, CCFs remain near-ubiquitous across the mid-latitudes in both hemispheres (Levy et al., 2014), although this apparent uniformity encompasses both classic and low-definition subtypes. However, unlike LDAs, LVFs, and GLFs—which are restricted to specific slope-bearing landforms—impact craters themselves are spatially ubiquitous across the mid-latitudes. The relevant question is therefore not the availability of host topography per se, but why only a subset of otherwise suitable craters host CCFs, and why preservation state varies so strongly among those that do. This distinction highlights that, for CCFs, distribution cannot be explained by simple topographic opportunity alone and instead requires additional climatic, geometric, or preservation-based controls, a topic explored within this Thesis.

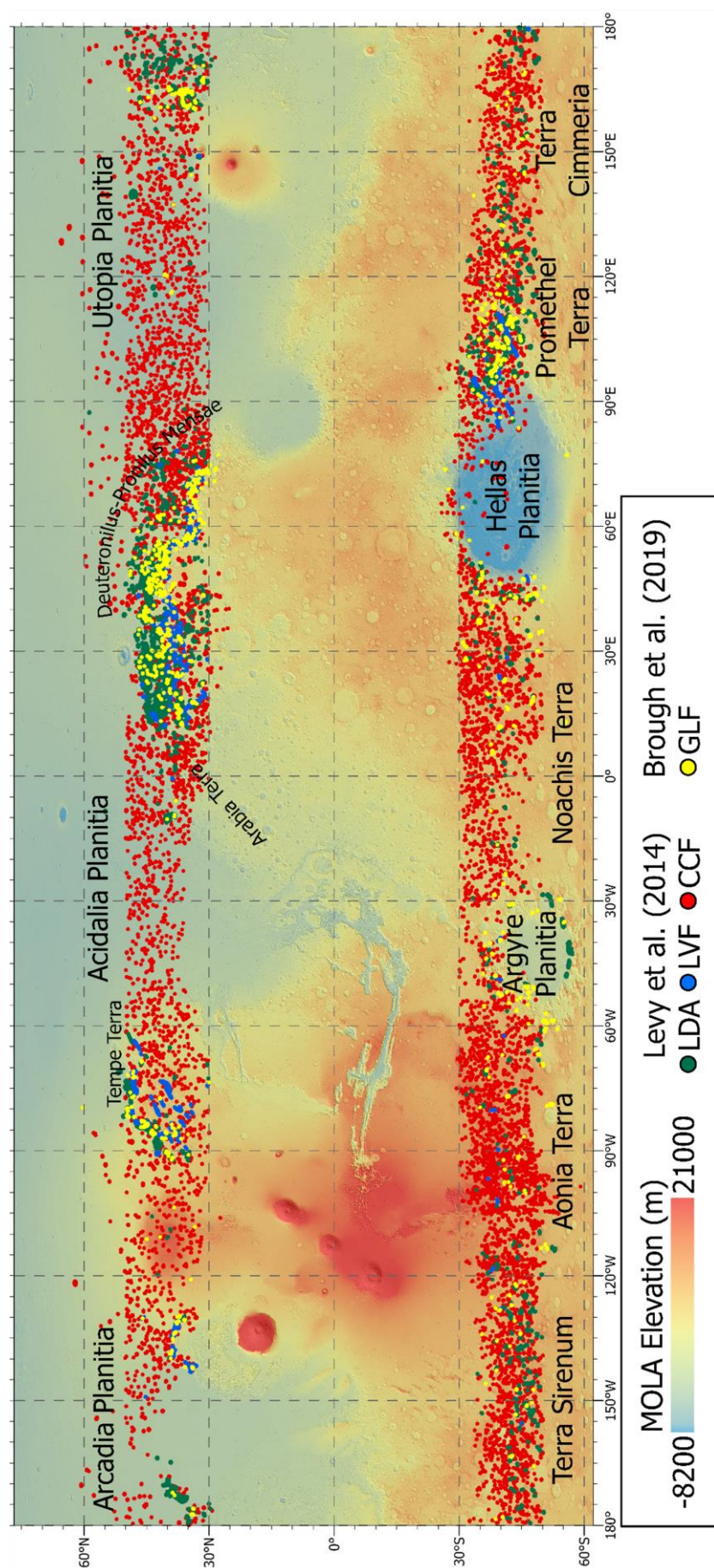


Figure 2.15. Global distribution of Viscous Flow Features overlain on MOLA topography. MOLA elevation map overlain on MOLA hillshade map, showing the mapped distributions of lobate debris aprons (LDAs), lineated valley fill (LVFs), and concentric crater fill (CCFs) are from the global inventory of Levy et al. (2014), while glacier-like forms (GLFs) are from the inventory of Brough et al. (2019).

2.2.3.6. Ice content of VFFs

Direct constraints on VFF internal composition are presently limited to LDAs, for which orbital radar sounding by SHARAD demonstrates a volumetrically dominant, high-purity water-ice interior (>90% ice) overlain by a relatively thin debris cover (Holt et al., 2008; Plaut et al., 2009; Petersen, 2018; Gallagher et al., 2021; Berman et al., 2021; Perry et al. 2024; Figure 2.16). The frequencies of SHARAD and MARSIS are insufficient to resolve the internal structure in most VFFs, such that, only the basal reflector of large LDAs are detectable, precluding direct radar confirmation for smaller forms (e.g., GLFs; Brough et al., 2019) or more complex topographic setting (e.g., CCFs).

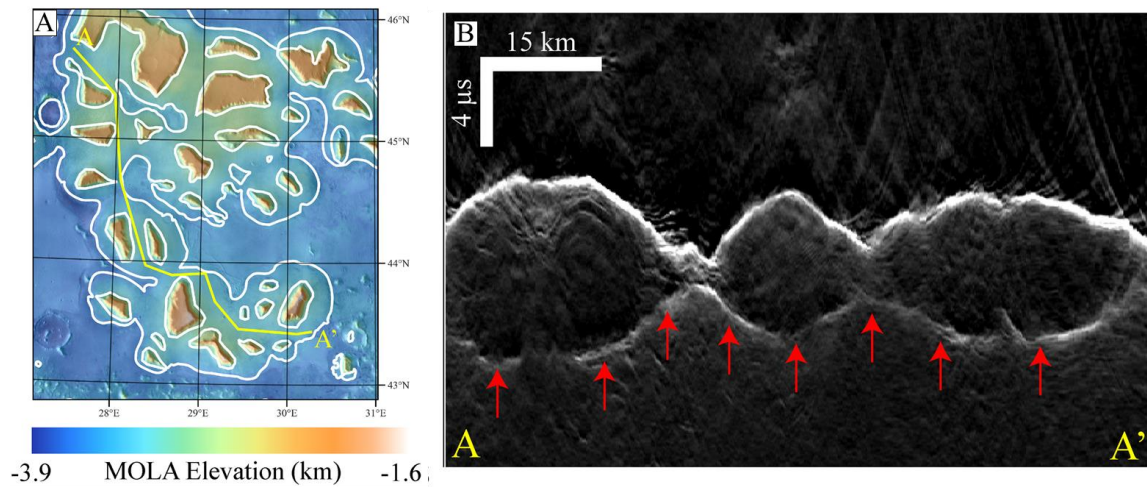


Figure 2.16. Tracing subsurface ice–bed interfaces beneath lobate debris aprons using SHARAD radar sounding. (A) Location of the radar transect (A–A'; yellow line) across a lobate debris apron mapped from the THEMIS daytime infrared global mosaic, with MOLA–HRSC blended topography overlain; the glacial deposit is outlined in white. (B) Corresponding SHARAD radargram extract showing a laterally continuous subsurface reflector (highlighted by red arrows) that can be traced beneath complex surface topography. Redrawn from Perry et al. (2024).

Nevertheless, the strong morphological and contextual continuity between VFF classes supports the inference that they share broadly similar, ice-rich internal compositions, at least at the time of emplacement (e.g., Baker and Carter, 2019; Morgan et al., 2021). Smaller landforms such as GLFs, however, are expected to be more sensitive to volumetric ice loss, and some may plausibly occupy transitional states toward more debris-rich or rock-glacier-like forms in the present day (e.g., Brough et al., 2019). Longitudinal and transverse surface structures occur across all VFF subclasses, including flow-parallel tracelines and curvilinear bands that closely resemble foliation and strain patterns on terrestrial glaciers and are widely interpreted as expressions of ductile flow within ice-rich interiors (e.g., Lucchitta, 1984; Baker et al., 2010; Souness and Hubbard, 2013; Levy et al., 2014). Transverse ridges show similar consistency across VFFs and have been linked to crevassing, moraine formation, rhythmic accumulation–ablation under fluctuating mass balance (e.g., Pierce and Crown, 2003;

Arfstrom and Hartmann, 2005; Parsons et al., 2020; Levy et al., 2021), or the surface expression of internal layering (Butcher et al., 2024; Arnold and Butcher, 2026).

This interpretation of a shared ice-rich internal compositions is further supported by small-scale surface expressions that constrain the shallow structure of VFFs. Sublimation pitting (e.g., Mustard et al., 2001; Soare et al., 2017), ring-mould craters with morphologies diagnostic of impacts into ice-rich substrates (Mangold, 2003; Kress and Head, 2008), and terrains characterised by polygonal or “brain” textures attributed to sublimation of ice beneath a thin debris mantle (e.g., Noe Dobrea et al., 2007; Levy et al., 2009) are all widely interpreted to indicate ice-rich material immediately beneath a debris lag. This inference is directly supported by fresh impacts into “brain” terrain on VFF surfaces that expose bright subsurface ice (Dundas et al., 2021, 2023; Figure 2.11A), demonstrating that near-surface ice occurs in discrete bodies overlain by ice-poor or ice-free debris rather than being distributed primarily as pore-filling ice within an ice-cemented regolith. Radar sounding provides a partial quantitative constraint on the thickness of this debris mantle: for LDAs, SHARAD data indicate that the debris cover is on the order of ~10 m thick, but that its full thickness cannot be directly resolved because the upper interface lies below the vertical resolution of the radar in ice (~10 m) (Holt et al., 2008). The available evidence therefore favours a structure consisting of a high-purity ice interior mantled by a thin, spatially variable, ice-poor debris cover, analogous to terrestrial debris-covered glaciers.

Collectively, radar detections, surface morphologies, and shared glacial histories indicate that VFFs consist of predominantly water-ice interiors mantled by supraglacial debris layers at least several metres thick, analogous to terrestrial debris-covered glaciers (Baker and Carter, 2019), though the data necessary to determine the debris cover thickness and content on VFFs (including CCFs) are not currently available (e.g., Gallagher et al., 2021).

2.2.3.7. The interconnected histories of VFFs

Multiple strands of geomorphic and stratigraphic evidence indicate that VFFs share interconnected glacial histories, with neighbouring ice masses frequently interacting, merging, or recording related patterns of flow and modification (e.g., Head et al., 2006, 2010; Souness et al., 2012). Building on the evidence for their shared, ice-rich composition (Section 2.2.3.6), the critical additional observation is that many VFFs do not occur as isolated bodies but instead commonly interact, merge, or overprint one another, implying coordinated evolution within larger glacial systems rather than discrete, locally confined accumulations.

In several regions, adjacent VFFs form composite flow units, with LVFs, and LDAs frequently converging or coalescing into one another (e.g., Figure 2.17; Head et al., 2006; Souness et al., 2012; Brough et al., 2016). These mergers form integrated glacial systems and thus suggest that VFF subsets did not develop through entirely separate mechanisms, but instead share interconnected histories. In

other cases, GLFs superimpose onto other VFFs, producing cross-cutting flow relationships that have been interpreted to record polyphase glaciation within the same topographic systems (e.g., Hepburn et al., 2020).

This interpretation of composite and interacting VFF systems is consistent with the broader hypothesis that many mid-latitude ice deposits are remnants of a formerly more extensive and laterally continuous ice mass (e.g., Head et al., 2003, 2006, 2010; Dickson et al., 2008, 2010; Levy et al., 2010; Sinha and Murty, 2013; Hubbard et al., 2014), with some authors proposing that regional-scale mid-latitude ice sheets may have developed during peak glaciation phases (e.g., Head et al., 2006, 2010; Levy et al., 2010a; Fastook et al., 2014), with the present VFF populations representing disconnected remnants of this broader cryospheric system.

This interpretation of VFFs as components of integrated glacial systems aligns with proposals that many mid-latitude deposits represent remnants of formerly more extensive and more continuous ice bodies (e.g., Head et al., 2003, 2006, 2010; Dickson et al., 2008, 2010; Levy et al., 2010; Sinha and Murty, 2013; Hubbard et al., 2014). Some authors have further suggested that regional-scale ice sheets or ice-cap systems may have existed during peak mid-latitude glaciation (e.g., Head et al., 2003, 2006, 2010; Levy et al., 2010a; Fastook et al., 2014), with the present VFF populations representing disconnected and differentially preserved remnants of this once broader glacial system.

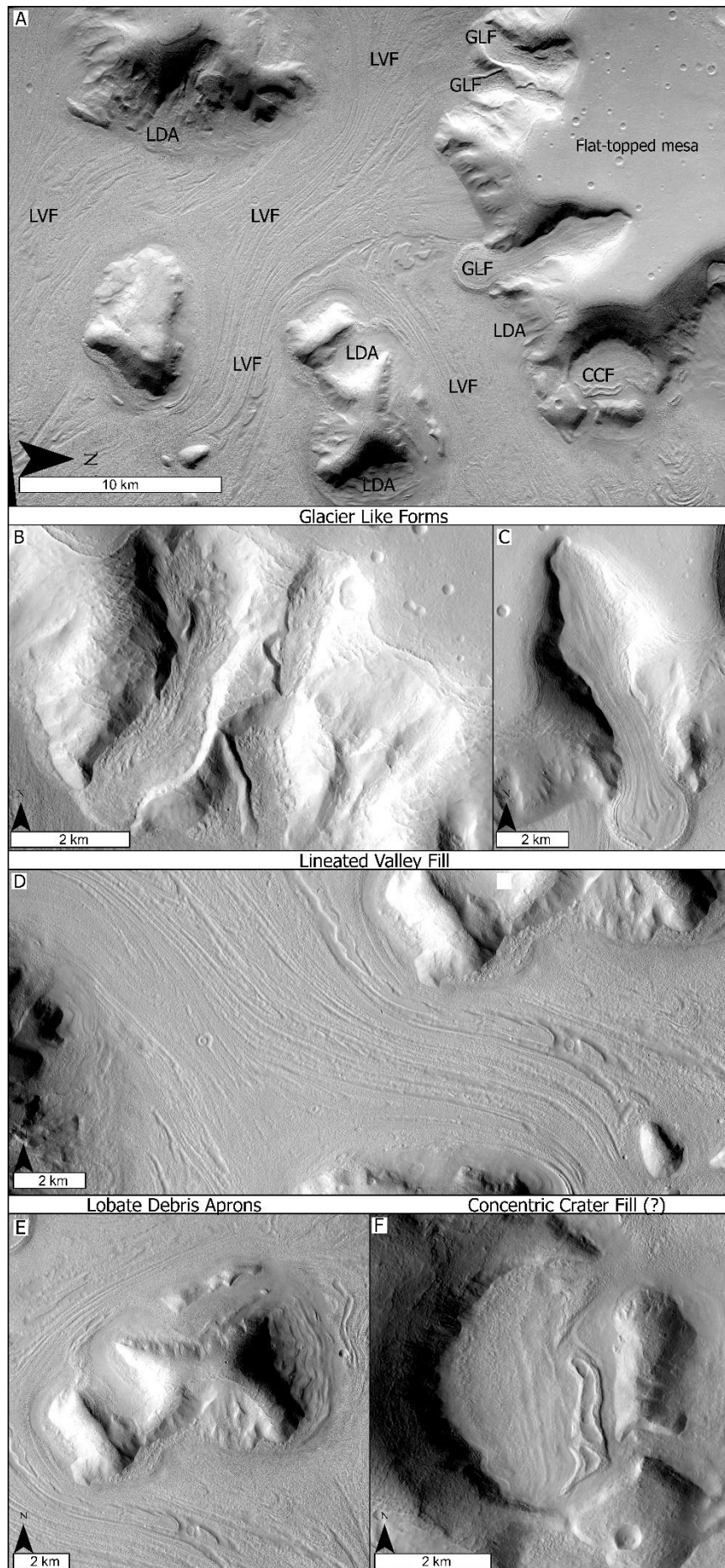


Figure 2.17. Interconnected VFF glacial systems in Deuteronilus Mensae. CTX mosaic (G02_018857_226_XI) showing a VFF-rich region at 42.63° N, 50.52° E in which multiple viscous flow features coalesce and merge into a single, integrated glacial system. (A) Regional context showing the interconnected VFF assemblage. (B–F) Close-up examples of different VFF subtypes within the system. (C) A glacier-like form superimposed onto the broader VFF assemblage, interpreted as evidence for polyphase glaciation (e.g., Hepburn et al., 2020). (F) A crater-filling VFF tentatively classified as a CCF which, despite partial preservation of the crater rim, is morphologically and dynamically merged into the wider system, highlighting the ambiguity in defining when a crater-confined fill constitutes a distinct CCF entity.

2.3. Theory of glacial cycling between the poles, mid-latitudes and equator

2.3.1. Timing and redistribution of Amazonian ice reservoirs

The ages of ice-related landforms on Mars are primarily inferred from stratigraphy and impact crater size–frequency distributions, which constrain surface exposure ages rather than the timing of initial ice emplacement (e.g., Hartmann and Neukum, 2001; Berman et al., 2015). These estimates therefore likely record the most recent modification of the surface layer and may substantially underestimate the emplacement age of ice below. The following sections review the age ranges of the primary ice reservoirs discussed in Sections 2.2.1 and 2.2.2, before situating the multi-episode history of VFFs within this chronological framework.

Crater retention ages for drop-moraine and fan-shaped deposits yield surface exposure ages of 100s Myr, up to ~700 My (Shean et al., 2007; Kadish et al., 2014). These ages are thought to record the age of the surfaces exposed by ice retreat, which initially accumulated and persisted under high-obliquity conditions ($>45^\circ$), when equatorial ice stability was enhanced, and polar reservoirs were unstable (Forget et al., 2006; Fastook et al., 2008). As obliquity declined toward intermediate states ($\sim 35^\circ$), these equatorial reservoirs may have acted as a source, contributing to mid-latitude ice accumulation (Madeleine et al., 2009).

The PLDs preserve a stratified structure of dust and ice layers, long thought to contain climate records (e.g., Murray et al., 1973). For the north PLD (NPLD), there is strong consensus these dust and ice layers record orbitally driven accumulation and that long-term survival and growth of the NPLD is closely linked to the evolution of Mars' mean obliquity (e.g., Murray et al., 1973; Laskar et al., 2002; Putzig et al., 2009; Smith et al., 2016). Before ~5 Ma mean obliquity was 35° (e.g., Laskar et al., 2004; Figure 2.3A), with mean summer insolation at the poles frequently exceeding the threshold for ice stability, favouring NPLD net ice loss and redistribution to lower latitudes (Laskar et al., 2004; Levrard et al., 2007; Fastook et al., 2008). The transition to a lower mean obliquity ($\sim 25^\circ$) ~5 Ma (e.g., Laskar et al., 2004; Figure 2.3A) initiated sustained transfer of water to the polar caps at the expense of mid- and low-latitude reservoirs that had accumulated during earlier high-obliquity states (Levrard et al., 2007). This aligns with radar sounding and stratigraphic analyses indicating that the bulk of the NPLD

accumulated primarily within the past ~4 Myr (Levrard et al., 2007; Smith et al., 2016). Following this transition, it is thought that Mars experienced short-lived excursions to intermediate obliquity ($\leq 35^\circ$) (e.g., Laskar et al., 2004; Figure 2.3A). These excursions reached lower peak values and occurred within a lower mean obliquity state than pre-5 Ma high-obliquity cycles; they are interpreted to have permitted only limited, transient redistribution of ice away from the poles before polar accumulation resumed (Head et al., 2003; Levrard et al., 2007). Obliquity-driven models predict a significant, geologically recent transfer of water from the mid-latitudes to the poles, driven by the transition from an intermediate obliquity regime ($\sim 35^\circ$) to the current lower obliquity state ($\sim 25^\circ$) (Head et al., 2003; Levrard et al., 2007; Bramson et al., 2017). The accumulation of the upper ~300 m of the northern PLD within the past 4 Myr implies a net transfer of ~1 m GEL (~4% of the north polar cap volume) and is thought to be sourced from mass loss of mid-latitude ground ice (Kostama et al., 2006; Smith et al., 2016; Bramson et al., 2017).

In contrast, the stratigraphic record of the South PLD (SPLD) is significantly more complex and exhibits much older surface ages of ~10–100 Myr (Koutnik et al., 2002; Herkenhoff and Plaut, 2007), largely predating the ~5 Ma obliquity transition (Figure 2.3A; Laskar et al., 2004). However, recent spectral analyses of SPLD outcrops have detected stratigraphic periodicities similar to those observed in the NPLD (Becerra et al., 2019). To account for the long-term survival of such ancient ice, researchers have investigated the role of protective lag deposits; earlier models by Levrard et al. (2007) assumed that sublimation continued beneath dust lags of up to 1 m, whereas more recent modelling by Bramson et al. (2019) suggests that as little as a few millimetres of dust may be sufficient to significantly inhibit sublimation and stabilise this reservoir prior to 5 Ma.

The same cycles of ice accumulation, stabilisation, and removal inferred from the PLDs are also thought to underpin the formation of the LDM, which represents one of the youngest ice-rich units on Mars and records this transient ice redistribution particularly clearly. Crater retention ages indicate that most LDM deposits were emplaced within the past ~2 Myr (e.g., Mustard et al., 2001; Schön et al., 2012), with a systematic poleward decrease in age to emplacement ages of <100 kyr near the poles (Kostama et al., 2006; Schön et al., 2012). Modelled crater-based ages support episodic emplacement between ~0.1 and 2 Ma, consistent with repeated, short-lived excursions to intermediate obliquity (30–35°) following the ~5 Ma transition, during which ice was redistributed away from the PLDs, forming the LDM at mid- to high latitudes (Head et al., 2003; Laskar et al., 2004; Levrard et al., 2007). This juxtaposition exposes a fundamental unresolved tension in obliquity-driven ice redistribution models: if the SPLD supplied ice for episodic LDM emplacement between ~0.1 and 2 Ma, how can its surface preserve crater retention ages of ~10–100 Myr?

Mid-latitude VFFs—including LDAs, LVFs, CCFs, and GLFs—exhibit a broad and heterogeneous range of surface crater retention ages spanning much of the Mid-Late Amazonian. Published age

estimates derived from crater size–frequency statistics generally place VFF surfaces between ~ 0.1 and 1 Ga (Levy et al., 2007; Kress and Head, 2009; Morgan et al., 2009; Fassett et al., 2014; Baker and Head, 2015; Berman et al., 2015; Baker and Carter, 2019; Hepburn et al., 2020). Within this range, two recurring age populations are commonly reported for LDAs, LVFs, and CCFs: older surface ages approaching ~ 0.8 –1 Ga (e.g., Levy et al., 2007; Baker and Head, 2015; Berman et al., 2015), and younger ages of several Myr to several hundred Myr (e.g., Neukum et al., 2004; Hauber et al., 2005; Morgan et al., 2009; Baker et al., 2010; Hartmann et al., 2014).

For CCFs, the broad spread in reported surface ages—a commonality across various VFF subtypes—may reflect distinct episodes of ice emplacement or instead may reflect geologically recent resurfacing. Crater dating of brain-terrain surfaces within crater interiors yields surface ages of ~ 10 –100 Ma (Levy et al., 2009), indicating relatively recent modification of the overburden rather than the age of ice emplacement. Generally, crater dating places CCF surface ages as mid-late Amazonian (e.g., Kreslavsky and Head, 2006, 2009; Levy et al., 2010a; Butcher et al., 2025). Importantly, these crater retention ages do not constrain the timing of initial ice emplacement, but instead record the history of debris mantling, erosion, and surface renewal (Baker and Carter, 2019). If palaeo-CCFs were formerly thicker and subsequently experienced substantial mass loss (Section 2.4.3), then their present-day surfaces may record the timing at which they stabilised following mass loss.

GLFs consistently yield younger surface ages than other VFF subclasses, typically ~ 2 –65 Ma (Souness et al., 2012; Hepburn et al., 2020), and some locally superpose older LDAs or LVFs (Hepburn et al., 2020), indicating renewed mid-latitude ice accumulation long after earlier VFF emplacement. Collectively, the wide dispersion of VFF surface ages may therefore reflect polyphase mid-latitude glaciation operating across multiple orbital cycles, and the tendency of crater counts to preferentially date the most recent modification of debris-covered surfaces rather than the underlying ice bodies themselves.

Collectively, these age constraints indicate that individual Amazonian ice reservoirs should not be interpreted in isolation; instead, they should be considered as an interconnected system recording the redistribution of water ice under changing orbital conditions. Importantly, the substantial volumes observed among mid-latitude VFFs imply that such favourable conditions either recurred repeatedly or occurred at different times in different places. This distinction motivates the question addressed in the following section: whether VFFs record single episodes of mid-latitude glaciation or the cumulative imprint of multiple orbitally paced glacial cycles.

2.3.2. VFF formation: single-episode emplacement or multiple glacial cycles?

A central unresolved question in the interpretation of VFFs is whether they record a single episode of mid-latitude glaciation or the cumulative effects of multiple glacial cycles. VFF formation is generally linked to intervals of intermediate obliquity (Section 2.3.1), when ice stability and atmospheric

conditions permit net accumulation at mid-latitudes (Mellon and Jakosky, 1995; Forget et al., 2006; Madeleine et al., 2009). What remains uncertain is whether palaeo-VFF accumulation was confined to one discrete orbital interval or proceeded episodically through repeated phases of accumulation and loss across successive obliquity cycles. Resolving this ambiguity is critical for determining whether present-day VFF morphologies reflect a single emplacement event or the integrated response to long-term climatic forcing.

Clustered boulder bands (Levy et al., 2021) and exposures of inclined internal layering (Butcher et al., 2024) provide critical evidence that VFF interiors may contain fine-scale debris structures, suggesting that these glaciers are not the product of a single, monolithic formation event. While orbital SHARAD radar observations reveal exceptionally pure ice (Section 2.2.3.6; e.g., Holt et al., 2008; Plaut et al., 2009; Petersen et al., 2018), such fine-scale internal complexity remains below current radar detection limits. Thus, radar transparency alone is unable to diagnose single-episode formation, and that some VFFs likely preserve internal stratigraphy produced by multiple phases of growth. While such layering is frequently attributed to high-obliquity cycles, it may also reflect shorter-term climatic signals, including seasonal or annual accumulation cycles. Distinguishing between layers driven by orbital obliquity and those representing higher-frequency cyclicity remains a significant challenge in diagnosing the specific climatic pacing of VFF growth.

The boulder-band clustering observed on LDA surfaces has been interpreted by Levy et al. (2021) as a potential record of repeated phases of ice accumulation and mass loss, rather than the product of a single, continuous glacial episode. In their model, clustered boulder bands represent the surface expression of supraglacial debris layers that accumulate in headwall hollows during intervals of reduced or negative mass balance, become buried during subsequent accumulation phases, and are advected down-flow during renewed glacier advance. Progressive sublimation-driven surface lowering may then re-expose these debris-rich horizons, producing arcuate surface bands and aligned boulder concentrations.

Levy et al. (2021) argue that the number and spacing of these bands are consistent with episodic accumulation operating over long timescales. Specifically, if boulder-band formation is distributed across the inferred ~800 Myr interval of LDA emplacement, the observed number of bands implies accumulation hiatuses on the order of ~10–100 Myr. Variable resurfacing associated with such episodic behaviour could help explain the wide spread of crater-retention ages reported for LDAs (Section 2.3.1), with different surface segments retaining exposure ages that may differ by tens to hundreds of Myr.

An alternative explanation for inclined internal layering has been proposed by Butcher et al. (2024) and Arnold and Butcher (2026), who show that such structures can arise in a geometrically steady-state glacier without any imposed time variation in climate or surface mass balance. In their models, the upwarping of internal layers results from the spatial pattern of strain imposed by longitudinal thickness gradients within the ice body, rather than from changes in bed geometry or from discrete accumulation–

ablation cycles. Importantly, this mechanism affects the entire stratigraphy: all internal layers are progressively folded upward toward the terminus, and some of the material intersecting the surface may originate from deep within the ice, including near the bed. This result demonstrates that the mere presence of inclined internal layers or surface-expressed tracelines does not, by itself, require climate change or repeated glacial episodes. However, this does not suggest that all inclined structures formed in a single glacial interval, nor does it invalidate interpretations of debris-rich bands as the products of episodic accumulation and burial (Levy et al., 2021). The two mechanisms are not mutually exclusive and could operate within the same glacier.

Numerical modelling of VFF formation provides first-order constraints on the timescales required for ice accumulation under plausible Martian conditions, but the resulting estimates are strongly dependent on the assumed climate, rheology, and mass-balance history. Early flow-law calculations suggested that kilometre-scale VFFs require of order $\sim 10\text{--}30$ Myr to develop under conservative assumptions, including low accumulation rates, cold ice temperatures, and relatively coarse grain sizes that imply stiff ice rheology and slow flow (Milliken et al., 2003). Subsequent, feature-specific models demonstrate that substantially shorter formation times are obtained when higher net accumulation rates and/or warmer or softer ice rheologies are prescribed. For example, Weitz et al. (2018) modelled that CCFs can infill craters up to ~ 80 km in diameter within ≤ 7.6 Myr under deposition rates of ~ 1 mm yr $^{-1}$, while Parsons and Holt (2016) modelled the development of several-hundred-metre-thick VFFs at Euripus Mons over ~ 1 Myr using accumulation rates of ~ 5 mm yr $^{-1}$. Similarly, Brough et al. (2019) inferred formation times of order $\sim 10^4$ years for GLFs given mean ice thicknesses of ~ 130 m and accumulation rates of ~ 10 mm yr $^{-1}$.

General circulation models predict that under obliquities of $\sim 35^\circ$ and eccentricities of ~ 0.1 , mid-latitude regions experience sustained net ice accumulation exceeding ~ 6 mm yr $^{-1}$, reaching $\sim 10\text{--}16$ mm yr $^{-1}$ in areas such as Deuteronilus Mensae, with the hemisphere of peak deposition controlled by the longitude of perihelion (Madeleine et al., 2009). These rates substantially reduce formation times relative to early Myr-scale estimates discussed previously, shifting plausible growth times for small-scale LDA, LVFs and CCFs toward kyr–Myr scales. However, even under optimal conditions, predicted formation durations commonly exceed 120 kyr for obliquity, implying that many VFFs likely record ice accumulation integrated across multiple episodes of orbitally modulated climate forcing. Only small-scale features, such as some GLFs, appear capable of forming entirely within one favourable interval (e.g., Brough et al., 2019).

Obliquity controls the latitude of ice stability through variations in insolation and atmospheric moisture transport (e.g., Laskar et al., 2004; Levrard et al., 2007), whereas precession determines which hemisphere preferentially accumulates ice (Madeleine et al., 2009). For single-episode VFFs formation to be viable, substantial ice deposition would need to occur within one ~ 120 kyr obliquity cycle,

favourable to mid-latitude ice stability and within one precession half-cycle of ~ 26 kyr (Levrard et al., 2007), favouring a given hemisphere (Madeleine et al., 2009).

It remains possible, however, that accumulation rates during past glacial episodes exceeded those adopted in current models. The simulations of Madeleine et al. (2009) assume an equatorial source for mid-latitude ice and complete exhaustion of the polar caps. However, they note that if mid-latitude glacial activity was instead sourced—partially or entirely—by mass loss from PLDs, net accumulation rates would differ markedly from those represented in existing GCMs. The PLDs become unstable at obliquities above $\sim 35^\circ$, with predicted polar mass loss of $\sim 26.5\text{--}80$ mm yr $^{-1}$ at obliquities of $35\text{--}40^\circ$ and eccentricity ~ 0.09 (Levrard et al., 2007). If even a fraction of this flux were transferred to mid-latitudes, accumulation rates would significantly exceed those used in most modelling studies. Under such higher accumulation conditions, small-scale VFFs could plausibly form within a single glacial episode. Elevated ice supply from polar reservoirs would therefore broaden the range of viable formation scenarios and weaken the necessity for multi-cycle accumulation in all cases.

Overall, existing observational and modelling evidence does not uniquely support either single-episode or multi-cycle formation for all VFFs. Instead, they point to a spectrum of histories in which some deposits may record repeated accumulation and ablation across multiple orbital cycles. In contrast, others could form rapidly under especially favourable climatic conditions.

2.4. The formation, evolution, and scientific importance of CCFs: state of knowledge

2.4.1. Hypotheses for the formation of CCFs

Since their initial identification in Viking orbiter images (Figure 2.18), a range of early interpretations have been proposed to explain the formation of CCFs. Early interpretations of CCF suggested a weaker association with ice-assisted flow compared to LDAs and LVF (Squyres et al., 1992). Evidence for an ice-related origin included morphological similarities to other flow features, transitional forms between LDAs and CCFs (Squyres, 1989), and a distribution consistent with early ground-ice stability models (Carr and Schaber, 1977). Alternative hypotheses proposed mass wasting of crater walls (Sharp, 1973), aeolian infill (Zimbelman et al., 1989), or limited ice involvement such as talus accumulation stabilised by ground ice (Squyres, 1978), frozen flood water from Amazonian-aged floods (Kreslavsky and Head, 2006), or analogies to terrestrial rock glaciers (Squyres and Carr, 1986; Squyres, 1989; Haeberli, 2006). Voluminous clean-ice glaciers models soon followed, invoking deformation of buried ice masses (Lucchitta, 1984) or viscous relaxation of ice-rich substrates into impact craters (Senft and Stewart, 2008). With higher-resolution imaging from the Mars Orbiter Camera, CCFs were interpreted as debris-covered glaciers emplaced during mid-latitude glaciation alongside LDA and LVF (Milliken et al., 2003) and later GLFs (Souness et al., 2012). It is now widely accepted that CCFs primarily consists of water

ice covered by debris, evidenced by its morphological resemblance to LVF and LDA (Head et al., 2006, 2010; Levy et al., 2007, 2010, 2014; Dickson et al., 2008, 2012; Morgan et al., 2009; Figure 2.13), spatial association with other VFFs (e.g., Squyres, 1978, 1979; Milliken et al., 2003; Levy et al., 2007, 2014; Dickson et al., 2008, 2012; Figure 2.15) LDA ice content confirmed by radar (e.g., Holt et al., 2008; Plaut et al., 2009; Petersen, 2018; Gallagher et al., 2021; Berman et al., 2021; Perry et al. 2024 ; Figure 2.16) and the presence of periglacial features such as brain terrain and ring-mold craters (Mangold, 2003; Kress and Head, 2008; Levy, 2010), numerical modelling of CCF formation (Weitz et al., 2014), and climate models indicating widespread mid-latitude glaciation and VFF formation (Head et al., 2003; Madeleine et al., 2009; Fastook and Head, 2014; Fassett et al., 2014).

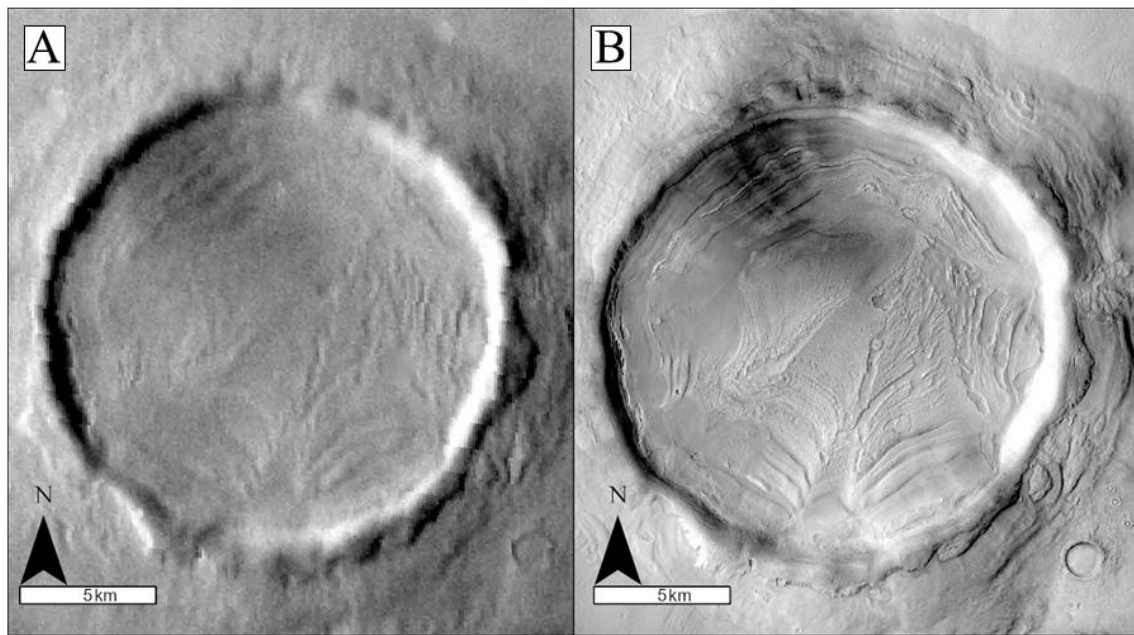


Figure 2.18. Concentric crater fill surface tracelines observed in Viking and CTX images. (A) Viking Orbiter image showing a concentric crater fill (CCF) at 41.51° N, 87.72° E, originally documented by Squyres (1979) (Viking frame 010B70). (B) CTX image of the same CCF (P22_009533_2219_XI_41N272W).

More recent work has converged on a glacial origin for CCFs, interpreting them as debris-covered ice bodies whose mass balance has been modulated by orbital climate variability (e.g., Kreslavsky and Head, 2006; Dickson et al., 2008, 2010, 2012; Levy et al., 2009, 2010a, 2014; Beach and Head, 2013; Fastook and Head, 2014; Weitz et al., 2018). A key unresolved issue concerns the mode of ice emplacement within craters (either on crater floors or on their rims), specifically whether accumulation occurred primarily in situ through atmospheric deposition within individual craters (e.g., Kreslavsky and Head, 2006; Levy et al., 2010a; Fastook and Head, 2014; Weitz et al., 2018), or whether CCFs represent remnants of more extensive regional ice masses that overtopped crater rims and flowed inward (e.g., Head et al., 2010; Levy et al., 2010; Fastook and Head, 2014). Distinguishing between these end-

member mechanisms is central to interpreting whether present-day CCF geometries primarily record local accumulation histories or are the degraded remnants of formerly more extensive ice bodies.

GCMs (e.g., Forget et al., 2006; Madeleine et al., 2009) and ice-sheet flow models (e.g., Fastook et al., 2008, 2010, 2014) have been widely used to argue that LDA and LVF represent remnant fragments of regionally extensive ice sheets, whose extent and persistence were modulated by obliquity-driven climate change (e.g., Head et al., 2003, 2005; Forget et al., 2006; Madeleine et al., 2009; Fastook et al., 2014). This interpretation is supported by multiple lines of evidence, including mapped glacial highstands and regional flow patterns (Dickson et al., 2008, 2010; Head et al., 2010), the presence of distal ice-rich deposits beyond present LDA margins (Baker et al., 2010), the integrated spatial organisation of glacial landsystems (Head et al., 2010), and the consistency between GCM-predicted accumulation patterns and ice-sheet flow simulations (Madeleine et al., 2009; Fastook et al., 2010, 2014). In these models, LDAs and LVF correspond to the debris-mantled, stagnating margins of—perhaps still—retreating regional ice sheets (Dickson et al., 2008, 2010; Fastook et al., 2014). The GCM simulation by Madeleine et al. (2009) will be elaborated on further in Chapter 4.

If such ice sheets contributed to CCF formation, they would need to attain sufficient thickness during intermediate-obliquity intervals to overtop crater rims and flow inward, infilling basins before retreating and leaving isolated ice remnants protected by sublimation lags and/or debris derived from crater walls (Head et al., 2010; Levy et al., 2010a; Fastook and Head, 2014). However, numerical modelling suggests that driving inward ice flow across the mid-latitude plains and into crater interiors would require regional ice thicknesses of several hundred metres (≈ 750 m; Fastook and Head, 2014). This modelled thickness required to initiate flow is significant; when considered alongside the near-ubiquitous distribution of CCFs across the mid-latitudes (Levy et al., 2014), a purely regional-overtopping mechanism would require ice volumes sufficient to blanket the entire mid-latitude belts, implying a hemispheric water GEL far exceeding independent current Amazonian estimates of ~ 20 – 50 m GEL (Carr and Head, 2015, 2019; Alsaedi and Jakosky, 2019; Jakosky and Hallis, 2024; Section 2.2). The volumetric inconsistency suggests that regionally extensive ice sheets cannot have been the dominant or universal mechanism responsible for CCF formation. Nevertheless, geomorphic evidence demonstrates that ice was locally capable of flowing across crater boundaries and between adjacent basins (Figure 2.19; Carr and Head, 2010b), indicating that some CCFs may have received ice from external accumulation zones such as alcoves or massif slopes during favourable obliquity states, without requiring continuous hemispheric ice-sheet cover.

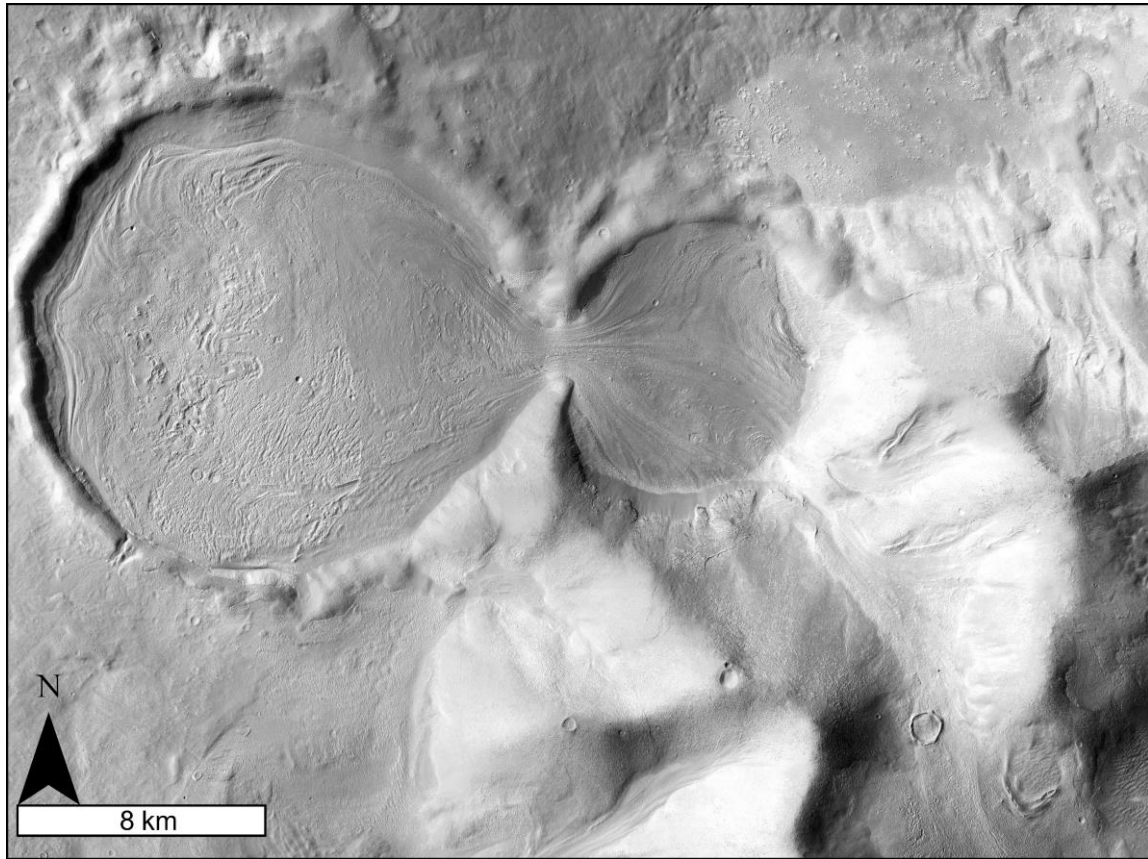


Figure 2.19. Inter-crater ice flow north of Reull Vallis. CTX images (P03_002149_1409; B09_013304_1407) showing ice-derived material flowing from an alcove in the south east and directly to the east, traversing a smaller crater and entering a larger downstream basin north of Reull Vallis (38.57°S, 102.46°E). Redrawn and modified from Carr and Head, 2010b.

An alternative and increasingly supported hypothesis is that CCFs formed primarily through crater-confined accumulation of snow and ice. Early observations noted inward-advancing lobate debris aprons sourced from individual crater-wall sectors that had not yet coalesced across the floor, suggesting progressive infilling driven by continued accumulation (Squyres, 1989). Levy et al. (2010) formalised this concept by proposing that snow and ice accumulated preferentially on steep interior crater walls and within wall alcoves, allowing discrete lobes to grow and merge over time. Fastook and Head (2014) extended this framework using numerical modelling, showing that repeated deposition of thin ice layers, focused onto steep interior slopes and progressively insulated by wall-derived debris, could generate crater-wide debris-covered ice masses. Repetition of this process over multiple orbital cycles produces concentric ridges, flow textures, and partially ablated surfaces consistent with observed CCF morphologies.

Numerical modelling indicates that crater-confined accumulation alone can generate substantial ice volumes over geologically short timescales. Simulations by Weitz et al. (2018) (Figure 2.20) show that repeated deposition of ~50 m thick ice layers during intermediate-obliquity intervals can infill craters <80 km in diameter within several million years, even under modest accumulation rates (~1 mm yr⁻¹).

Higher accumulation rates predicted by GCMs (up to $\sim 10\text{--}16\text{ mm yr}^{-1}$; Madeleine et al., 2009) would further accelerate infill. These models demonstrate that crater-confined glaciation provides a physically plausible mechanism for CCF formation without requiring regionally extensive ice sheets.

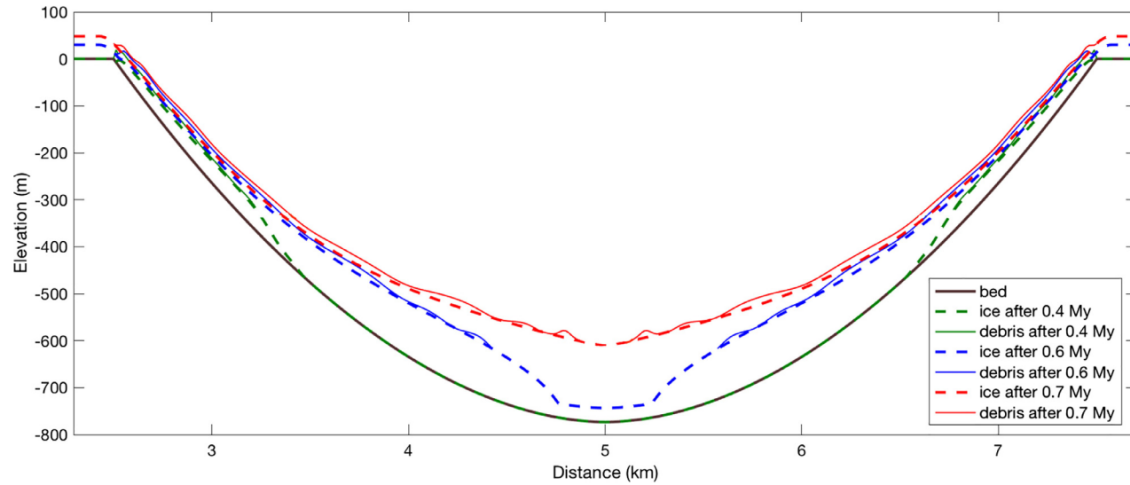


Figure 2.20. Early-stage numerical simulation of the early stages of CCF formation in a 5 km diameter crater. Two-dimensional ice-flow model illustrating the initial stages of crater infilling by ice and debris. Dashed lines show the ice surface and solid lines the debris layer at model times of 0.4, 0.6, and 0.7 Myr. The simulation shows the development of ice-flow lobes from the crater walls (~ 0.4 Myr), progressive inward flow with eventual complete coverage of the crater floor (~ 0.6 Myr), representing the transition from wall-fed lobes to a palaeo-CCF configuration. Redrawn from Weitz et al. (2018).

Limited geomorphic evidence nevertheless suggests that, in some localised settings, ice may have exceeded the confines of craters. Lobate ridges interpreted as moraines on the outer margins of a small number of CCF-hosting craters imply transient outward flow during periods of enhanced accumulation (Figure 2.21; Dickson et al., 2010; Levy et al., 2010). To date, however, such examples are confined to the northern hemisphere, and it therefore remains unclear whether this behaviour was once widespread across the global CCF population or instead reflects a small number of regionally or locally exceptional cases.

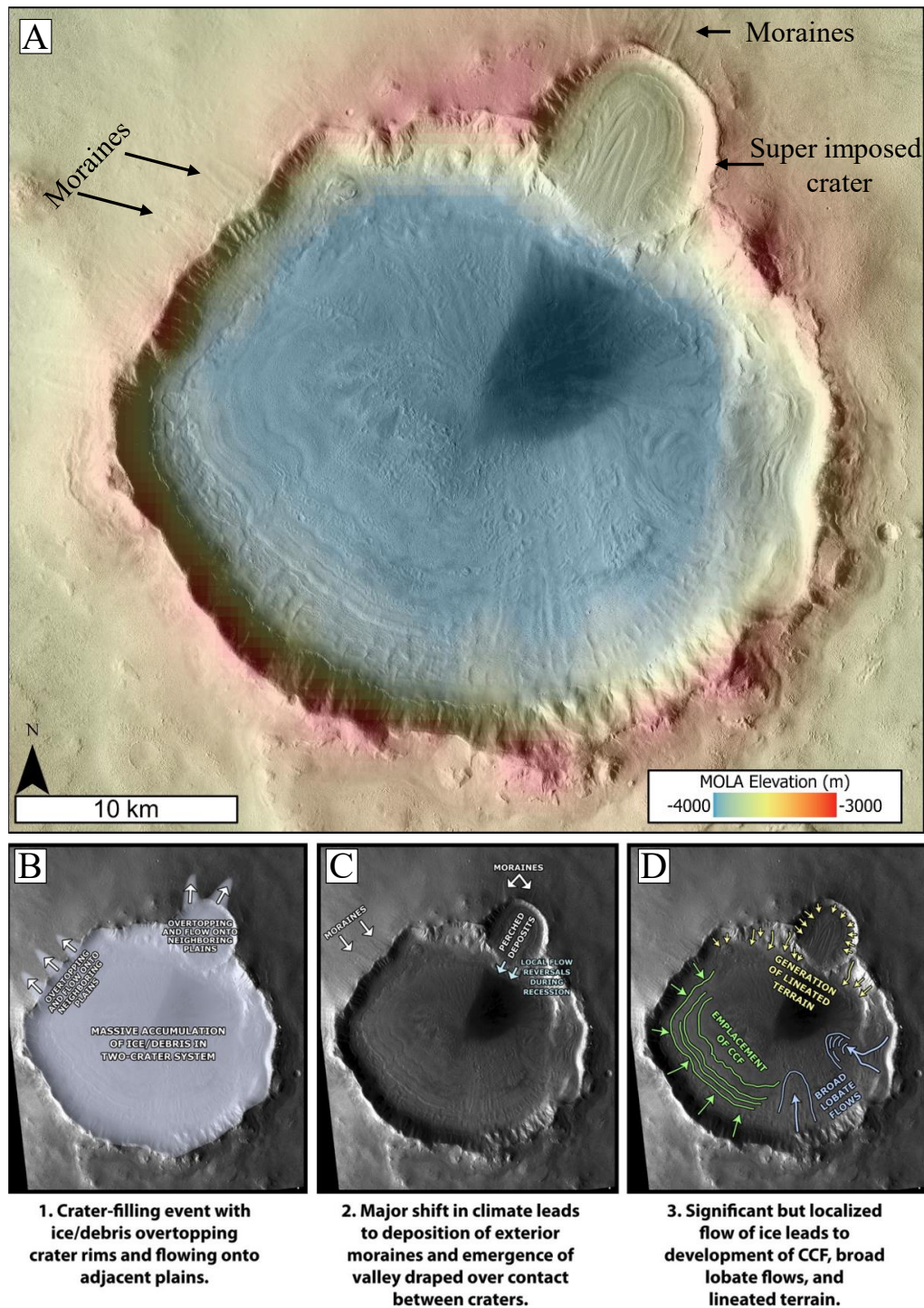


Figure 2.21. Evidence for rim overtopping and subsequent modification in a two-crater CCF system.

(A) CTX image of the two-crater system with MOLA topography overlain, showing the present geomorphic setting in Phlegra Montes (42.91° N, 157.79° E; CTX images: P01_001619_2232, P01_001553_2232). (B–D) Schematic reconstructions from Dickson et al. (2010) illustrating the inferred sequence of ice accumulation, rim overtopping, mass loss, and post-glacial modification within and around the crater system. Panels (B–D) are redrawn and modified from Dickson et al. (2010).

2.4.2. Modification of CCF surfaces

CCFs display significant surface-morphological diversity, ranging from well-preserved examples with sharply defined concentric lineations and diagnostic viscous-flow indicators to more ambiguous forms in which such structures are muted or absent (e.g., Squyres, 1979; Levy et al., 2010). This variability is commonly interpreted to reflect differential degrees of post-emplacement modification acting on a broadly similar primary glacial structure, rather than fundamentally distinct formation mechanisms (Levy et al., 2010; Pedersen and Head, 2010). Existing inventories formalise this continuum through the distinction between ‘classic’ CCFs (Figure 2.14A)—characterised by clearly resolved concentric tracelines—and ‘low-definition’ CCFs (Figure 2.14B), which lack unambiguous lineations and are instead identified through subdued undulations, pitting, fractures, or other ancillary indicators of ice-rich flow (Levy et al., 2010; Head et al., 2010; Levy et al., 2014). While this approach captures a spectrum from well-preserved to heavily modified examples, it also groups deposits with potentially distinct modification histories under a single label. In practice, geologically recent CCF surface modification can be attributed to three dominant modification pathways: periglacial reworking that may alter, but often preserves concentric organisation (e.g., Levy et al., 2010); partial to complete burial by superimposed deposits such as LDM (e.g., Levy et al., 2009; Section 2.2.2.2) or sand dunes; and superposition of younger crater-interior ice lobes that obscure earlier surface textures (e.g., Sinha and Vijayan, 2017). The following sections examine these processes in turn, establishing how surface modification mediates the transition between classic and low-definition CCF morphologies (Figure 2.14).

Periglacial modification overprints older CCF surfaces and records repeated near-surface ice cycling under Late Amazonian climate variability (e.g., Kress and Head, 2008; Levy et al., 2009). A characteristic expression of this modification is termed *brain terrain*, a surface texture composed of arcuate or ovaloid, cell-shaped mounds and ridges typically ~10–20 m wide, ~10–100 m long, and ~4–5 m high, or similarly shaped regions bounded by sinuous ridges (Levy et al., 2009, 2010). On CCFs, brain terrain develops within the debris cover that overlies the ice-rich interior and commonly co-occurs with polygonal ground and sublimation pits (e.g., Levy et al., 2010; Figure 2.22). It is interpreted to form through episodic accumulation and loss of shallow ground ice (e.g., Mangold, 2003; Levy et al., 2009), indicating sustained volatile exchange rather than a single post-glacial modification phase. However, its origin remains debated, with proposed mechanisms including preferential sublimation of ice along steep fracture walls perpendicular to flow direction (Mangold, 2003), stone sorting driven by cyclic freeze–thaw processes (Noe Dobrea et al., 2007), or sublimation-driven topographic inversion of flow-aligned polygons (Levy et al., 2009). CCF surfaces hosting brain terrain are typically sparsely cratered and preserve highly degraded impact morphologies (Malin and Edgett, 2001; Mangold, 2003; Levy et al., 2009), consistent with geologically recent resurfacing driven by repeated episodes of debris emplacement and sublimation-driven modification (e.g., Levy et al., 2010; Morgan et al., 2025). Crater-

retention ages derived from brain-terrain-bearing CCF surfaces define multiple populations, with younger surfaces (~ 25 Ma) overprinting older substrates (~ 80 – 100 Ma; Levy et al., 2010; Morgan et al., 2025), indicating successive episodes of resurfacing, interpreted as being linked to multiple obliquity-driven climate cycles since the Middle Amazonian (e.g., Levy et al., 2010; Morgan et al., 2025). Spatial offsets between brain-terrain ages on CCFs and those measured on adjacent LDAs and LVFs further indicate that resurfacing and near-surface ice stability were not synchronous across all VFF classes (Morgan et al., 2025), reflecting regional climate gradients, variable accumulation histories, and crater-scale microclimatic controls (Levy et al., 2009, 2010; Morgan et al., 2025). Although its surface expression varies, periglacial modification appears to be widespread across the CCF population and, in low-relief or highly degraded examples, may provide the primary evidence for underlying ice (Levy et al., 2010).

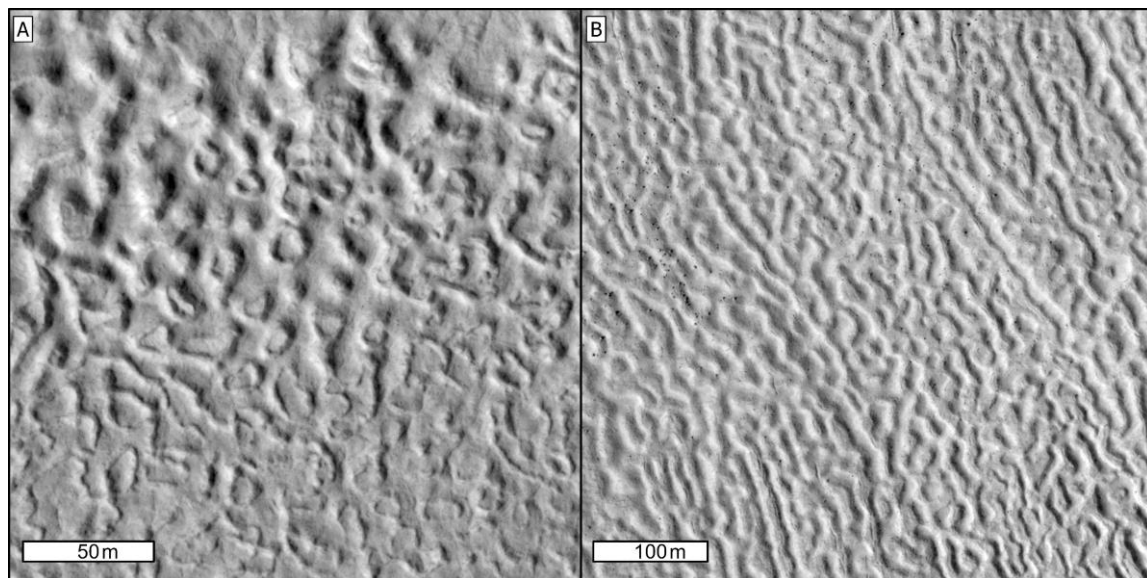


Figure 2.22 Brain terrain on a CCF surface. HiRISE image showing examples of (A) open-cell and (B) closed-cell brain terrain developed on the surface of a CCF, interpreted as the surface expression of sublimation-modified, debris-mantled ice (40.81° N, 99.48° E; HiRISE image: PSP_002175_2210).

LDM emplacement represents a second major late-stage modification pathway affecting CCFs, draping their surfaces during recent obliquity cycles and variably obscuring primary glacial morphology. LDM is interpreted as a thin, ice–dust deposit which drapes the topography covering $\sim 23\%$ of the surface poleward of $\sim 45^{\circ}$ (Mustard et al., 2001; Kreslavsky and Head, 2002; Head et al., 2003; Section 2.2.2) and is common on VFF surfaces (e.g., Head et al., 2003; Hepburn et al., 2019). Curiously, some GLFs appear to be flowing LDM, and for some of these small-scale features it remains unresolved if they are merely mantled by LDM or partially sourced from it (Hepburn et al., 2019). Thick, ice-rich mantles may deform under their own weight, producing morphologies comparable to small VFFs and contributing debris to their surfaces (e.g., Conway and Balme, 2014; Hepburn et al., 2019). On CCFs specifically, LDM expression varies spatially: thin or discontinuous mantling permits windows to the

surface tracelines of CCFs (Figure 2.23A), whereas thicker, cohesive deposits—especially at higher latitudes—completely obscure surface lineations (Figure 2.23B) and generate some of the observed low-definition CCF morphology (e.g., Head et al., 2003; Levy et al., 2009, 2010). This heterogeneity implies a global-scale latitudinal control on mantle accumulation and preservation, likely governed by surface roughness, crater geometry, and the degree of prior ice loss. It highlights the role of repeated late-stage mantling in modifying, rather than erasing, underlying glacial structures.

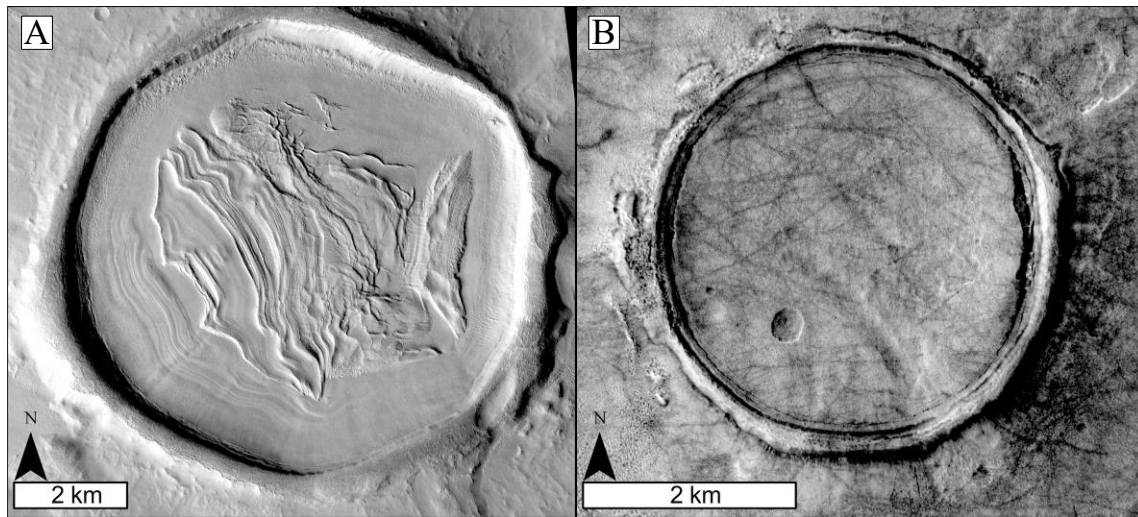


Figure 2.23. Partial and complete LDM mantling of crater infilling surfaces. (A) Example from Acidalia Planitia showing partial LDM cover toward the outer margins of the CCF surface, concentric tracelines remain visible, thus the deposit is still readily identifiable as a CCF (38.58° N, 64.27° E; CTX image: B18_016549_2207). (B) Example from Utopia Planitia showing complete smoothing of the crater-filling deposit by LDM, where any underlying concentric tracelines (if present) are fully obscured (44.81° N, 106.91° E; CTX image: D01_027426_2245).

A further, poorly understood class of crater-interior ice-related landforms is included here because it may record either a late-stage modification of CCFs or a separate, younger phase of crater-interior glaciation whose relationship to CCF evolution remains unresolved. These features take the form of pole-facing crater-interior lobes that originate from the pole-facing crater wall and extend partway across the interior, resembling small, crater-confined LDAs or pole-facing flow features rather than ‘concentric’ crater fills (Levy et al., 2007; Dickson et al., 2008, 2012; Morgan et al., 2009; Sinha and Vijayan, 2017; Figure 2.24). Crater-count data place their emplacement at ~10–100 Ma, consistent with renewed but more limited mid-latitude glaciation long after the main CCF-forming interval (Sinha and Vijayan, 2017). Similar morphologies have also been described for degraded equatorial crater infill deposits (Shean et al., 2010; Figure 2.8). The genetic relationship between these pole-facing lobes and CCFs remains unresolved. They could represent (i) an early or supply-limited stage of crater infill that, given sufficient accumulation rate and/or time, might evolve into a classic CCF; (ii) a heavily degraded remnant of an earlier CCF; or (iii) a younger ice-rich deposit emplaced onto a pre-existing, already

degraded crater fill. Geomorphically, these lobes terminate partway across the crater interior, and the surface beyond their margins is not exposed crater floor (Sinha and Vijayan, 2017), implying the presence of an underlying infill unit. However, it remains unclear whether this substrate is itself ice-rich or instead consists primarily of dust and debris. Regardless of their precise origin, these pole-facing crater-interior lobes occur in many of the same regions that host abundant CCFs (Sinha and Vijayan, 2017), raising an interesting, currently unresolved question: why did some craters maintain stable, CCFs whilst locally adjacent craters—which clearly possess the climate-driven ability to accumulate ice—host small, pole-facing lobes superposed on degraded interior fills? This discrepancy is investigated in Chapter 4.

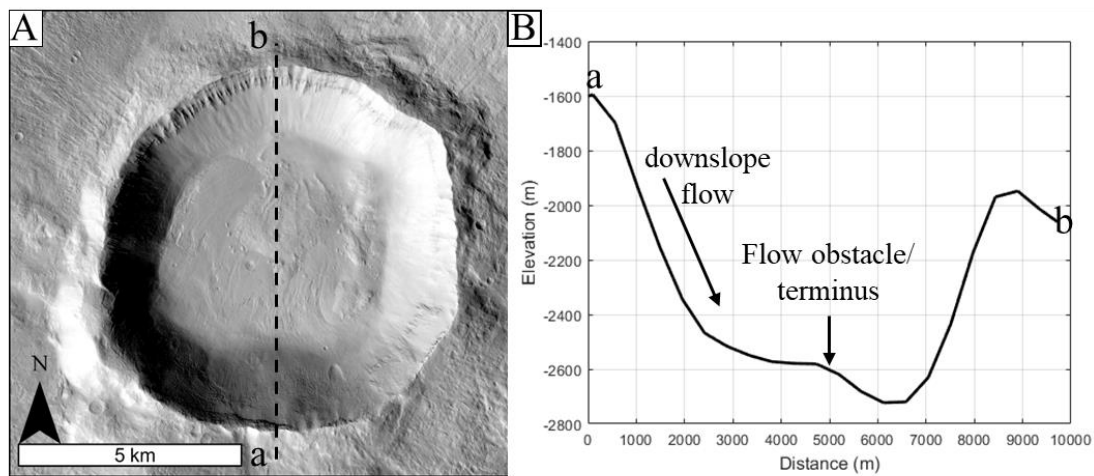


Figure 2.24. Pole-facing flow feature within a crater interior and corresponding topographic transect. (A) CTX image showing a pole-facing flow feature descending from the interior wall of a crater (39.66°N, 135.37° W; CTX image P17_007669_2196). (B) Elevation profile extracted along the labelled transect using MOLA topography. Redrawn and modified from Sinha and Vijayan (2017).

A speculative but physically motivated interpretation is that some degraded CCF record Late Amazonian ablation associated with ice transfer from mid-latitudes to the poles. As discussed in Section 2.3.1, climate models and stratigraphic constraints indicate that the upper ~300 m of the north PLD accumulated within the past ~400 kyr (Head et al., 2003; Levrard et al., 2007). This implies a net transfer of ~1 m GEL from mid-latitude sources, typically attributed to mass loss from mid-latitude ground ice (Kostama et al., 2006; Smith et al., 2016). However, the ultimate fate of mid-latitude ice during low-obliquity intervals remains poorly constrained (Bramson et al., 2017), and there is no requirement that ground ice was the exclusive source of this transfer. VFFs constitute a volumetrically significant mid-latitude reservoir (presently ~3 m GEL; Levy et al., 2014; Brough et al., 2019), such that even a partial contribution to this flux would imply substantial downwasting of debris-covered glaciers, including CCFs. Whilst evidence for large-magnitude mass loss is observed across the wider VFF population (e.g., Levy et al., 2007, 2010; Dickson et al., 2008, 2010; Morgan et al., 2009; Pedersen and Head, 2010; Brough et al., 2016), the spatial extent and magnitude remain unknown. For CCFs, direct geomorphic

evidence for much greater former ice thicknesses is rare, with only a small number of examples preserving indications of rim overtopping or markedly greater extents (Dickson et al., 2010; Levy et al., 2010; Section 2.4.1). This scarcity, however, does not preclude substantial former mass loss, because significant thinning could occur without generating or preserving clear external markers of greater extent, particularly if the ice was cold-based and performed limited geomorphic work on its substrate, and/or if any such markers were later obscured by the superposed deposits discussed above. In this framework, some ‘low-definition’ CCFs—namely those that appear morphologically degraded and deflated rather than merely subdued by latitude-dependent mantle cover (e.g., Figure 2.14B; Levy et al., 2010)—may represent systems that failed to reach long-term mass-balance equilibrium beneath a protective overburden and instead underwent prolonged mass loss to their present states, potentially during low-obliquity intervals ($<15^\circ$) that recurred repeatedly over the last ~ 5 Myr (Laskar et al., 2004). By contrast, ‘classic’ CCFs may represent systems that developed sufficiently thick debris insulation to stabilise their ice bodies early, or that have since experienced only extremely low net ablation rates.

Collectively, these surface modification pathways, and the current extent of potential mass loss demonstrate that present-day CCF surface morphology reflects a continuum of preservation states rather than discrete endmembers. Periglacial reworking, variable mantling by LDM, and superposition of younger crater-interior ice lobes can each modify, obscure, or selectively preserve primary concentric structures to differing degrees, producing the observed spectrum from sharply defined ‘classic’ CCFs to more ambiguous, low-definition forms. Despite the widespread use of this terminology, a consistent, morphology-based classification of CCF preservation states has not yet been systematically applied, and surface variability has rarely been characterised beyond qualitative descriptions. Addressing this gap is essential to isolating well-preserved examples that retain diagnostic information about CCF formation and evolution. Accordingly, Chapter 3 develops a structured classification framework for CCFs, distinguishing high-confidence, uncertain, and non-CCF deposits and documenting the range of surface morphologies within the classic CCF population.

2.4.3. Connecting CCF evolution to orbital forcing

Here I synthesise current understanding of CCF evolution within the broader context of Amazonian mid-latitude glaciation. The discussion is structured around three broad phases commonly invoked in previous work—glacial advance, glacial retreat, and geologically recent surface modification or mantling—and summarises how each phase is likely linked to changes in obliquity-driven climate conditions. Rather than proposing a new evolutionary model, the aim is to consolidate existing interpretations of how CCFs reflect cumulative ice accumulation and ablation during the Mid-Late Amazonian (Figure 2.25).

In Figure 2.25, I present a schematic framework for CCF evolution that organises the principal formation and modification pathways proposed in the literature into a single, internally consistent

lifecycle. This framework spans three primary phases: glacial accumulation, glacial retreat, and post-glacial modification. I link alternative accumulation scenarios and potential peak states—some exceeding the present extent—to subsequent net ablation pathways that generate the modern spectrum of CCF morphologies. In the following sections, I treat these phases in turn, with reference to the left, middle, and right-hand panels of the figure to track the progressive evolution of the CCF through time.

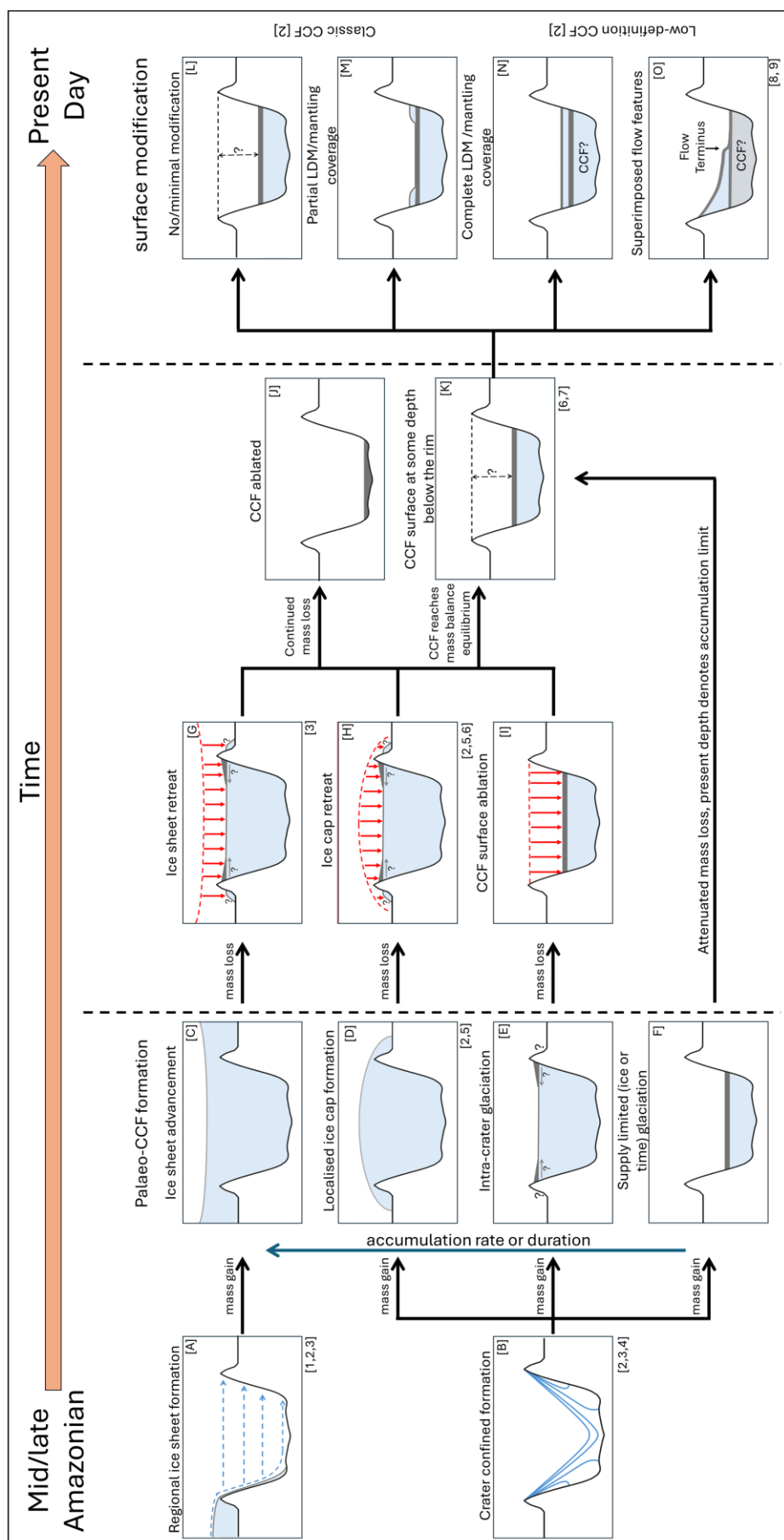


Figure 2.25. Schematic of CCF evolution. Conceptual diagram summarising the sequence of processes proposed in the literature for the evolution of concentric crater fillings, from initial ice accumulation, through phases of growth and subsequent ablation, to progressive surface modification and degradation leading to the present-day classic and low-definition morphology. Numbered stages correspond to published conceptual and modelling frameworks: [1] Head et al. (2010); [2] Levy et al. (2010); [3] Fastook and Head (2014); [4] Weitz et al. (2018); [5] Dickson et al. (2010); [6] Levy et al. (2014); [7] Pedersen and Head (2010); [8] Dickson et al. (2012); [9] Sinha and Vijayan (2017).

Glacial Advance

During intervals of intermediate obliquity ($\sim 35^\circ$), GCMs predict enhanced mid-latitude net ice accumulation, creating conditions favourable for the formation of VFFs (Madeleine et al., 2009). Under such regimes, palaeo-CCFs have been proposed to form either through inflow from regionally extensive ice masses capable of overtopping crater rims (Head et al., 2010; Levy et al., 2010; Fastook and Head, 2014; Figure 2.25A) or through crater-confined accumulation within individual basins (Levy et al., 2010a; Fastook and Head, 2014; Weitz et al., 2018; Figure 2.25B). Direct constraints on the timing of CCF emplacement remain limited because many surfaces have been modified or mantled (Mangold, 2003; Kostama et al., 2006; Levy et al., 2009; Section 2.4.2), yet crater-based surface ages place CCF surfaces within the mid to Late Amazonian ($\sim 0.1\text{--}1$ Ga; Levy et al., 2010; Kress and Head, 2009; Fassett et al., 2014; Butcher et al., 2025), overlapping with surface ages of other VFFs (Levy et al., 2007; Kress and Head, 2009; Morgan et al., 2009; Fassett et al., 2014; Baker and Head, 2015; Baker and Carter, 2019; Hepburn et al., 2020).

Orbital variability provides a mechanism for repeated alternation between accumulation-favourable and ablation-favourable conditions at mid-latitudes over Amazonian timescales. Mars' obliquity and precession undergo quasi-periodic oscillations (~ 120 kyr obliquity cycles; ~ 51 kyr precession; Levrard et al., 2007), which modulate insolation and hence the stability of snow and ice at mid-latitudes (Laskar et al., 2004; Levrard et al., 2007). GCMs indicate that specific orbital configurations— 35° obliquity, 0.1 eccentricity, and perihelion aligned with northern and southern summers ($L_p \approx 90^\circ/270^\circ$)—can yield sustained mid-latitude net accumulation with local rates approaching $\sim 10\text{--}16$ mm yr $^{-1}$ (Madeleine et al., 2009). These configurations display hemispheric asymmetry, such that precession determines which hemisphere experiences enhanced accumulation at a given time, while the opposite hemisphere remains unfavourable for ice accumulation (Madeleine et al., 2009). Under a crater-confined formation pathway, this orbital variability implies incremental growth through repeated depositional episodes rather than continuous accumulation. Ice would preferentially accumulate during intervals of favourable orbital geometry. At the same time, intervening configurations would limit further growth and promote partial surface lowering, with overburden progressively accumulating to suppress sublimation and enhance long-term preservation (Schorghofer and Aharonson, 2005; Fastook and Head, 2014; Weitz et al., 2018).

Although all CCFs are presently crater-confined, modelled accumulation rates and geomorphic evidence suggest that some palaeo-CCFs achieved substantially greater thicknesses and extents (Figure 2.25E)—potentially overtopping their rim (Figure 2.25D). Outward-facing lobate ridges interpreted as rim-overtopping moraines at the external margins of a small number of CCF-hosting craters imply rim overtopping and transient outward flow during peak accumulation at those sites (Dickson et al., 2010; Levy et al., 2010). Such examples support the possibility that localised ice caps or domes developed

under particularly favourable accumulation conditions (Figure 2.25D). The near-ubiquitous present-day distribution of CCFs across the mid-latitudes (Levy et al., 2014) raises the possibility that the current population represents only a remnant of a once more extensive ice-rich crater inventory. In this scenario, additional palaeo-CCFs either failed to develop sufficient debris insulation or were removed during subsequent phases of net ice loss. It is also plausible that CCFs, either locally or globally, experienced some form of supply limited (either time or ice accumulation) glaciation and never ascertained rimful conditions (Figure 2.25 E, F).

Glacial Retreat

Following intervals of mid-latitude net accumulation, subsequent changes in orbital configuration are expected to have shifted the mass-balance regime toward net ablation, promoting glacier thinning and retreat (Figure 2.25). Although orbital solutions can only be resolved robustly for the past ~20 Myr (Laskar et al., 2004), large-amplitude obliquity variations are widely inferred to have repeatedly redistributed volatiles between the poles, equator, and mid-latitudes throughout the Amazonian (Laskar et al., 2004; Schorghofer and Aharonson, 2005; Chamberlain and Boynton, 2007; Levrard et al., 2007). During such transitions, mid-latitude surface ice would no longer experience sustained accumulation and would instead undergo progressive thinning as sublimation exceeded deposition.

In addition to orbital forcing, limitations on ice supply may have contributed to the stagnation and thinning of CCFs. GCMs indicate that mid-latitude ice accumulation is ultimately sustained by water vapour sourced from equatorial reservoirs established during high-obliquity states (Madeleine et al., 2009; Forget et al., 2013). Depletion or reduction of these finite-source reservoirs would therefore curtail further accumulation, even in the absence of a major orbital transition, potentially reinforcing retreat once accumulation ceases.

Irrespective of the dominant driver, glacial retreat would promote progressive CCF ablation until a state of mass-balance equilibrium beneath a protective overburden layer. This overburden is interpreted to develop through a combination of atmospherically deposited dust, debris supplied from crater walls, and/or sublimation lag generated during ice loss, which together suppress further ablation and enable long-term preservation of subsurface ice (Schorghofer and Aharonson, 2005; Fastook and Head, 2014; Weitz et al., 2018). Under such conditions, ablation rates may become extremely low or effectively arrested, allowing CCFs to persist through extended intervals of unfavourable climate despite ongoing surface modification.

Under such conditions, any paleo-CCFs that overtopped their host-crater rims would be expected to undergo progressive ablation until retreating back into their host craters (Figure 2.25G-I), resulting in either complete ablation (Figure 2.25J) or eventually approaching mass-balance equilibrium beneath an insulating debris cover (Figure 2.25K). Direct geomorphic evidence for substantial post-glacial retreat is preserved in a limited number of CCFs, where steep talus aprons, truncated margins, and interior-

wall retreat indicate losses of at least tens to hundreds of metres of ice (Levy et al., 2010; Figure 2.25H-I), although the extent to which similar degrees of mass loss characterised the broader CCF population remains uncertain. Comparable evidence for large-magnitude mass loss is observed across the broader VFF population, where linear ridges high along valley walls have been interpreted as lateral moraines, implying hundreds of metres of ice-surface lowering since emplacement (Morgan et al., 2009; Levy et al., 2016).

For CCFs, if post-glacial mass loss was spatially heterogeneous, this may explain the observed variability in present-day CCF surface elevations relative to their crater rims (Pedersen and Head, 2010; Levy et al., 2014; Figure 2.25K). Such variability may reflect differences in local climatic conditions during interglacial periods, initial ice volumes, spatial variations in debris-cover development, and/or the modulating influence of host-crater geometry on CCF preservation. Spatially heterogeneous ablation may also account for the occurrence of low-definition CCFs, which may represent end-member cases in which insufficient debris insulation developed to stabilise the palaeo-CCF, leading to strongly degraded or near-vanished deposits (Figure 2.25J). A potential extreme end-member of this degradation pathway may be represented by the equatorial crater-floor deposits described by Shean (2010; Section 2.2.1.3), which may preserve geomorphic evidence for former ice-rich infill but for which the survival of any relict ice remains uncertain. If these equatorial craters are indeed the remnants of former ice infill, they may represent the terminal products of equatorial ‘palaeo-CCF’s that failed to develop or maintain a sufficiently protective debris cover during post glacial periods, and instead progressed toward near-complete loss of their original ice content. Perhaps these equatorial features represent the long-term fate of all mid-latitude CCFs, which, given sufficient time, may progressively degrade into low-definition forms and ultimately into near-vanished end-members, but it remains unclear whether some “classic” CCFs have already entered a regime of reduced climatic sensitivity, in which a sufficiently thick debris overburden suppresses further ablation and allows subsurface ice to maintain mass-balance equilibrium through successive future orbital and post-glacial climate states. The variability in CCF surface depth observed by Pedersen and Head. (2010) and Levy et al. (2014) is explored further in Chapter 5.

Any theory of the evolution of CCFs would involve at least one build-up phase and at least one retreat phase (unless CFFs follow panel F). The crudest working hypothesis would summarise the evolution as overall advance/growth, followed by overall retreat, hence the model in Figure. 2.25. But the true evolution is likely very complex and remains uncertain. Current understanding of CCF (and other VFF) evolution does not resolve whether palaeo-CCFs formed during a single interval of sustained accumulation or through multiple episodes of ice deposition and ablation operating across successive orbital cycles. In a single-episode scenario, CCFs would represent the product of one prolonged period of favourable climatic conditions, producing a coherent, high-purity ice body. Alternatively, a multi-cycle history would imply that CCFs record the cumulative effects of repeated accumulation and partial loss, with internal structure reflecting the superposition of multiple glacial phases and surface debris

layers which become internal with subsequent burial layers. Importantly, both pathways could plausibly generate palaeo-CCFs that were formerly thicker and more extensive than observed today (Figure 2.25C-E), provided that accumulation was sufficiently large or repeated over enough cycles. Subsequent ablation, partially attenuated through armouring by the overburden, would then reduce these palaeo-CCFs to their present configurations, making it difficult to distinguish between single- and multi-cycle origins based on present morphology alone.

CCF Surface Modification

Geologically recent surface modification establishes a continuum of post-emplacement overprinting on the CCF population (Section 2.4.2; Figure 2.25L–O), ranging from subtle periglacial texturing (Figure 2.25L, M) to superposed units/mantling that fully obscure the infilling unit beneath (Figure 2.25N, O). This spectrum reflects progressive climatic forcing during the Late Amazonian and records repeated episodes of ice redistribution, degradation, and renewed emplacement. Within this continuum, some CCFs—termed ‘classic’ (Levy et al., 2010; Figure 2.25L, M)—retain well-preserved, high-confidence morphologies that likely preserve signals of their primary accumulation and ablation history, whereas others—termed ‘low-definition’ (Levy et al., 2010; Figure 2.25N, O)—include heavily degraded features, in addition to crater fills which possess surfaces shaped by successive phases of mantling, periglacial modification, and younger ice emplacement. These latter cases record composite surface histories in which multiple climatic episodes are superimposed, complicating direct inference of original ice geometry or volume. Distinguishing between these endmembers is therefore essential for reconstructing Mid- to Late-Amazonian glacial behaviour and for assessing whether present-day CCF morphometry reflects peak ice accumulation states or the cumulative effects of post-glacial modification. Accordingly, isolating the best-preserved CCFs provides the most robust basis for evaluating the climatic, geometric, and volumetric controls on crater-confined glaciation, and for distinguishing primary accumulation signatures from later overprinting.

2.4.4. Scientific importance of CCFs

Water ice is a fundamental component of the Martian system, exerting a first-order control on surface evolution and providing a principal constraint on the planet’s climatic and habitability history. As the most widespread class of mid-latitude debris covered glaciers, CCFs potentially constitute an extensive archive for assessing the Martian H₂O inventory and the long-term stability of the cryosphere over million-year timescales.

Are CCFs a significant contributor to the current Martian H₂O inventory?

Constraining the modern water budget is essential for understanding Mars’ climatic evolution. Current syntheses place the accessible H₂O inventory at approximately ~20–50 m GEL, stored primarily within the polar layered deposits and widespread subsurface ice reservoirs (Carr and Head, 2015, 2019;

Alsaeed and Jakosky, 2019; Jakosky and Hallis, 2024). Within this global budget, mid-latitude viscous flow features are estimated to represent a non-negligible component, contributing of order $\sim 0.9\text{--}2.7$ m GEL in total (Levy et al., 2014; Brough et al., 2019). Recomputing this budget for CCFs alone using the Levy et al. (2014) CCF inventory yields a total ice volume of $\sim 8.75 \times 10^4$ km³, equivalent to ~ 0.6 m GEL. Under these existing estimates, CCFs would therefore account for only a few percent ($\sim 1\text{--}3\%$) of the total present-day Martian water inventory. Although CCFs do not dominate the global GEL inventory, these values suggest that they nevertheless constitute a volumetrically significant secondary component of the mid-latitude cryosphere, with a cumulative volume comparable to that inferred for LVFs ($\sim 6.5 \times 10^4$ km³; Levy et al., 2014). More importantly, their scientific importance may lie less in their fractional contribution to the total H₂O inventory and more in their near-ubiquitous spatial distribution and geometric confinement within craters, which make them unusually well-suited for systematic, population-scale analysis of mid-latitude ice accumulation and preservation. It should be emphasised, however, the CCF volume estimate of Levy et al. (2014) include both low-definition and classic CCFs and, as will be shown in Chapter 6, their current global volume estimates for CCFs require revision.

CCFs as a spatially ubiquitous climate record (record of environmental gradients)?

The near-ubiquitous distribution of CCFs across the Martian mid-latitudes (Levy et al., 2014) makes them uniquely suited for assessing how large-scale environmental gradients influenced ice accumulation and preservation during the Amazonian. In contrast, non-CCF VFFs (LDAs, LVFs, and GLFs) show pronounced spatial clustering tied to regional relief, with strong preferences for steep slopes and high topographic gradients in either hemisphere (e.g., Milliken et al., 2003; Souness et al., 2012). Though it remains unresolved whether the present spatial distribution of VFFs other than CCFs reflects conditions necessary for their initial accumulation or subsequent preservation during post-glacial modification (Souness et al., 2012). CCFs differ fundamentally in this respect: because they are confined within impact craters, their formation does not require regionally extensive slopes, and suitable host geometries provided by impact craters are widespread across the mid-latitudes. Consequently, spatial variations in CCF distribution and morphology may reflect environmental modulation—such as latitudinal or elevational controls on their accumulation and/or ablation—rather than the distribution of suitable topography alone, although variation in background crater density might also influence the CCF distribution.

This near-ubiquity therefore provides a unique opportunity to use CCFs as a population-scale test of how environmental gradients modulate ice accumulation, modification, and preservation across the mid-latitudes—an opportunity that is taken up explicitly in this Thesis.

CCF as a potential target for future exploration (life, ice cores, In-site resource utilisation)

VFFs, including CCFs, represent potential targets for future exploration by orbital and surface missions. Although the probability of present-day habitability within CCFs is expected to be low, it is not zero. Terrestrial studies demonstrate microbial persistence for 10^5 – 10^7 years within glacial ice and permafrost (e.g., Christner et al., 2014), indicating that long-term preservation of biosignatures within Mid-Late Amazonian, debris-covered Martian ice is at least plausible (Lavagna et al., 2022). Importantly, CCF stratigraphy seems likely to preserve a layered archive of Amazonian climate. Perhaps then VFFs represent potential long-term cryogenic archives rather than contemporary habitable environments?

More broadly, the primary scientific value of CCFs lies in their potential to preserve a time-integrated record of Amazonian climate. Whether individual CCFs formed during a single prolonged glacial interval or through multiple phases of accumulation and retreat, glacier ice inherently traps dust, volatiles, and atmospheric gases during snowfall and firnification, which may be retained as internal stratigraphy. In this context, recovery of ice cores from ‘classic’ CCFs—via deep drilling missions (e.g., Niles et al., 2023)—may offer long-term records of atmospheric composition and orbitally forced climate variability, because CCFs (compared to ‘low-definition’ CCFs) show the least evidence for post-depositional mantling and reworking. However, the strong evidence for flow convergence, compression, and internal deformation in crater-confined systems implies that such stratigraphy is likely to be significantly warped and reoriented, complicating any simple “ice-core” style interpretation. Accessing CCF ice requires drilling through an overburden of uncertain thickness to reach ice of unknown purity and thickness. These parameters remain poorly constrained across the broader VFF population. SHARAD observations suggest overburden debris/dust cover may reach ~15 m (Holt et al., 2008; Plaut et al., 2009).

CCFs nevertheless present substantial operational challenges for in-site resource utilisation (ISRU) investigation, owing to their confined topography, depth below the crater rim, steep interior slopes, rough surface textures, and the likely strong deformation and complexity of their internal stratigraphy. However, closely analogous constraints are already being confronted in proposed in situ exploration concepts for permanently shadowed regions on the Moon, where access, mobility, illumination, and sampling of ice-rich substrates within steep-walled, topographically enclosed environments pose comparable engineering problems (e.g., Russell et al., 2022). Technologies being developed to enable landing, traversal, and subsurface sampling in these lunar settings could plausibly be adapted to Martian CCFs. In this sense, although CCFs are not an obvious or low-risk target, they constitute a technically feasible class of sites for future ISRU.

2.5. CCFs: remaining knowledge gaps

Whilst CCFs are the most numerous and spatially extensive class of mid-latitude glaciers on Mars, they remain surprisingly understudied. Consequently, substantial uncertainties persist regarding their origin, evolution, and climatic significance. In this section, I distinguish between (i) the long-standing,

unresolved problems concerning CCF origin, structure, and climatic significance that define the broader scientific context of this Thesis (Section 2.5.1), and (ii) the targeted, testable knowledge gaps that provide the direct impetus for the research presented in this Thesis (Section 2.5.2).

2.5.1. Broader scientific questions concerning CCF formation, structure, and climatic interpretation

This subsection summarises broader scientific questions concerning the origin, structure, and climatic significance of CCFs. These questions define the wider field context for this Thesis, but they are not each resolved directly or completely by the analyses presented here. Instead, they identify the larger unresolved problems to which the targeted, chapter-level research questions in Section 2.5.2 contribute.

What is the origin of CCFs as part of Mars' mid-latitude glacial history and cryospheric evolution?

As discussed in Sections 2.4.1 and 2.4.4, the glacial processes that formed CCFs is still heavily debated. The fundamental problem is that surface morphology alone cannot tell us if the ice accumulated entirely confined within host craters or if it is remnant of a much larger, regional ice sheet (Sections 2.4.1; 2.4.4). Current hypotheses span crater-confined accumulation analogous to alpine-style glaciation, in which ice builds locally on the internal crater walls and remains topographically bounded; the generation of localised ice caps, over-topping crater rims; and emplacement during more extensive ice-sheet-like conditions that would have promoted regional connectivity and subsequent isolation during retreat.

All three pathways are physically plausible under different combinations of accumulation rate, equilibrium line altitude and duration. Yet, the key discriminants—peak ice thickness relative to rim height, timing of emplacement and the frequency of rim overtopping—are poorly constrained. Part of this stems from the limited evidence of more extensive deposits (e.g., Levy et al., 2010; Dickson et al., 2010). More fundamentally, however, any interpretable imprint of past ice extents is likely to be extremely subtle—or non-existent if the ice never breached the rim—and current orbital observations lack the ground-based field techniques required to probe subsurface sediments or identify potentially eroded paleo-surfaces. Consequently, without a clear view of CCF internal stratigraphy through either radar sounding or ice coring, it remains an open question whether a CCF is a single, continuous mass of high-purity ice or a composite feature built over multiple climatic cycles. Because of these uncertainties, it is possible that the global CCF population records more than one glacial accumulation pathway, rather than a single, uniform mode of ice emplacement. Resolving this problem is central to reconstructing CCF formation pathways and, by extension, to constraining the true style and scale of Amazonian mid-latitude glaciation.

What is the origin of concentric features on CCFs?

Arguably, the most defining and visually striking characteristic of a classic CCF is its concentricity—a system of nested ridges and tracelines arranged parallel to the crater interior (e.g., Squyres, 1979). While

curvilinear structures are common across VFFs (e.g., Squyres, 1979; Milliken et al., 2003; Head et al., 2010), CCF organisation into concentric patterns is a unique geometry that is still not fully understood. This distinction suggests that the concentric geometry of CCF tracelines may reflect a specific combination of flow confinement, internal stratigraphy, and surface exhumation that is not captured within other non-CCF VFFs, motivating closer examination of terrestrial processes capable of generating concentric surface outcrop patterns.

Terrestrial evidence suggests that these nested patterns can emerge when internal glacier stratigraphy is folded and subsequently exposed through surface lowering. Under conditions of lateral compression and simple shear, primary ice layers can deform into domed syncline/anticline geometries (e.g., Goodsell et al., 2005; Jennings and Hambrey, 2021). As ablation thins the glacier, these internal folia intersect the surface to create concentric outcrop patterns. A terrestrial example is found in the Sherman Glacier, Alaska (Figure 2.26) where the melting of folded ice has revealed a nested stratigraphy of concentric layers (Molnia, 2004). By extension, the tracelines on a CCF may not require a specific surface-forming process; they may simply be the "top-down" view of internally structured ice that has been exposed as the feature thinned and manifests through the surficial dust/debris layer over time.



Figure 2.26. Nested internal layering exposed by folding and ablation in a terrestrial valley glacier (Sherman Glacier, Alaska). East-looking oblique photograph of the middle reaches of Sherman Glacier, Chugach National Forest, showing concentric foliation revealed by structural folding and subsequent surface lowering. Image credit: Molnia, 2004.

Numerical models specific to Martian craters offer a complementary perspective based on the physics of bounded, convergent ice flow. The “crumpling” model of Weitz et al. (2018) simulates CCF formation by accumulating ice on crater walls during intermediate-obliquity states, with discrete ice lobes flowing inward and converging toward the crater centre. As the crater fills in the model, ice thickens and flow velocities decrease, producing low-relief, concentric undulations in the simulated ice surface. Debris is advected with the ice and becomes locally concentrated during phases of reduced flow or stagnation, enhancing the visibility of these surface features, but the topographic pattern itself arises from the convergence of ice flow within a confined circular basin. When translated into three dimensions, these surface undulations correspond to basin-parallel, concentric ridges comparable in scale and geometry to those observed on CCFs (Weitz et al., 2018; Figure 2.27). Although the model explicitly tracks only surface topography and debris thickness, it demonstrates that crater-bounded convergence of ice is, in principle, sufficient to generate concentric surface structure.

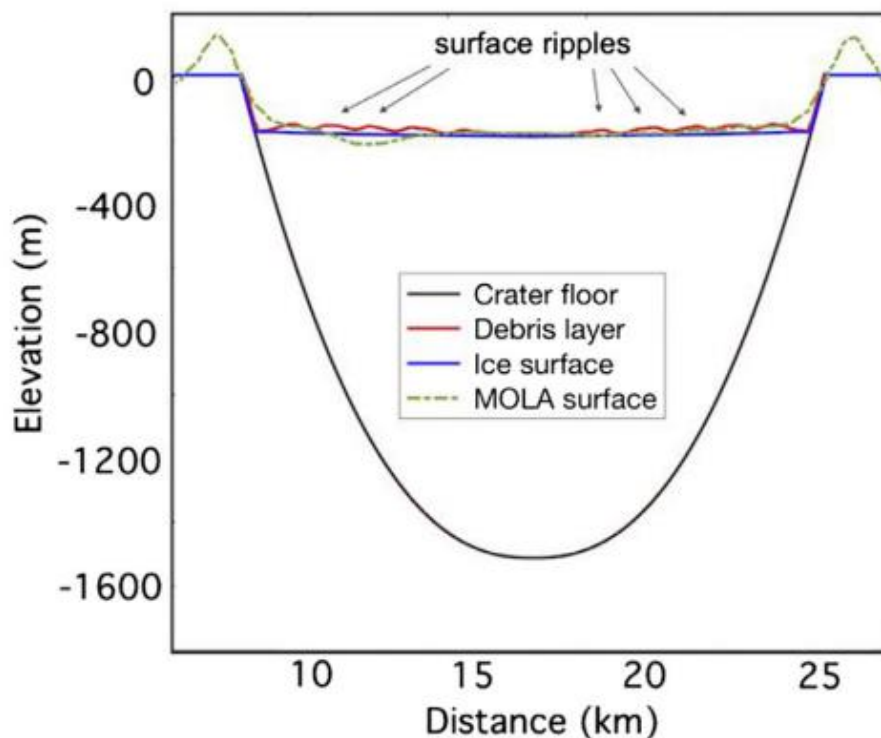


Figure 2.27. Modelled surface topography produced by inward flow of crater-wall ice. The figure shows simulated present-day ice thickness and debris distribution, illustrating the development of surface ridges and undulations generated by convergence and thickening of ice toward the crater centre. Modified from Weitz et al. (2018).

Although the Weitz et al. (2018) model does not explicitly represent internal stratigraphy, the development of systematic surface undulations by converging crater-wall flows is likely to be accompanied by distributed deformation of the underlying ice mass. In glacier-like materials, ice flow involves both horizontal and vertical motion, such that compressive flow in down-glacier or basinward settings is expected to promote upward advection of deeper ice and associated internal layers toward

the surface (e.g., Nye, 1952; Hooke, 2019; Arnold and Butcher, 2025). Under such conditions, lateral convergence and longitudinal compression are unlikely to be accommodated solely at the surface, and would instead be expressed throughout the CCF, reorienting internal stratigraphy as well as the surface topography. Recent observations of Martian VFFs are consistent with this expectation: Butcher et al. (2024) and Arnold and Butcher (2025) document inclined and progressively rotated internal layering exposed in incised VFF margins, with layer geometries that steepen down-flow and connect directly to arcuate surface structures near flow termini, closely analogous to fold-and-thrust architectures at terrestrial glacier snouts. These observations indicate that surface lineations on VFFs can plausibly represent the surface expression of bulk ice deformation and internal stratigraphic warping, rather than features confined to a superficial debris mantle. In this context, the surface “ripples” (Figure 2.27) produced in the Weitz et al. (2018) simulations are best interpreted as the surface expression of convergent flow within the ice body, even though internal layering was not explicitly modelled in that study.

Do CCFs record a history of Martian glaciation and environmental change?

As discussed, multiple formation models suggest that CCF concentric manifestations are the surface expression of internal, ice-rich layers deformed by flow and sublimation. If tracelines are indeed manifestations of internal stratigraphy shaped by orbital cycles, then perhaps a CCF surface is essentially a climate record—analogous to the stratigraphic layers found in terrestrial glaciers.

The challenge is that CCF surfaces are not preserved equally across all features. There is a clear morphological divide between “classic” CCFs, which retain sharp, well-developed tracelines, and ‘low-definition’ features that appear subdued or buried (Levy et al., 2010). While both likely share a glacial origin, their current states represent a spectrum of preservation. Low-definition features may primarily reflect recent burial or degradation, whereas classic CCFs likely retain the most legible signals of past ice-flow and accumulation history.

Despite this, most previous work has grouped CCFs into a single population (e.g., Levy et al., 2010, 2014; Dickson et al., 2012), potentially masking primary glacial signals with noise from post-glacial modification. Consequently, the capacity for ‘classic’ CCFs to act as high-fidelity climate archives remains largely untested. By isolating and analysing this well-preserved subset, we can move beyond mere degradation assessments to explore whether their distribution, geometry, and surface properties preserve a systematic record of past orbital forcing.

Why do CCFs reside in some craters and not others?

Although CCFs are near-ubiquitous across the Martian mid-latitudes (Levy et al., 2014), they are absent from most individual craters. CCFs are uncommon in craters smaller than ~2 km in diameter and are also rare in the largest basins (>~80 km), indicating that crater size imposes first-order constraints on

their formation or long-term survival (Levy et al., 2014). Yet even within the favourable diameter range, CCFs do not occur in every crater, including among craters of similar size, morphology, and latitude, highlighting the role of additional controlling factors beyond simple geometric suitability.

A simple, non-climatic explanation would be that craters of similar size and morphology can differ substantially in formation age and therefore in their availability during mid-late Amazonian glaciation. However, the secular decline in impact flux from the Noachian into the Hesperian and Amazonian (Section 2.1.1; Hartmann and Neukum, 2001) suggests that most mid-latitude craters in the relevant diameter range likely originate from the Noachian or Hesperian, pre-dating palaeo-CCF ice emplacement, making systematic youthfulness of the unfilled crater population an unlikely explanation for their close spatial intermixing with filled craters.

The close spatial intermixing between craters that host CCFs and adjacent, morphologically similar craters which do not therefore suggests that large-scale regional variations in ice mass balance alone cannot explain their current distribution and instead points to crater-scale controls on whether ice can accumulate or survive within a given basin. Variations in rim degradation, interior slope geometry, and crater-scale microclimates may plausibly modulate both palaeo-CCF ice accumulation and long-term preservation by influencing debris and dust supply to the ice surface, the degree of topographic shading and insolation-driven ablation, and local wind and vapour-transport conditions within the crater interior.

However, these local variables do not operate in isolation. Spatial variations in latitude, elevation, or broader topographic context may regulate spatial-temporal variations in ice mass balance, thereby conditioning whether local crater environments fall above or below thresholds for sustained initial ice accumulation or post glacial retention. In this view, the absence of CCFs from otherwise suitable craters may reflect limits imposed by regional mass balance—such as insufficient accumulation or enhanced post-glacial ablation—rather than the absence of favourable crater geometry alone. Solving this requires disentangling the influence of the host crater from broader climatic forcing to determine if the present CCF distribution primarily records spatial variability in past glacial conditions or differential preservation through subsequent modification.

Beyond the presence or absence question, it remains unclear if well-defined ‘classic’ CCFs follow the same spatial distribution as the combined inventory (Levy et al., 2014) or instead cluster regionally. If the ‘classic’ subset and the global general population distributions differ, it would effectively map out regions of preferential CCF preservation. Finally, while it is established that slope availability dictates the location of non-CCF VFFs (e.g., Milliken et al., 2003), it remains unclear if spatial variations in CCF abundance simply track crater density, such that areas with greater CCF simply have a higher density of craters, or if they instead encode environmental signals. Testing whether CCF occurrence varies systematically with latitude and elevation—independent of crater counts—is essential to separating simple geometric opportunity from true climatic influence.

2.5.2. Research questions addressed in this Thesis

The broader scientific questions outlined above motivate this Thesis, but the empirical analyses address a more targeted set of chapter-level research questions. These questions correspond to the empirical chapters introduced in outline in Section 1.2 and are developed here in greater detail following the literature review. Together, they test how far CCF distribution, surface depth, preserved volume, and surface orientation can be interpreted as records of Amazonian mid-latitude glaciation rather than as consequences of host-crater geometry and preservation state.

Before these chapter-level research questions can be addressed, a controlled CCF population suitable for comparative geometric analysis must first be established. Squyres (1979) described CCFs as crater-filling deposits characterised by sharply defined concentric lineations, termed *tracelines* in this Thesis. Subsequent studies recognised a broader population of subdued or ‘low-definition’ examples, commonly interpreted as mantled or degraded forms that may share a glacial origin but whose surfaces have been modified or obscured by later mantling, deformation, and/or ice loss (e.g., Levy et al., 2010, 2014). The global inventory of Levy et al. (2014) contains 9,533 candidate CCFs distributed across the Martian mid-latitudes but combines well-preserved, low-definition, and uncertain examples. Chapter 3 therefore re-evaluates this inventory to isolate high-confidence CCFs retaining clear and unambiguous viscous-flow morphologies, distinguish them from low-definition and uncertain candidates, and characterise their surface morphologies and traceline expressions. The resulting population provides the controlled dataset used throughout the subsequent analyses.

Alongside the chapter-level questions introduced below, one cross-cutting research question runs through the empirical chapters:

What is the effect of the host crater on CCF evolution (Chapters 4–7)?

It is likely that a crater’s specific geometry and modification history—during and prior to CCF occupation—dictate whether a CCF can form or persist. Crater depth and shape (from simple bowl-shaped to more complex-shape), and volumetric capacity scale systematically with diameter (Garvin et al., 2003), thus as all present-day CCFs are crater-confined, the host crater directly sets the maximum attainable CCF thickness and volume. Although the observed diameter range of CCF-hosting craters is well constrained (~2–80 km; Levy et al., 2014), it remains unclear whether CCFs preferentially occupy particular crater sizes once background crater availability is taken into account. In other words, it is not known whether CCF occurrence simply tracks where craters exist, or whether certain crater sizes are disproportionately likely to host CCFs.

The condition of the crater rim likely plays a decisive role in microscale influences on CCF mass balance. A well-preserved, high rim may provide topographic shading to the CCF surface and/or a steady supply of debris delivery to the crater interior. Conversely, a degraded or breached rim may leave

the interior more vulnerable to insolation or fail to provide the debris necessary for an insulating lag, resulting in reduced net accumulation during glacial periods and enhanced ablation during interglacial periods. However, the true extent of the crater rim's influence likely cannot be determined without identifying the origin of the overburden—specifically the relative contributions of hillslope-derived mass wasting versus regional atmospheric dust fallout.

Existing modification within the crater also matters. If a crater was partially resurfaced before the palaeo-CCF accumulation, the volumetrically reduced cavity space may prevent CCF preservation or formation. Host-crater position may also modulate local energy balance through background topographic shading, slope orientation, or regional dust availability, introducing additional microclimatic controls on ice stability. Ultimately, a CCF is not just a product of regional climate; it is the result of host crater-specific variables that govern whether ice can be preserved (or perhaps even form) over geological time. Despite their importance, these crater-scale controls remain largely under-researched. The influence of the host crater (namely its diameter, depth and volume) is tested throughout Chapters 4–7.

Does the global distribution of high-confidence CCFs reflect environmental controls or the availability of craters suitable for glaciation? (Chapter 4)

Whilst VFFs cluster within the Martian mid-latitudes (e.g., Squyres, 1978, 1979; Milliken et al., 2003; Souness et al., 2012; Levy et al., 2014), and non-CCF VFFs generally cluster on steep slopes (e.g., Milliken et al., 2003; Levy et al., 2007; Dickson et al., 2008; Souness et al., 2012; Brough et al., 2019), it remains unresolved whether the spatial distribution of CCFs—the most abundant VFF—reflects glacial–environmental controls or simply the availability of suitable host craters. Global maps document the combined distribution of classic and low-definition CCFs (Levy et al., 2014), but they do not test whether this distribution differs in a statistically meaningful way from the background crater population. As a result, it is currently unclear whether latitudinal, elevational, or regional variations in CCF abundance merely mirror variations in crater density or instead encode signals of spatially variable accumulation and/or preservation.

Furthermore, it remains unknown whether the classic CCF subset follows the same near-ubiquitous spatial distribution as the full inventory or instead exhibits regional clustering. If classic and low-definition CCFs display different spatial distributions, this may imply that they record different combinations of climatic controls, preservation filtering, or geometric opportunity. If, however, both populations share the same widespread distribution, this introduces an additional constraint to the question: not only must we explain why some craters host CCFs and others do not, but also why some craters preserve well-defined classic CCF forms whereas others retain only degraded, low-definition remnants. These questions regarding classic-CCF spatial distribution are addressed in Chapter 4.

Why do CCFs have variable surface depths below their rim (Chapter 5)?

A persistent but poorly explained property of CCFs is the large variability in their present-day surface elevation relative to their host crater rims, with some nearly rim-full and others inset by several hundred metres even among craters of similar size and morphology (Pedersen and Head, 2010; Levy et al., 2014). This variability could, in principle, arise from differences in the total amount of ice originally emplaced, differences in the magnitude of post-glacial mass loss, or purely geometric effects associated with crater size and shape, which control both accommodation space and the potential for topographic shading (e.g., Garvin et al., 2003). The key unresolved question is therefore not whether CCF surfaces differ in their elevation below the rim, but whether this variability is dominated by crater geometry alone or whether it also encodes systematic environmental controls related to latitude, elevation, and regional climate setting.

This problem is addressed directly in Chapter 5, which tests whether CCF surface elevation relative to the crater rim scales only with host-crater geometry or whether additional spatial structure emerges once geometric effects are accounted for. By examining how CCF surface elevations vary with crater size, latitude, elevation, and regional context, Chapter 5 evaluates whether these differences primarily reflect crater-controlled accommodation and preservation or instead record spatially organised signals of accumulation and post-glacial ablation.

Do preserved CCF volumes record climate, or only crater capacity (Chapter 6)?

Because crater depth and shape scale systematically with diameter (e.g., Garvin et al., 2003), larger craters provide greater geometric capacity for crater-confined ice. It remains unresolved whether present-day CCF volumes are controlled primarily by this geometric capacity or whether they retain a measurable imprint of spatially variable accumulation, post-glacial mass loss, or long-term preservation efficacy.

Existing volume estimates (e.g., Levy et al., 2014) have not been analysed within a framework that explicitly separates geometric capacity from environmental signal, nor used to test whether different regions occupy systematically different parts of that capacity space. As a result, it is not known whether CCF volumes primarily record what each crater can hold, or what climate and glacial history supplied and removed. This gap motivates the volume-scaling analysis in Chapter 6.

Do CCF surface slopes and aspects record asymmetric accumulation or ablation (Chapter 7)?

While non-CCF VFFs show strong pole-facing and secondary east-facing orientation biases (e.g., Milliken et al., 2003; Souness et al., 2012), CCFs have generally been assumed to be orientation-neutral due to their concentric geometry. This assumption has never been tested quantitatively. It therefore remains unknown whether CCF surfaces record asymmetric snow deposition, preferential insolation-driven ablation, post-glacial modification, or simply crater-imposed shape.

Dickson et al. (2012) inferred flow directions from surface textures but did not perform a quantitative slope–aspect analysis, nor isolate CCFs as a distinct population. As a result, the potential climatic or glaciological significance of CCF surface geometry remains essentially unexplored. This gap motivates the slope–aspect analysis in Chapter 7.

Addressing these questions requires moving beyond treating CCFs as a single, uniform population. A necessary first step is therefore to establish a robust classification framework that distinguishes well-preserved, morphologically diagnostic ‘classic’ CCFs from degraded or ambiguous examples, and to construct a controlled subset of the Levy et al. (2014) inventory. Chapter 3 now develops this classification scheme and defines the CCF datasets used throughout the remainder of this thesis.

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Chapter 3: CCF Variety, Classification and Inventory

CCFs are the most spatially widespread and numerous mid-latitude ice-related landforms on Mars (Levy et al., 2014). However, they display considerable morphological diversity, ranging from well-preserved examples with sharply defined concentric lineations and distinctive viscous flow features (e.g., Squyres, 1979), to more ambiguous forms described as “low-definition” CCFs (e.g., Levy et al., 2010). A clear and rigorous CCF classification system is therefore essential for any global investigation of their climatic significance and glacial history.

This chapter’s purposes is to establish explicit criteria for the identification of CCFs, to describe the range of morphologies observed, and to present the final CCF inventory used as the input to all subsequent analyses in this Thesis. This Thesis uses the global CCF inventory assembled by Levy et al. (2014) as the initial inventory, with each feature re-assessed according to the classification criteria defined here, rather than generating a new global catalogue independently.

The Levy et al. (2014) CCF inventory contains 9,533 mapped CCFs across $\sim 30\text{--}55^\circ\text{N/S}$, identified primarily from CTX mosaic images ($\sim 18\text{ m/pixel}$) and supplemented by THEMIS-VIS data. As the only global catalogue of CCFs, it intentionally adopted broad inclusion criteria to capture the full range of crater-confined ice candidates recognised at the time. Accordingly, the inventory encompasses classic examples with prominent concentric lineations, low-definition CCFs where surface expression is muted, pole-facing flow features now interpreted as younger intra-crater ice flows (e.g., Sinha and Vijayan, 2017), and deposits where mantling, aeolian cover, partial infill, or degradation complicate interpretation. This comprehensive starting point motivated the need for a more narrowly defined classification to enable isolation of high-confidence CCFs for quantitative analysis in subsequent chapters. This chapter addresses this need.

From the 9,533 entries in that inventory, I identify and extract a subset that meet my classification of a CCF – as outlined in Section 3.2. My definition is deliberately strict, aimed at retaining only those CCFs showing unambiguous evidence of ice-related processes, while excluding low-definition features with ambiguous or uncertain characteristics that cannot be confidently attributed to glacial origins. This reduces interpretative uncertainty and ensures that the retained sample comprises CCFs with the highest likelihood of preserving diagnostic evidence of past glacial activity. The resulting high-confidence CCF inventory will be used as the input data in the later chapters for evaluating the spatial pattern of mid-latitude glaciation and its influence by environmental factors/gradients.

This chapter is structured in three parts. Section 3.1 reviews and clarifies how CCFs have been defined and classified in previous studies (e.g., Levy et al., 2010), highlighting ambiguities around partially infilled and “low-definition” forms—establishing the conceptual motivation for an explicit classification standard. Section 3.2 sets out my CCF identification criteria in this Thesis and applies

them to the Levy et al. (2014) inventory to derive my CCF inventory; I also discuss low-definition cases that were excluded. Section 3.3 presents the inventory and examines the variety of surface morphologies and traceline expressions of its CCFs, from well-defined ‘classic’ examples to a spectrum of ‘non-classic’ morphologies encompassing degraded, mantled, and otherwise irregular forms.

3.1. Definition of Concentric Crater Fill

CCFs (Figure 3.1), a morphological subtype of viscous flow features, were first observed and described by Squyres (1979) as “crater fill possessing a concentric pattern of surface undulation” (shown in Figure 3.1A). Broadly, they are glacial units which infill impact craters. Their surfaces host curvilinear tracelines which typically form multiple concentric loops (Figure 3.1B). CCFs are found in Mars’ mid-latitude regions (approximately 29–60°N and 29–55°S; Levy et al. 2014), coexisting with other morphological subtypes of Viscous Flow Features (VFFs) (Squyres, 1979; Squyres and Carr, 1986; Milliken et al., 2003; Souness et al., 2012). Given the morphological overlap with related viscous-flow deposits, this Thesis applies strict, morphology-based criteria to identify CCFs and ensure a high-confidence inventory for subsequent analyses.

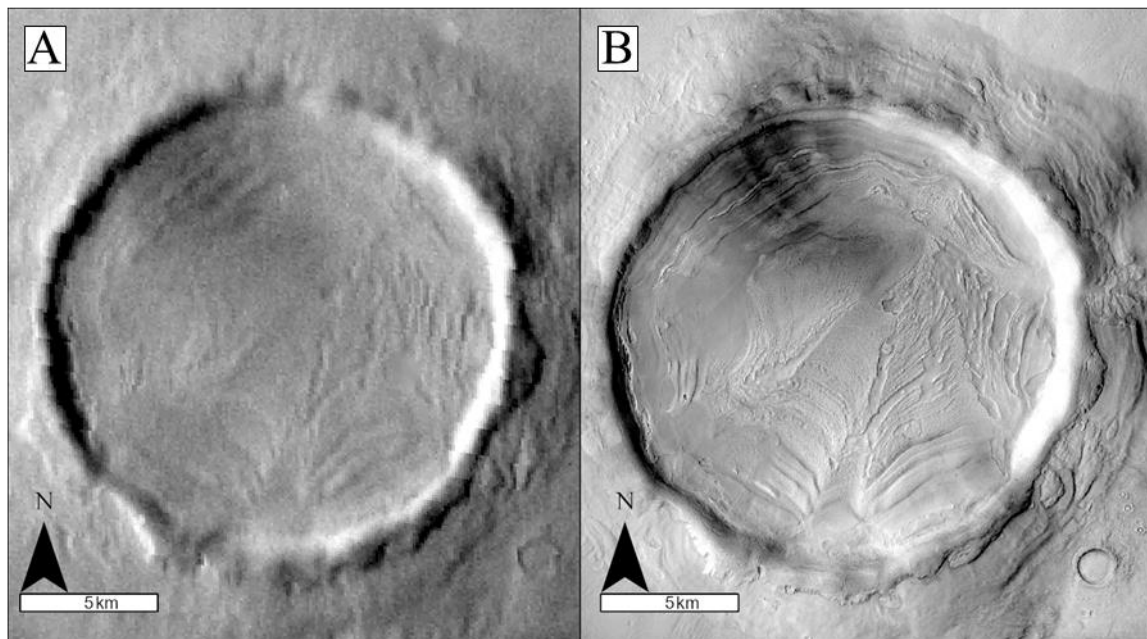


Figure 3.1. Comparison of early and modern imaging of a Concentric Crater Fill (CCF). (A) The CCF at 41.51° N, 87.72° E as originally identified in Viking images (Squyres, 1979; Viking frame 010B70). (B) The same CCF in CTX images, where the higher spatial resolution reveals clearly defined tracelines (CTX image P22_009533_2219).

Although CCFs are widely recognised as crater-confined VFFs characterised by surface lineations and convex profiles (e.g., Squyres, 1979, 1989; Milliken et al., 2003), the operational definition of what constitutes a CCF has not been applied consistently. In Viking images, Squyres (1979) identified CCFs as crater-filling deposits with concentric undulations, and later clarified that complete floor coverage is

fundamental, explicitly distinguishing inward-facing wall aprons that had not yet buried the crater floor as not CCFs—though they could evolve into CCFs if infilling continued (Squyres, 1989). In later mapping efforts, however, such partially filled craters were often incorporated into CCF inventories as “low-definition” examples where lineations were subtle or convex flow forms extended downslope from crater walls (Levy et al., 2009, 2010, 2014; Pedersen and Head, 2010; Dickson et al., 2012). These inward-facing deposits share some surface textures with CCFs (e.g., flow ridges, brain terrain; Levy et al., 2010; Pedersen and Head, 2010; Sinha and Vijayan, 2017) but are now interpreted as smaller-scale, more recent glacial features emplaced under different climatic regimes, with crater-retention model ages of ~10–100 Ma (Sinha and Vijayan, 2017).

Concurrently, inventories expanded to include “low-definition” CCFs lacking clearly resolved lineations, based on muted undulations, subtle stippling or widened fractures where concentric tracelines could not be confidently identified (Levy et al., 2010). Supplementary viscous-flow indicators—such as brain terrain, ring-mold craters, and pitting—were also used to imply glacial origin of these deposits and support CCF interpretation in the absence of tracelines (e.g., Levy et al., 2010; Head et al., 2010; Levy et al., 2014). While this approach captured a continuum from well-preserved to heavily modified/degraded CCFs forms (Pedersen and Head, 2010), it also resulted in inventories in which deposits with differing glaciological histories and potentially distinct emplacement modes (e.g., pole-facing wall lobes) appeared under a single label.

In a broad geomorphological sense, imposing a hard boundary between ‘CCF’ and related viscous-flow landforms is arguably artificial; these features likely form a spectrum of crater-confined and wall-sourced ice accumulations (e.g., Squyres, 1989; Pedersen and Head, 2010; Levy et al., 2010). However, this Thesis requires a constrained subset of unambiguous CCFs to enable robust quantitative analysis of their geometry, volume, spatial controls, and glacial histories. Accordingly, the criteria adopted in Section 3.2 distinguish fully crater-filling examples displaying diagnostic concentric tracelines from incomplete, ambiguous, or heavily mantled cases. The aim is not to redefine the landform class for the wider field, but to isolate a high-confidence CCF population appropriate for reproducible morphometric and volumetric investigation.

3.2. CCF identification Criteria

Here, I outline the CCF identification criteria used in this Thesis (Figure 3.2). The classification scheme is based on observable properties in orbital CTX images and explicitly distinguishes definitive CCFs from ambiguous crater fills and non-CCF deposits. Figure 3.2 provides an overview of the decision pathway, while Table 3.1 presents the full criterion combinations and classification outcomes, with representative examples shown in Figure 3.3.

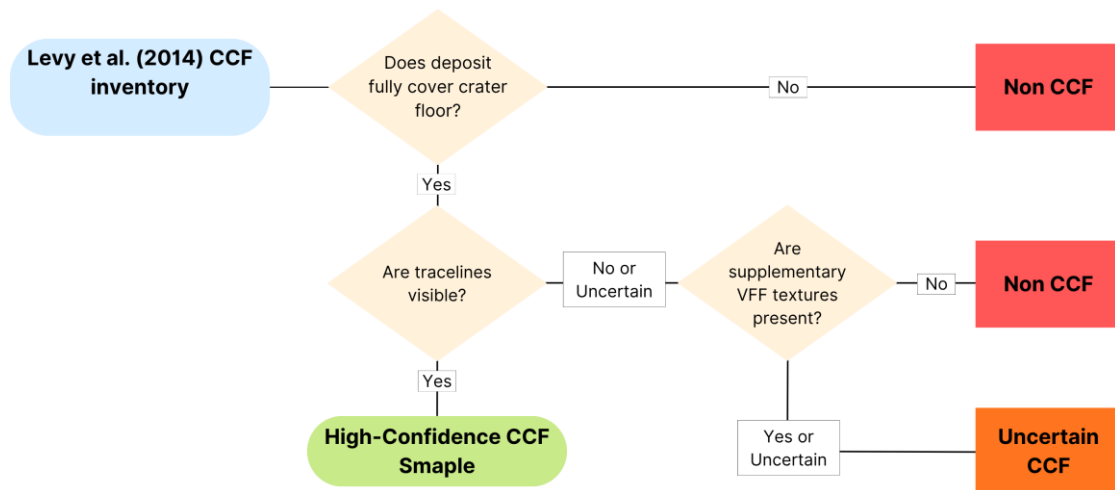


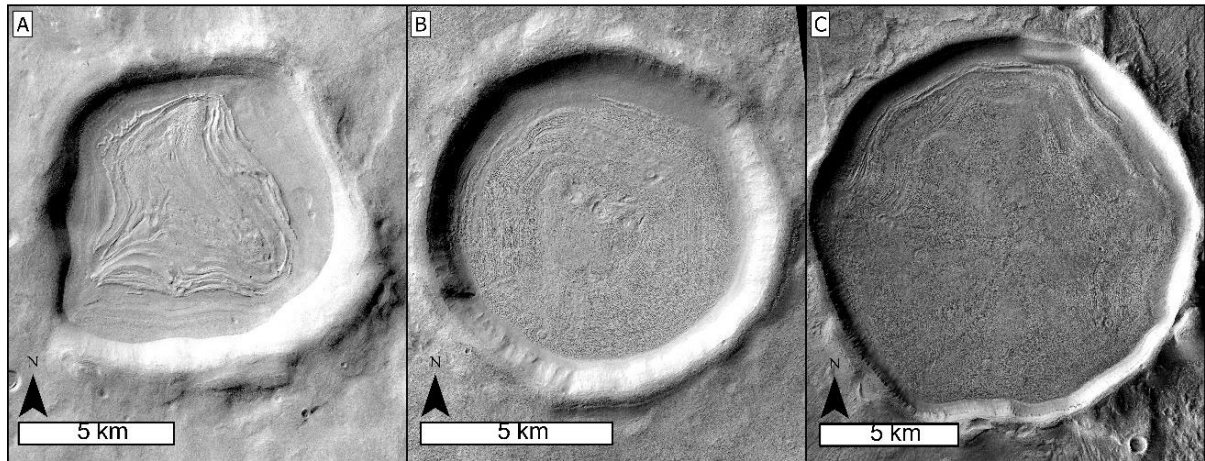
Figure 3.2. Workflow pathway used to classify candidate CCFs. The workflow summarises how crater-floor coverage, visible tracelines, and supplementary viscous-flow morphologies are combined to distinguish between high-confidence CCFs, uncertain CCFs and non-CCF deposits.

Table 3.1. Comprehensive list of potential combinations of my CCF identification criteria.

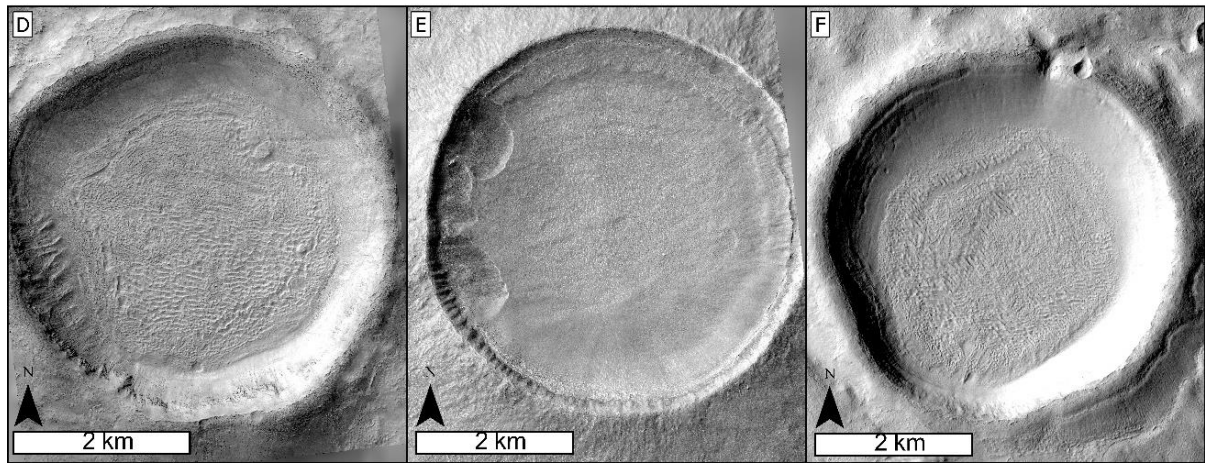
Classification	Covers the host crater floor (covering any central peak)	Tracelines present	Typical supplementary VFF morphologies present	Figure example
High confidence CCF	✓	✓	✓	3.3A-C
	✓	✓	uncertain	None found
	✓	✓	✗	None found
Uncertain CCF	✓	uncertain	✓	3.3D
	Uncertain	✓	✓	None found
	✓	✗	✓	3.3E
	✓	uncertain	uncertain	3.3F
Non CCF	✓	✗	✗	3.3G
	✗	✓	✓	3.3H
	✗	✓	✗	None found
	✗	✗	✓	3.3I
	✗	✗	✗	3.3J

Table 3.1 Comprehensive list of potential combinations of CCF identification criteria. The first column assigns each combination to one of three classification outcomes (CCF, uncertain CCF, non-CCF) based on the presence or absence of floor-filling, tracelines, and supplementary VFF morphologies. The order of criteria from column 2 (fills the crater) through to column 4 (typical supplementary VFF morphologies) indicates a broad priority-order of criteria for a feature being identified as a CCF. If a feature does not fill the crater floor, it is not defined as a CCF for the purposes of this study, regardless of whether it fulfils subsequent criteria in columns 3 and 4. At this stage of CCF identification, several combinations of criteria lead to an ‘uncertain’ categorisation. The figures corresponding to examples of each combination of criteria are provided for reference, with the rightmost column pointing to image figures showing specific examples of these combinations.

High Confidence CCFs



Uncertain CCF



Non CCFs

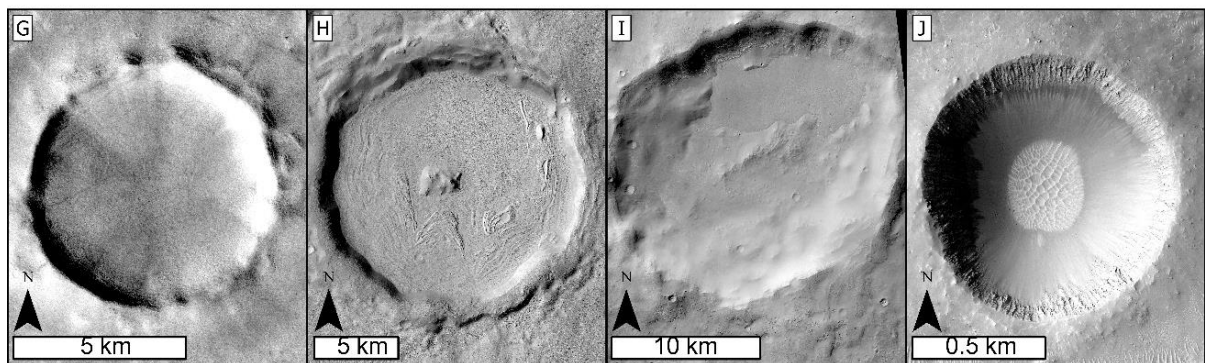


Figure 3.3. Examples illustrating the CCF classification scheme used in this Thesis (Table 3.1). (A–C) High-confidence CCFs: craters whose floors are completely covered by CCF and display clear trachene morphologies; supplementary viscous-flow features (e.g. brain terrain, ring-mold craters) are also present. (A) 32.32° S, 52.02° E; CTX D16_033349_1478. (B) 43.32° S, 85.55° W; CTX B16_016001_1358. (C) 44.11° S, 83.15° W; CTX P05_002947_1355. (D–F) Uncertain CCFs: craters whose floors are covered but where trachenes and/or viscous-flow morphologies are ambiguous or obscured, commonly due to mantling or surface degradation. (D) 43.86° S, 159.51° E; HiRISE ESP_033411_1360. (E) 42.99° S, 98.38° E; HiRISE ESP_023062_1365, showing partial mantling by LDM with pitted textures and scalloping. (F) 43.08° S, 169.92° E; CTX B05_011482_1375, where uncertainty arises from a single possible trachene and abundant parallel, aeolian-like surface textures. (G–J) Not CCFs: examples that fail one or both primary criteria. (G) 60.99° S 114.32° E; CTX D02_028085_2388, a smooth crater fill lacking trachenes or viscous-flow morphologies.

Figure 3.3. (continued) (H) 30.29° N, 41.56° E; CTX G22_026690_2098, a crater fill that does not completely cover the crater floor (central peak exposed) despite the presence of tracelines and supplementary viscous-flow features. (I) 33.68° S, 82.53° W; CTX B17_016225_1473, an incomplete crater fill with no tracelines, although brain terrain and ring-mold craters occur on adjacent smooth terrain. (J) 3.51° S, 3.76° W; HiRISE ESP_063878_1765, an unfilled impact crater with exposed bedrock.

Three observable properties underpin the classification scheme displayed in Figure 3.2 (and each elaborated in Sections 3.2.1–3.2.3):

1. **Coverage of the host crater floor (including any central peak)** – the deposit must completely cover the host crater floor, including any central peak. This establishes a true crater-filling deposit rather than a localised wall-slope accumulation.
2. **Presence of traceline lineations** – repeating elongated ridges and troughs, often concentric and curvilinear in expression, must be visible, indicating viscous deformation of an ice-rich interior.
3. **Supplementary viscous-flow morphologies** – such as brain terrain, ring-mold craters or pitted textures where present, that further indicate the presence of an ice-rich interior.

While all three properties strengthen confidence in glacial origin, only the first two are required to classify a feature as a definitive CCF. Complete floor coverage establishes crater-filling geometry, and the presence of tracelines represents diagnostic surface expression of viscous flow. Thus, both must be present (Table 3.1, Rows 2–4) for inclusion in the high-confidence CCF inventory used in this Thesis. Where tracelines are ambiguous or obscured, supplementary viscous-flow morphologies are used as supporting evidence to distinguish between uncertain CCFs and non-CCF deposits. Table 3.1 expresses these three properties as yes/no/uncertain combinations and links each combination to a classification outcome listed in the leftmost column of the table. The logic is as follows:

- **High-confidence CCF** – host crater floors are completely covered *and* tracelines visible (supplementary VFF morphologies may or may not be present).
- **Uncertain CCF** – key indicators are ambiguous or incomplete (e.g., crater floor covered but tracelines obscured possibly due to mantling or degradation). Such cases may represent degraded or ice-rich deposits but cannot be confidently interpreted; these are excluded from the primary inventory but noted for replication transparency.
- **Not a CCF** – crater-floor coverage is incomplete and/or tracelines are absent, consistent with partial wall accumulations, sedimentary infill, or other non-diagnostic crater-floor materials.

To compile the high confidence CCF inventory, I re-examined every candidate entry in the Levy et al. (2014) global catalogue. Each feature was inspected using MRO Context Camera (CTX; ~5–6 m/pixel; Malin et al., 2007) images, which provided the consistent resolution and complete spatial coverage

necessary for a global study (Table 1.1). Where available, higher-resolution HiRISE (25 cm/pixel; McEwen et al., 2007) images were used to examine fine-scale morphology, though it was not used to determine whether a feature met the CCF criteria due to partial spatial coverage. By applying this specific combination of criteria, I ensure that the CCF sample used in this Thesis is restricted to unambiguous, morphologically diagnostic examples of crater-confined glacial ice.

Although many mid-latitude craters contain infill deposits that may be ice-rich, not all such features meet the criteria for inclusion in the present CCF inventory. Craters in which the interior deposit lacks visible tracelines — including highly degraded examples or cases where thick mantling prevents confident interpretation — are excluded because their glacial origin cannot be demonstrated with sufficient certainty. Likewise, glacial deposits that only occupy part of the crater (e.g., pole-facing wall lobes or intra-crater LDAs that do not cover the crater floor) are not classified as CCFs. These exclusions ensure that subsequent quantitative analyses are based on a high-confidence CCF population. Nonetheless, prior studies have shown that degraded and partially preserved viscous-flow features can retain valuable palaeoclimatic information (e.g., Pedersen and Head, 2010), and so whilst these features are excluded, their broader significance is acknowledged even though they are not investigated further here.

3.2.1. Covering the host crater floor

The first criterion for CCF identification is that the post-impact fill completely covers the floor of the host crater, including any central peak structure (Figure 3.1, Figure 3.3A–C). Given the criterion, deposits that cover most of the crater floor but leave a central peak exposed are not classified as CCFs in this study (e.g., Figure 3.4). Such deposits may represent either incomplete crater-floor infill or former thicker deposits that have undergone sufficient post glacial ablation to re-expose the peak. While the latter possibility acknowledges a potential continuum of degradation states, the present-day surface expression does not satisfy the defining geometric requirement of full crater-interior coverage and is thus excluded from the high-confidence CCF population analysed here.

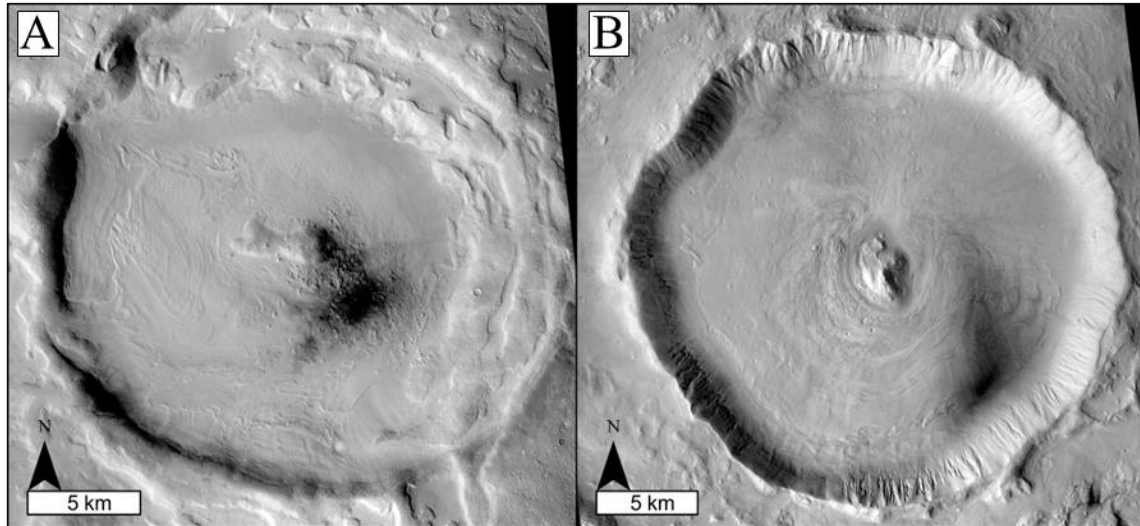


Figure 3.4. Examples of crater-confined exposed-floor VFF, that do not meet the CCF classification. (A) A crater partially occupied by a LDA that extends from the crater walls across the floor but does not constitute a fully crater-filling deposit (41.76° N, 38.32° E; CTX image: P16_007188_2198). (B) A crater-floor deposit that leaves the central peak exposed and therefore does not satisfy the crater-floor coverage criterion required for classification as a CCF (42° N, 16.69° E; CTX image: J03_045917_2223).

The criterion also excludes deposits that may eventually evolve into CCFs but presently only partially occupy crater interiors. In some cases, these deposits are recognisable as other VFF subtypes, such as LDAs which flow into craters and extend partway across the floor without fully infilling the basin (Figure 3.4A). In other cases, crater-confined lobate flows and tongues descend from the walls and spread across limited sectors of the floor. These wall-slope accumulations are widely documented across Mars' mid-latitudes (e.g., Dickson et al., 2012; Hartmann et al., 2014; Sinha and Vijayan, 2017; Figure 2.24) and can superficially resemble embryonic CCFs but differ from them morphologically and developmentally. Crater chronologies and model results suggest that wall lobes are typically young (~10–100 Ma) (Pearce et al., 2011; Sinha and Vijayan, 2017) and, if significantly older, would likely have progressively infilled more of the crater floor, perhaps developing into CCFs (Weitz et al., 2018). Wall-slope lobes also differ from CCFs in planform and flow extent, typically extending only a few kilometres downslope before terminating, rather than forming the basin-wide, concentric infill characteristic of CCFs (Sinha and Vijayan, 2017). Consequently, while both inflowing LDAs and crater-confined wall lobes may record processes relevant to CCF formation, neither satisfies the geomorphic criteria for CCF classification in this study.

3.2.2. Tracelines

The second essential criterion for CCF identification is the presence of tracelines (Figure 3.5). Tracelines are common on VFF surfaces and are diverse in form with variations in morphology and planform patterns (e.g., Milliken et al., 2003; Head et al., 2010). On CCFs, they typically manifest as repeating sub-parallel formations of elongated ridges and troughs (with widths on the order 10s of metres wide) that are often oriented quasi parallel (Figure 3.5B) and occasionally transverse to the crater

wall (see Section 3.3.2, Figure 3.18). The prominence of tracelines can vary across individual CCFs and different CCFs, reflecting variations in the relief of the ridges (Figure 3.5B-D). While concentric crater-parallel lineations dominate, some CCFs also exhibit differently oriented or locally complex traceline arrangements.

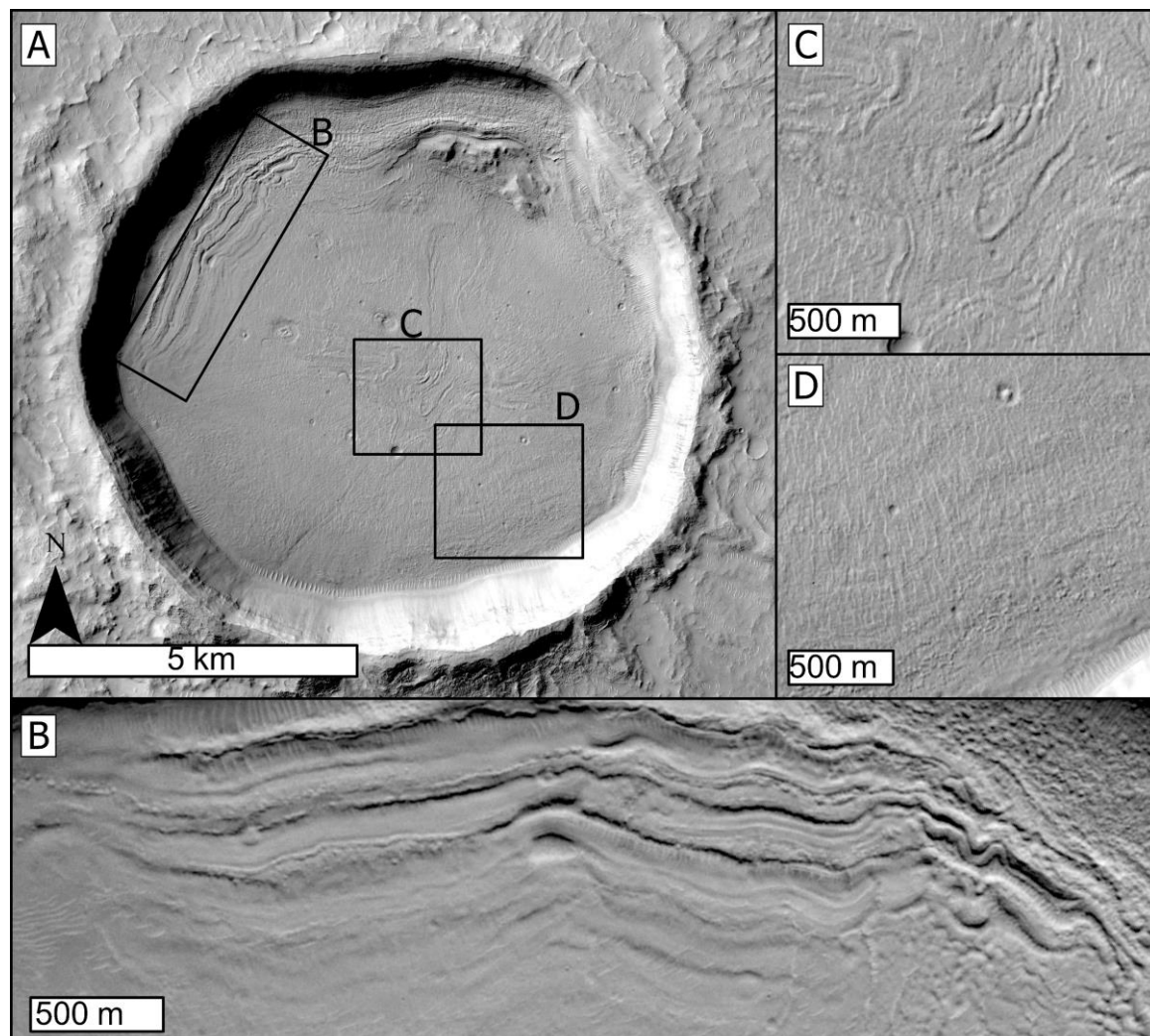


Figure 3.5. CCF typical traceline expressions and their variations. (A) A CCF located at 34.08°S 63.18°W with the extents of panels B-D shown. (B) illustrates the prominent parallel and repeating elongated ridges and troughs on the outer portions of the CCF. (C) displays an example of a significantly more subdued traceline pattern. (D) demonstrates a more intricate traceline variation, characterised by non-repeating, non-parallel, and complex arrangements of ridges and troughs. CTX image: B16_015921_1442.

CTX images (~5–6 m/pixel) formed the sole basis for global classification. HiRISE images (25 cm/pixel) and DTMs were consulted where available for detailed observation of surface morphology and confirm that CTX-scale tracelines correspond to real ridge–trough topography. Accordingly, while HiRISE provides critical validation of surface morphology, the diagnostic expression of tracelines—as observed consistently at CTX scale—forms the primary basis for identification. The key characteristics observable at each scale are summarised below.

Tracelines on CTX images: At CTX resolution ($\sim 5\text{--}6$ m/pixel), the elongated ridges and troughs forming tracelines often appear continuous over long distances. In classic examples, particularly on CCFs in large craters, individual tracelines along the outer margin of the CCF surface have been found to extend up to ~ 4 km in length and ~ 30 m in width (Figure 3.4; see also Figure 3.1B). On smaller CCFs, these lengths are proportionally reduced but still commonly exceed hundreds of metres. Close inspections reveal that where tracelines terminate, they often do so abruptly; in some cases, they are interrupted by more prominent ridges, while in others they fragment into more irregular, discontinuous forms.

In terms of traceline patterning across CCFs, circular loop arrangements are most common; however, morphologies vary considerably. Individual tracelines may adopt quasi-circular, polygonal, or cusped planforms, and patterns can be off-centred within the crater, producing asymmetry with compressed sectors opposite elongated ones. Overlapping or nested loops also occur (see Section 3.3). This diversity indicates that while circular concentricity is characteristic of CCFs, it is not universal.

Tracelines on HiRISE images: On these high-resolution data (25 cm/pixel), tracelines frequently look more complex (than on CTX images), exhibiting local deflections, and variable spacing. Elevation profiles extracted from the HiRISE DTM along transect A–A' (Figure 3.6) reveal two superimposed scales of relief. At the broader scale, the CCF surface forms a gentle, long-wavelength slope across the crater floor. Superimposed on this gradient are shorter-wavelength undulations corresponding to individual tracelines. The most prominent ridges display crest-to-trough amplitudes of $\sim 3\text{--}4$ m with wavelengths of $\sim 20\text{--}30$ m, and ridge widths of 2–3 m narrowing slightly at troughs. More subdued examples show relief of ~ 1 m and somewhat shorter wavelengths ($\sim 15\text{--}25$ m). These data confirm that CTX traces represent genuine geomorphic structure and show that some fine-scale curvature or segmentation visible in HiRISE will appear simplified or smoothed in CTX. Thus, HiRISE informs interpretation but does not control classification.

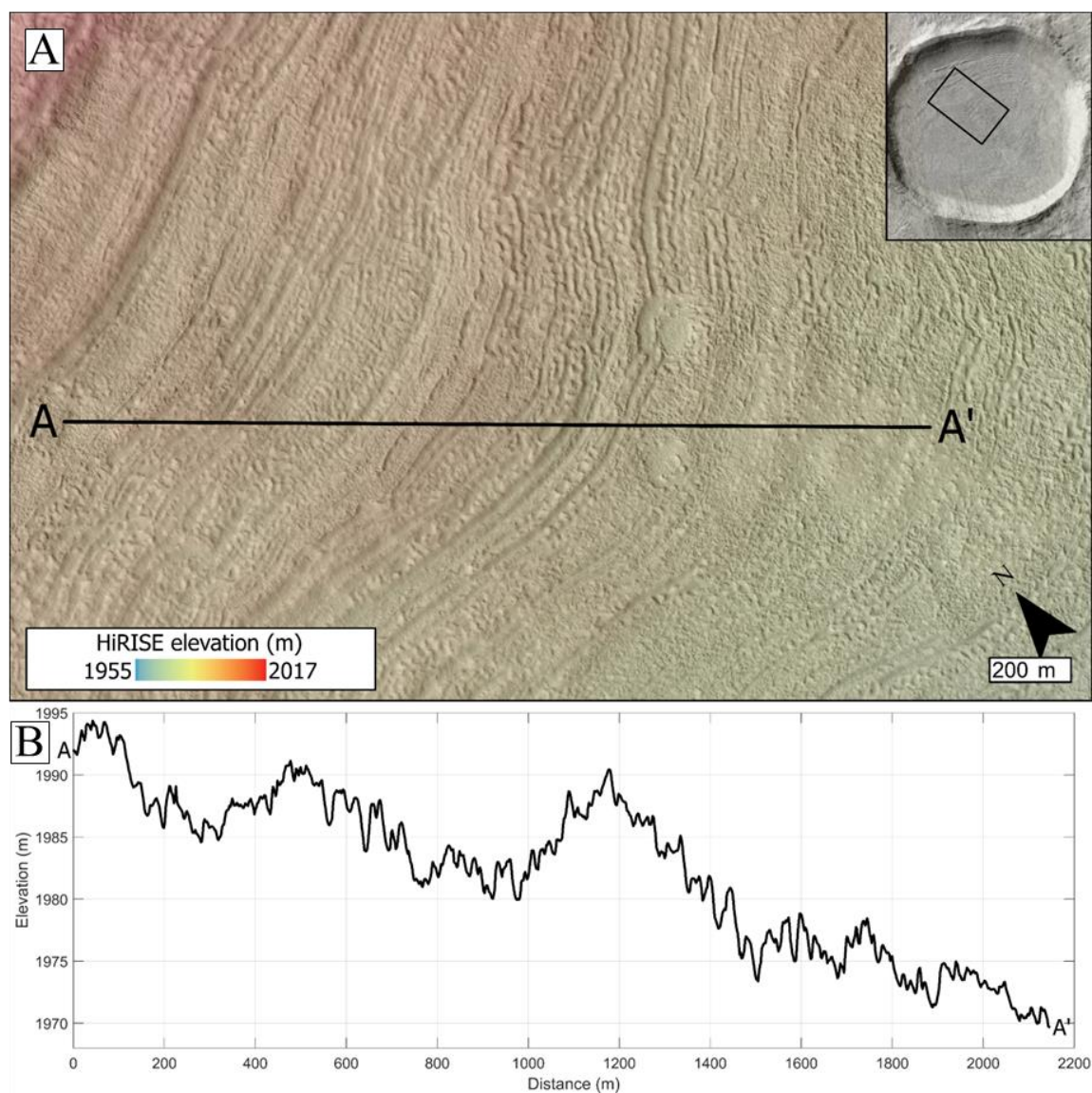


Figure 3.6. HiRISE image and elevation transect across CCF tracelines. (A) CCF tracelines and elevation transect overlaid on a HiRISE Digital Terrain Model (DTM) (43.17° S, 140.18° W; HiRISE image: ESP_046622_1365; HiRISE DTM: DTEEC_046622_1365_046688_1365_A01). (B) Elevation transect along path A–A' from panel A. The profile reveals two superimposed scales of relief, with longer-wavelength undulations underlying the finer-scale tracelines.

It is important to note that CCF surface expression may be partially obscured by overlying material, most notably the Latitude-Dependent Mantle (LDM). Because the LDM overlaps spatially with the CCF population, particularly at the poleward limits of their distribution, tracelines may be locally muted or undetectable (e.g., Figure 3.3E).

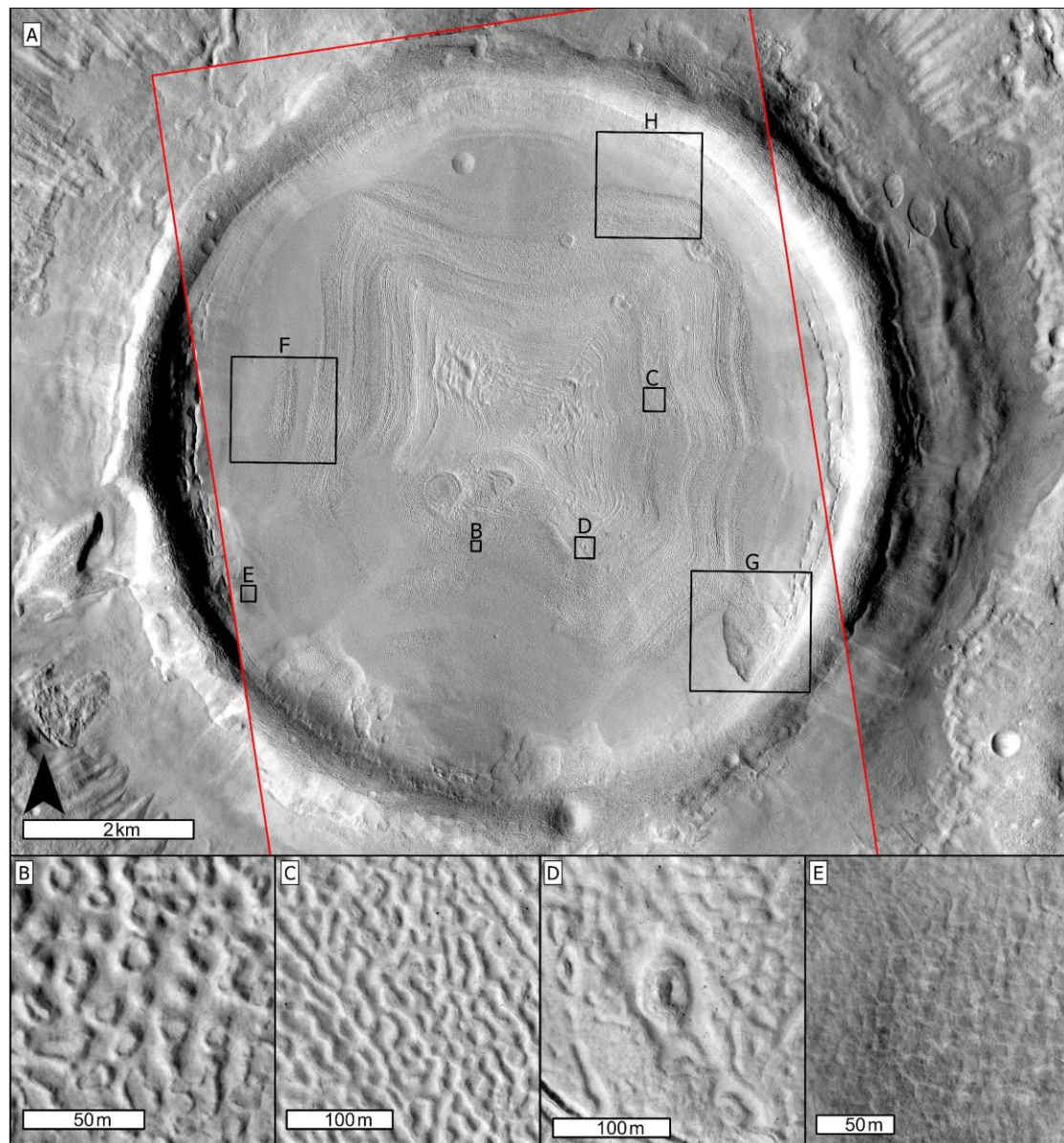
Where tracelines were visible—even if only in short or discontinuous segments due to mantling—they were taken as sufficient evidence that the underlying CCF is still present and the feature was classified as a CCF. Where no clear tracelines were visible but supplementary viscous-flow morphologies were present, the feature was recorded as an uncertain CCF on the basis that mantling may have obscured

original surface expression (e.g., Figure 3.3E). Conversely, in instances where the LDM completely drapes over the crater-filling unit such that neither tracelines nor supplementary VFF morphologies are identifiable, the feature is classified as a non-CCF (e.g., Figure 3.3G). Thus, partial mantling does not preclude identification as a CCF, but complete surface masking prevents confident attribution.

3.2.3. Other landforms observed on Viscous Flow Features

Besides tracelines, several other landforms and morphological characteristics commonly occur on the surfaces of VFFs that indicate the presence of ice within them (Figure 3.7). These include (a) brain terrain (e.g., Levy et al., 2009) (Figure 3.7B & C), (b) ring-mold craters (e.g., Kress and Head, 2008) (Figure 3.7D), and (c) small-scale pits/pitting (e.g., Soare et al., 2017) (Figure 3.7E). During my classification workflow, where a candidate CCF covers the host crater floor but its traceline identification is uncertain, I use these landforms (where possible) to distinguish between ‘uncertain CCF’ (Figure 3.3D–F) and ‘non CCF’ (Figure 3.3G–J). Specifically, if the deposit fully covers the crater floor but displays ambiguous or no tracelines, the presence of supplementary VFF morphologies (e.g., brain terrain, ring-mold craters, or pitting) results in an ‘uncertain CCF’ designation. Conversely, if tracelines are absent or ambiguous *and* the deposit does not fully cover the crater floor, the feature is classified as ‘non-CCF’. This procedure reflects the primacy of crater-floor coverage as a defining criterion. The following sections briefly describe these supplementary VFF morphologies and surficial deposits, providing the reader with the necessary context for their recognition and use in cases where tracelines are ambiguous.

CCFs exhibit various **brain terrain** morphologies, which comprise ridges arranged in cell-like patterns and are thought to be formed by periglacial processes modifying the surface debris layer, which may itself host ground ice (e.g., Noe Dobrea et al., 2007, Levy et al., 2009). These features are frequently observed on VFFs. Whilst brain terrain can be identified in CTX images, their detailed morphology is best resolved with HiRISE, revealing two principal forms of brain terrain distinguished by cell structure (Figure 3.7B and C). "Closed cells" (Figure 3.7C), comprise of arcuate or domed mound, typically measuring 10–20 metres in width, ~10–100 m in length, and ~4–5 m in height, (Levy et al., 2009). "Open cells" (Figure 3.7B), by contrast are narrower (4–6 m) and lower (2 m), characterised by convex-up ridges enclosing flat-floored depressions (Levy et al., 2009). Both cell types resemble terrestrial sorted stone circles in scale (Kessler et al. 2001; Kääb et al. 2014; Pearson et al., 2024).



Common morphologies of latitude-dependent mantle

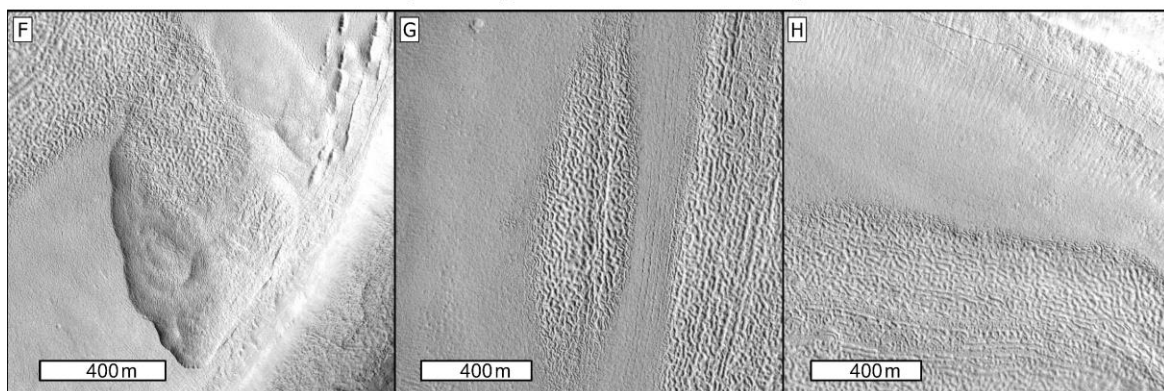


Figure 3.7. Examples of VFF surface morphologies and superposed LDM on a CCF. (A) CCF at 40.81° N, 99.48° E, showing the locations of panels B–H (black outlines). HiRISE image PSP_002175_2210 (red outline) overlain on CTX image P03_002175_2211. (B–C) Open-cell (B) and closed-cell (C) variants of brain terrain. (D) Ring-mold crater. (E) Pitting and polygonal textures. (F) Scalloped depression, typically oval in planform and associated with irregular pits and steep slopes, locally

Ring-mold Craters (RMCs) (Figure 3.7D) are interpreted to be the result of impacts into a relatively pure ice substrate below a thin regolith and differential sublimation into a subsurface glacial deposit (Mangold, 2003, Kress and Head, 2008). RMCs are a small-scale feature often less than a few hundred metres in diameter and are identifiable by their unusual morphology, with a raised circular outer border or moat extending around a central interior, similar to a truncated torus. The central floors of RMCs have been identified to display a range of interior morphologies including central plateaus, pits, bowls or mounds (Kress and Head, 2008).

Additional **pitted textures**, at the ten-metre scale and often accompanied by knobby terrain or networks of small buttes, may also occur on VFF surfaces (Mustard et al., 2001; Figure 3.7E). Such pitting has been previously associated with uneven heterogeneous sublimation of near-surface ground ice within and/or beneath the debris layer creating an uneven and bumpy surface (Soare et al., 2017), and thus serves as a potential indicator of ice-rich VFFs. Similarly, **Scallops**—oval or scalloped depressions that may appear individually or coalesce to form extensive knobby terrains (Figure 3.7G)—are interpreted as the result of erosional processes driven by sublimation of ground ice (Plescia, 2003; Lefort et al., 2010).

Among the most widespread surficial deposits on CCFs are mantling layers, which can obscure the underlying morphology and complicate identification of tracelines or other features. **Latitude-Dependent Mantle (LDM)** is a smooth, ice–dust layer that drapes pre-existing topography and is extensive above $\sim 45^\circ$ latitude in both hemispheres (Kreslavsky and Head, 2002; Head et al., 2003), while **Dark Aeolian Materials**—ubiquitous across Mars—cover CCF surfaces and crater walls (Figure 3.4A–B; Tirsch et al., 2011). Both deposits act as superficial mantling layers: LDM typically conceals concentric tracelines or smaller-scale VFF morphologies (Figure 3.7F–H), whereas the dark aeolian material can similarly mask surface textures and may reflect exposure of globally distributed subsurface material. Collectively, these mantling layers represent a significant modifying influence on CCF surfaces, with implications for both morphological interpretation and the detection of ice-related features. In some instances, these mantling layers completely drape over the crater-filling unit, preventing CCF classification, but frequently they are confined to either the outer margins or central portions of the CCF, leaving tracelines only partially obscured.

3.2.4. Low-definition CCFs (uncertain and non CCFs)

During the examination of the Levy et al. (2014) inventory, many features were identified that did not meet the strict ‘high-confidence’ classification and were subsequently excluded from the primary subset of this Thesis (see examples in Figure 3.3D–I). These excluded features correspond to what has been termed ‘low-definition’ CCFs in the literature (Levy et al., 2010; Figure 3.8), or as debris-covered lobate floor fill deposits resulting from downslope flow from crater walls onto crater floors (Head et al., 2008). These ‘low-definition’ CCFs warrant brief discussion.

Some of these deposits cover the host-crater floor (Figure 3.3D–G) but are defined by an absence of visible traceline patterning, or it is obscured by degradation and mantling. Accordingly, their identification as CCFs remains uncertain. Despite this uncertainty, these deposits likely represent significant reservoirs of buried ice and hold value for reconstructing Martian glacial history. Consequently, several previous studies have included these low-definition cases to provide a more comprehensive census of potential crater-confined mid-latitude ice (e.g., Dickson et al., 2012; Levy et al., 2014). However, for the purposes of this Thesis, these features are excluded to ensure that subsequent analysis is limited to morphologically unambiguous examples of crater-confined glacial ice. If they are true CCFs (contain ice), one could potentially investigate why they are so highly degraded relative to (the definite) CCFs, but this matter lies beyond the scope of this Thesis.

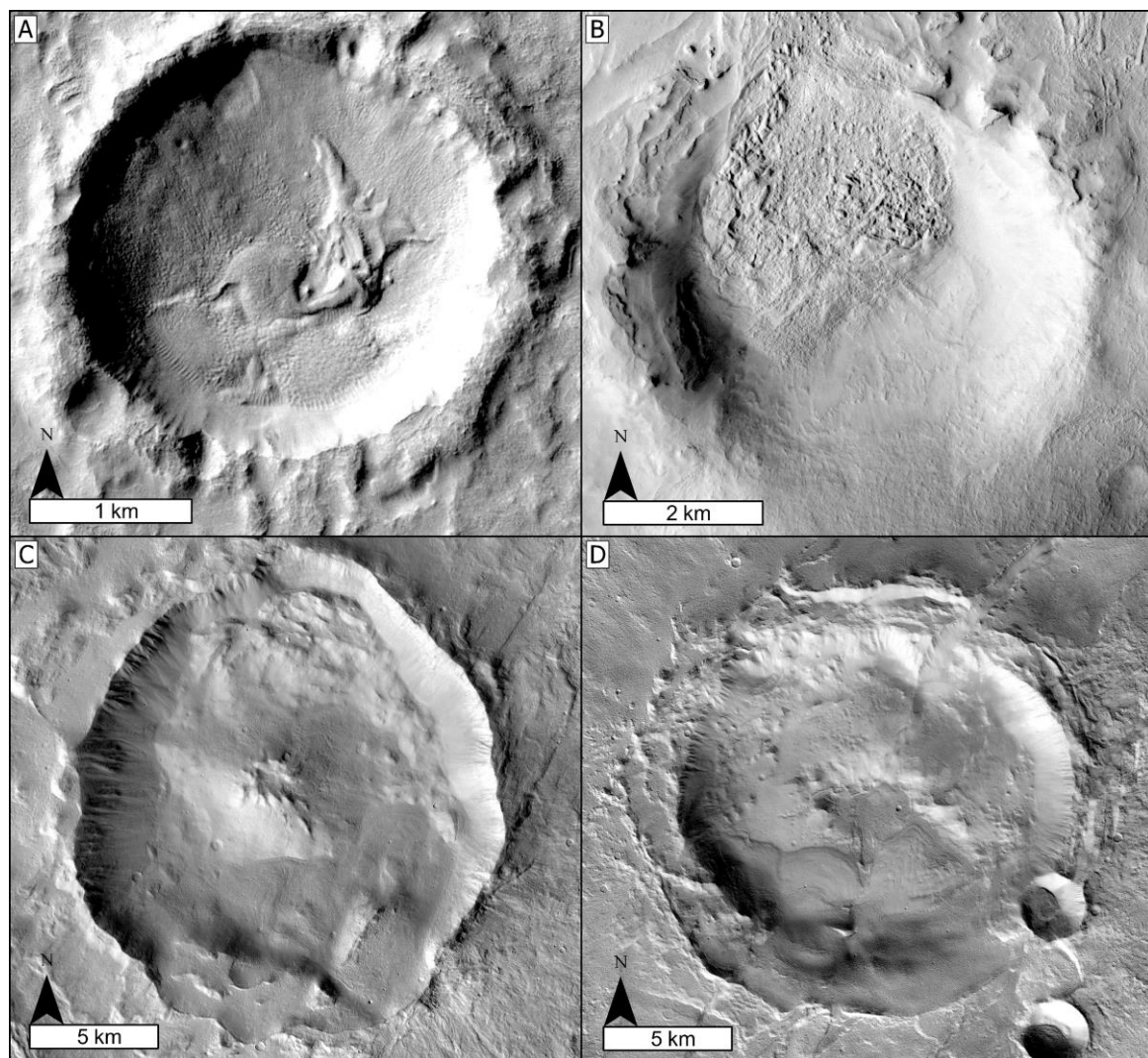


Figure 3.8. Examples of low-definition CCFs. Shown are four deposits from the Levy et al. (2014) inventory previously classified as low-definition CCFs (Levy et al., 2010) but excluded from the high-confidence CCF inventory used in this thesis due to the absence of diagnostic traceline morphologies and/or ambiguous crater-floor coverage. All panels are CTX images: (A) 37.26° S, 26.28° W; CTX J03_046064_1430. (B) 33.14° N, 60.38° E; CTX P16_007385_2149. (C) 45.82° N, 71.06° W; CTX D01_027538_2262. (D) 43.46° N, 113.21° W; CTX G14_023821_2238.

3.3 CCF Variation: classic vs. non-classic CCFs

After applying the classification criteria of Section 3.2, the 9,533 entries in the Levy et al. (2014) inventory were filtered to retain all “definite CCFs”. This refinement yielded 3,028 CCFs for my Thesis investigation—1,579 in the northern hemisphere and 1,449 in the southern hemisphere. Figure 3.9 plots their global distribution. Their spatial trends and latitudinal variations will be examined more in Chapter 4.

The 3,028 CCFs exhibit considerable morphological diversity, particularly in the form and arrangement of traceline patterning. The traceline expressions allowed me to separate them into two principal morphological groupings, which I call (i) “classic” CCFs (Section 3.3.1), which exhibit clear, well-preserved systems of tracelines, and (ii) “non-classic” CCFs (Section 3.3.2), where tracelines remain visible beneath later mantling or surface modification. The next subsections elaborate on these types. Classic CCFs will feature prominently in the work of some of the later chapters. As described below, both classic and non-classic CCFs show recognisable surface morphological variations that may be further sub-categorised (encompassing partial mantling and different kinds of degradation and morphological modification); however, in the following sections, I aim only to introduce the reader to the leading sub-categories, rather than analysing all variations comprehensively. The grouping into two major types is useful for structuring the analyses of the later chapters in terms of the CCFs’ appearance, without committing to any hypotheses about their origins.

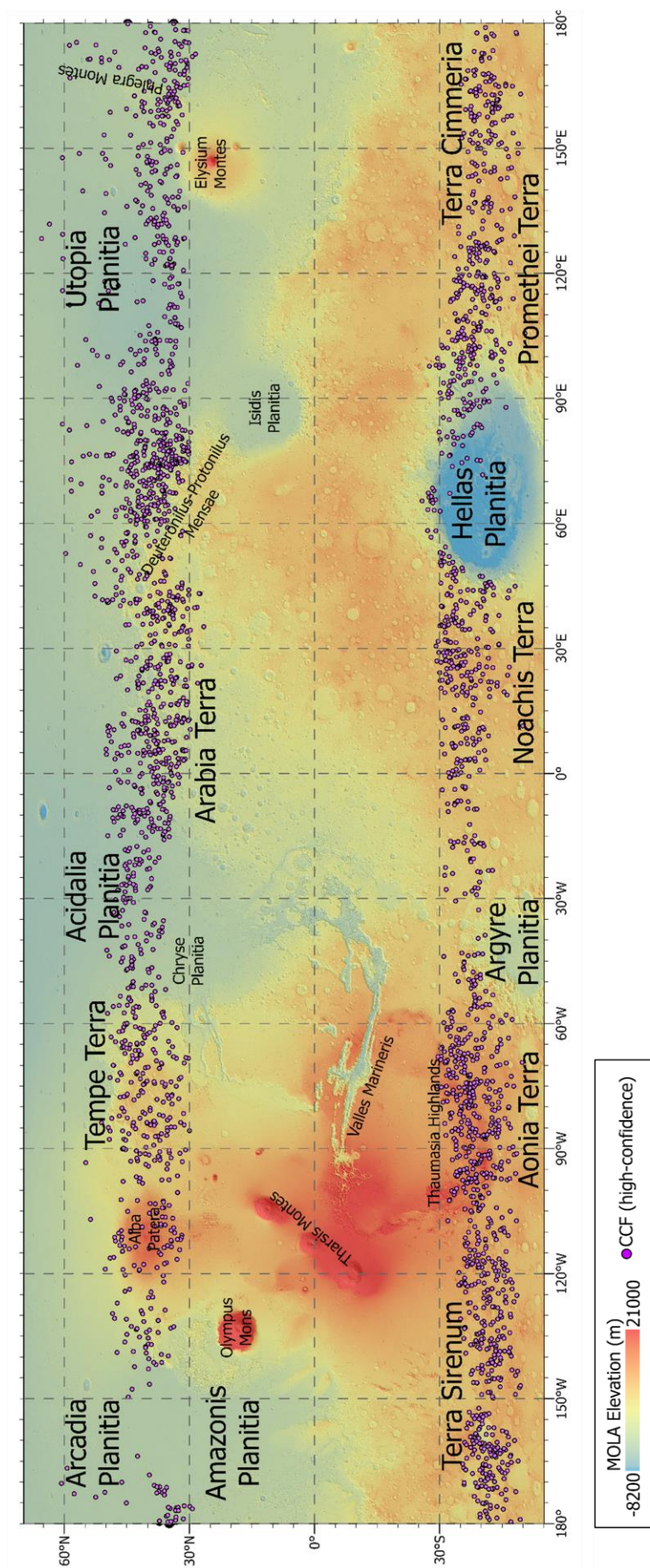


Figure 3.9. Global distribution of high-confidence CCFs. Map showing the locations of all high-confidence CCFs overlaid on gridded MOLA topography and MOLA-derived hillshade. Major physiographic regions are annotated for geographic reference.

3.3.1. Classic CCFs

In this Thesis, ‘Classic’ CCFs represent the most well-preserved examples of concentric crater fill. They are defined by clear, generally uninterrupted tracelines that remain sharply expressed and unmodified by superimposed mantling deposits and are relatively rare. I determined 122 southern and 168 northern examples meeting the definition, representing less than 10% of my CCF sample (3,028). Tracelines on classic CCFs typically display a distinctive clarity, with internal concentricity, often with broadly symmetrical, closed-loop patterns that are easy to trace along the crater’s interior margin, particularly near the wall-to-CCF surface transition. Whilst in larger host craters, the central portions of a classic CCF surface may exhibit minor modification, it is the integrity of the outer, wall-adjacent tracelines that defines these features as classic.

Classic CCFs (Figure 3.10) thus provide ideal references for identifying and comparing traceline morphology across the broader population. From my extensive observations through both CTX and HiRISE images, the traceline morphologies on classic CCFs can be categorised broadly by three levels of complexity: simple, intermediate, and complex. As crater size increases, the tracelines structure tends to become more complex.

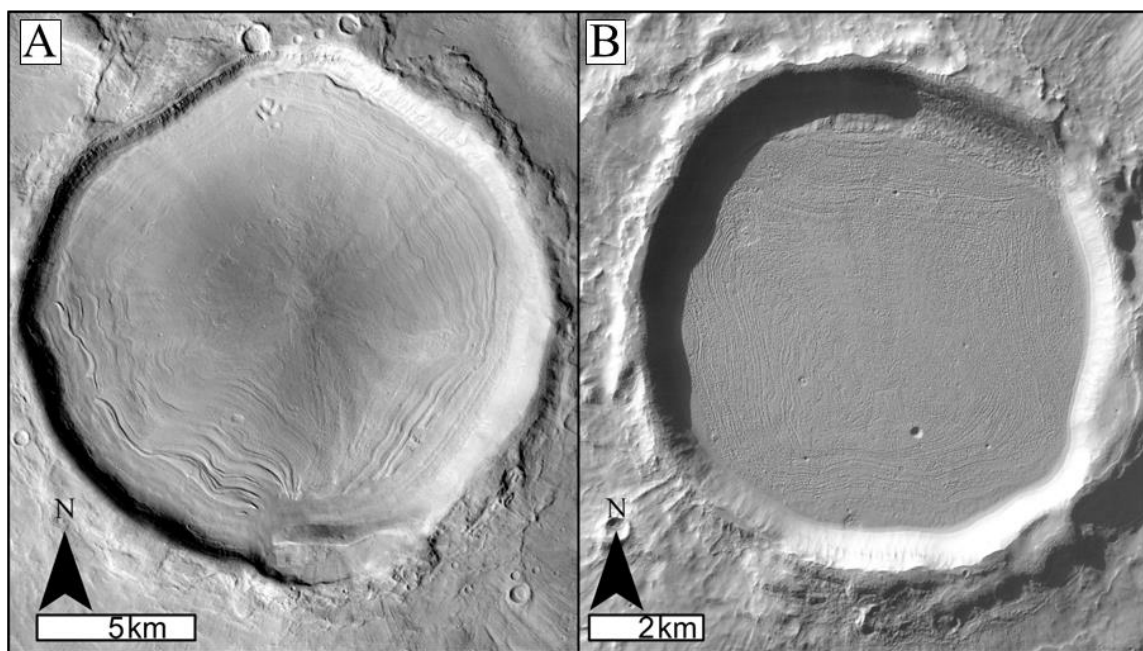


Figure 3.10. Two examples of classic CCF surface morphology. Despite differences in the visual expression of their traceline manifestations, both examples display clear internal concentricity, with broadly symmetrical, closed-loop tracelines characteristic of well-preserved, morphologically “classic” CCFs. (A) 36.41°N, 87.82°E; CTX image P15_007028_2164 and (B) 37.33 S, 76.11 W; CTX image P18_008155_1411.

Classic CCFs with ‘simple traceline morphologies’ are typically hosted by smaller craters, often less than 7 km in diameter. Simple traceline patterning are often circular or elliptical in form and exhibit uniform, repeating traceline patterns that form concentric loops. In many cases, the loops are circular

and centred on the host crater midpoint (Figure 3.11A). In others, the concentric patterning is slightly offset from the centre, resulting in oval or elliptical shapes that appear elongated in the direction of the offset (Figure 3.11B). Typically, tracelines become more compact along the major axis of the elliptical pattern, with wider spacing occurring along the minor axis. This off-centre geometry introduces asymmetry while maintaining an overall regular, low-complexity pattern.

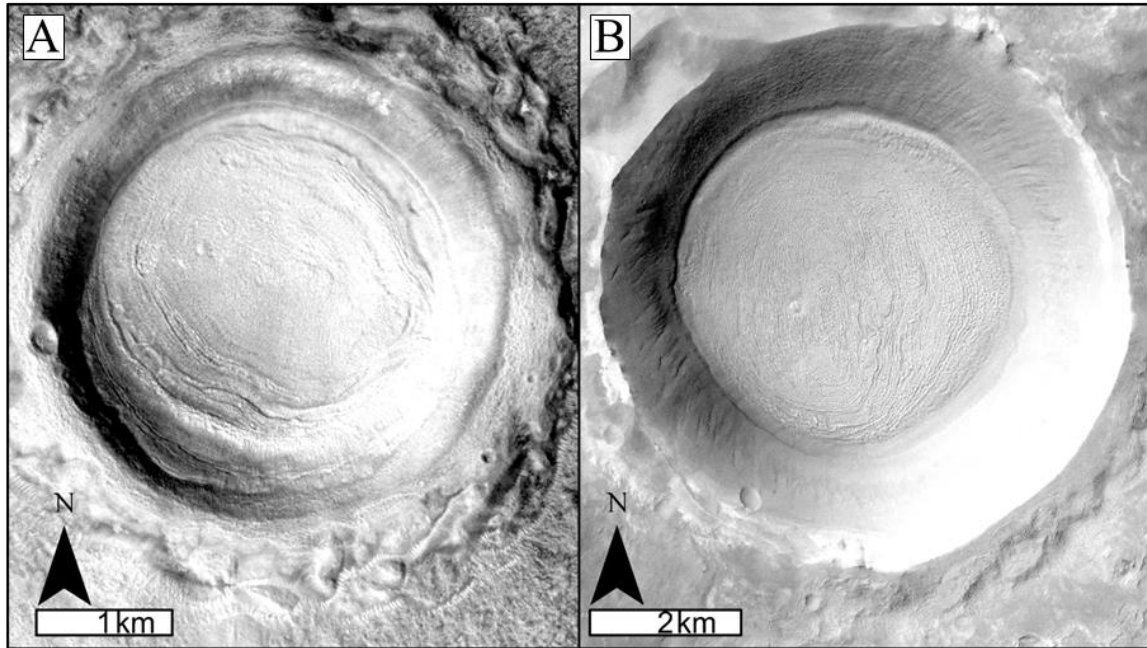


Figure 3.11. Examples of CCFs with simple traceline morphologies. (A) the tracelines form near-circular, centrally focused concentric loops (35.1° N, 95.81° E; CTX image P14_006619_2151). (B) the concentric pattern is slightly offset from the crater centre, producing an elliptical geometry with tracelines elongated along a preferred axis, while retaining a regular, repeating, low-complexity form (-34.79°S, 2.78°E; CTX image: P12_005647_1443)

Intermediate traceline morphologies are characterised by more angular, polygonal forms that depart from the circularity of simpler patterns. These patterns frequently appear as quasi-rectangular or irregular ‘geometric shape’, often exhibiting four to five distinct ‘sides’ (Figure 3.12). In many cases, these sides are sharply defined and exhibit linear or gently curving segments that contrast with the smoother arcs of simpler concentric fills. These can appear unbroken in CTX scale, often forming concentric evenly spaced loops. CCFs with intermediate traceline patterns are typically hosted by craters between ~7 and 15 km in diameter. While still broadly concentric, the tracelines show a noticeable gradient in form: outer tracelines tend to form adjacent or nearly parallel to sections of the crater wall, forming sharp-edged, almost linear structures. As the pattern progresses inward, the tracelines become increasingly curved and less angular, transitioning into smoother, more quasi-circular loops toward the crater centre. This inward softening of geometry creates a hybrid morphology that bridges the gap between simple and complex pattern types.

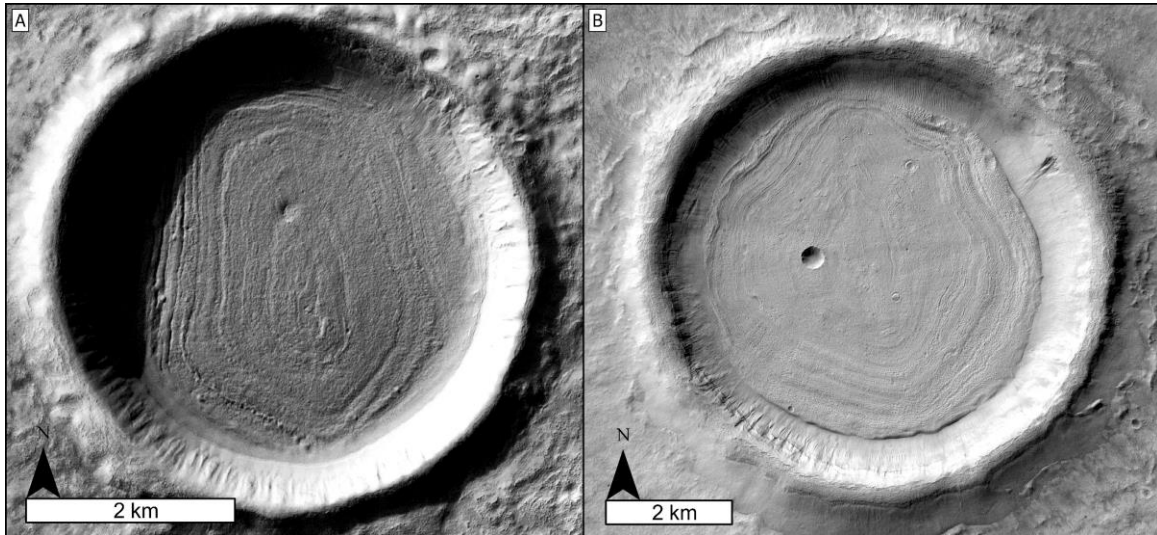


Figure 3.12. Examples of intermediate CCF traceline morphologies. (A) CCF at 35.91° S, 87.41° E; CTX B19_017155_1441. (B) CCF at 39.64° S, 123.07° E; CTX G17_025039_1401. Both examples show CCFs whose surface tracelines form irregular geometric shapes, commonly exhibiting four to five distinct “sides”, reflecting an intermediate level of morphological organisation and complexity.

Complex traceline morphologies are most commonly observed in CCF-hosting craters exceeding 15 km in diameter (Figure 3.13). These patterns are distinguished by their intricate, often highly cusped appearance, featuring numerous, irregular, broadly curvilinear tracelines that appear closely spaced largely because larger craters accommodate more tracelines, rather than due to an inherently higher density. Rather than forming smooth loops or quasi-rectangular polygons with clear ‘sides’, the traceline patterns on these complex CCFs exhibit a high degree of segmentation and internal variation. The outer margins of the CCF surface may still show a degree of concentric organisation, but this pattern is made more complex by the irregular, and often highly cuspy form of the tracelines, which introduce considerable internal variability (Figure 3.13C-F). Unlike ‘simple’ or ‘intermediate’ pattern morphologies, complex traceline patterns in this outer margin layer of the CCF can appear broadly concentric, yet closer examination reveals pronounced irregularity and variability in their form and structure (Figure 3.13F). The concentric traceline pattern is primarily confined to the outer margins (~20–40%) of the CCF surface area, with the innermost regions displaying a smoother or more irregular surface texture. Although the centre of the CCF often lacks concentric repetition, it may still display faint or discontinuous tracelines. These are usually more irregularly distributed and lack the more ordered spacing seen at the margins, contributing to an overall more mottled appearance in the interior.

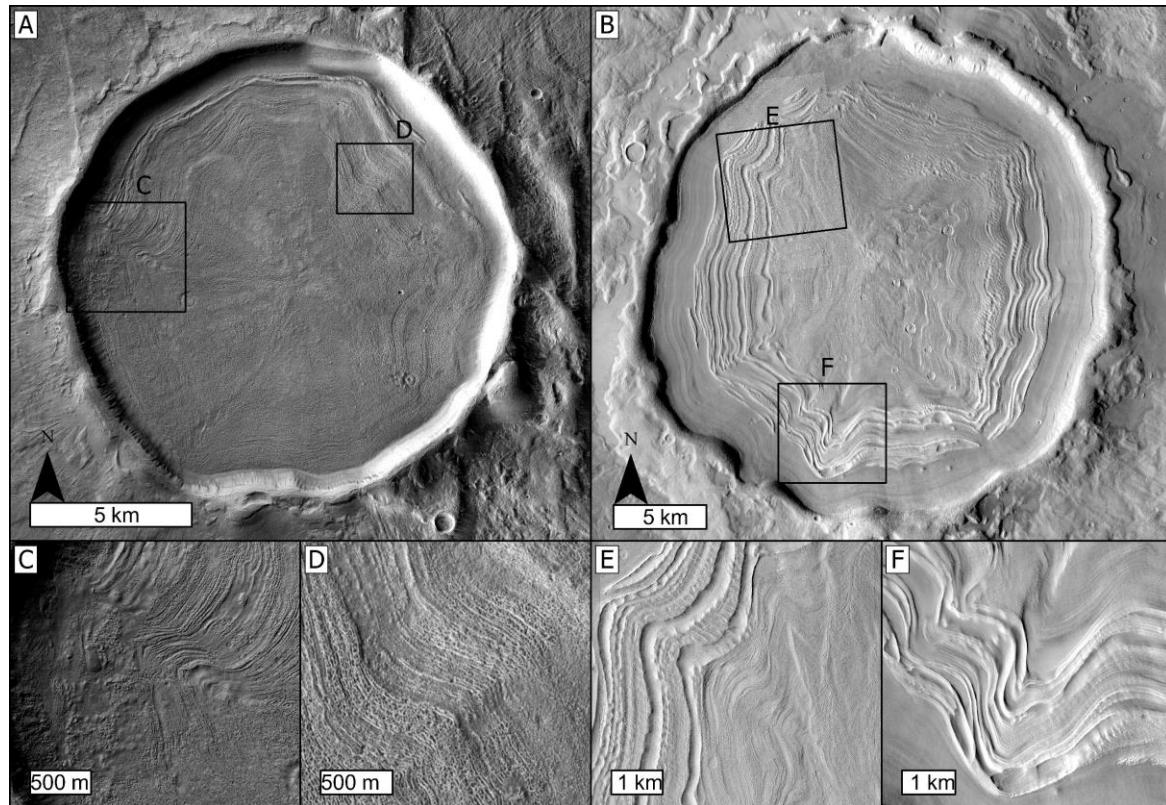


Figure 3.13. Examples of complex CCF traceline morphologies. (A) CCF showing a surface characterised by highly organised, repeating traceline patterns (44.14° S, 83.26° W; CTX image: P05_002947_1355). (B) CCF with more irregular, high-variance traceline morphology (37.89° N, 75.67° E; CTX images: P17_007714_2184 and B22_018144_2182, with HiRISE image: ESP_018144_2185). (C–D) Close-ups of (A) illustrate complex but internally consistent traceline spacing and curvature. (E–F) Close-ups of (B) illustrate strongly variable traceline spacing, width, and curvature, with frequent cusped and sharply pinched terminations. Both examples display complex traceline organisation, but (A) exhibits a more regular, rhythmically repeating surface structure, whereas (B) shows greater morphological variability and more strongly cusped, irregular deformation.

3.3.2. Non-classic CCF morphologies

Here, I describe and illustrate a spectrum of non-classic CCF traceline morphologies, illustrating to the reader their sheer diversity. The non-classic CCFs number 2738 in my collection. Due to the difficulty of establishing firm definitions/criteria for sub-categorising the morphologies, a sophisticated classification scheme for them is not attempted in this Thesis. However, future research might utilise some of the forms describe below to examine CCF history, notably with respect to questions about CCF accumulation and/or preservation histories.

The vast majority of non-classic CCFs exhibit some degree of surface obscuration, typically through mantling by LDM-like deposits, aeolian material, or other later surface processes. For some non-classic CCFs this smoothed mantle is ring shaped, superimposed on the outer edges or along one side of the central CCF (Figure 3.14, 3.15). This marginal mantle, which commonly hosts polygonal surface

morphologies at HiRISE scale, was first described in association with CCFs by Levy et al. (2009) as ‘polygonalized mantle’. It appears to extend from the crater wall and is typically between 0.6 and 1 km in width (see also Levy et al., 2009 for further discussion). Its morphology varies between craters: in some, it forms a relatively uniform band encircling the central CCF; in others, the mantle extent varies considerably, with certain sectors displaying broader, more extensive deposits while others are narrow.

In several examples, faint, partially-buried expressions of underlying CCF tracelines can be seen beneath the mantle surface (Figure 3.14), supporting the interpretation that this mantle represents a later emplacement phase superimposed on the original glacial surface. Where present, the interior of the CCF often appears slightly elevated relative to the surrounding mantle surface (Figure 3.15a). This subtle topographic contrast is below the resolution of MOLA but is detectable in 75 m/pixel HRSC DEMs (Neukum et al., 2004), with transects illustrating the relative elevation differences (Figure 3.15). In most cases, the marginal mantle forms a shallow gradient that slopes gently from the crater wall toward the CCF interior. Although commonly expressed as a continuous ring, partial mantling around crater margins is also observed.

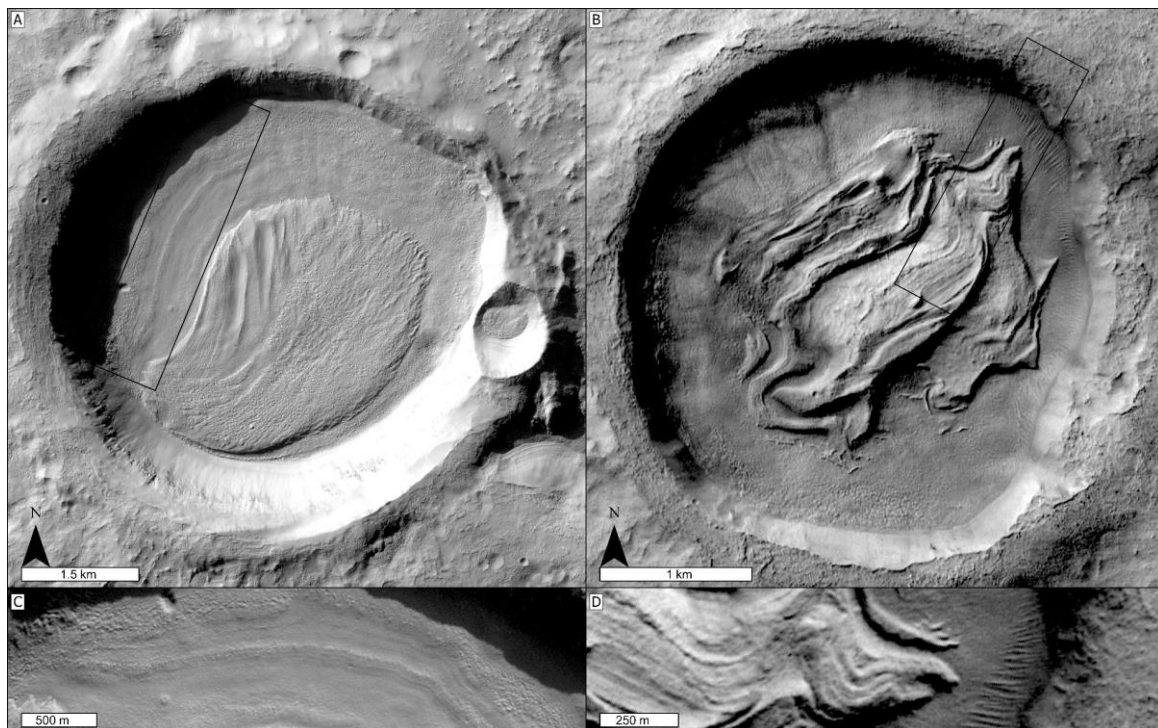


Figure 3.14. Examples of non-classic CCFs partially obscured by marginal mantling deposits. (A) showing a CCF in which a smooth mantling unit drapes one side of the central CCF deposit (32.91° S, 26.46° E; CTX image B19_016999_1460). (B) CCF surrounded by a ring-shaped marginal mantle developed along the outer edge of the deposit toward the crater wall (37.89° S, 18.15° E; HiRISE image H2694_0001). (C) Close-up of the mantled sector in (A), where faint, partially buried expressions of underlying CCF tracelines are still visible beneath the mantle surface. (D) Close-up of (B) illustrating the topographic relationship between the central CCF deposit, the marginal mantle, and the crater wall, with the CCF interior forming a subtle topographic high relative to the mantled zone.

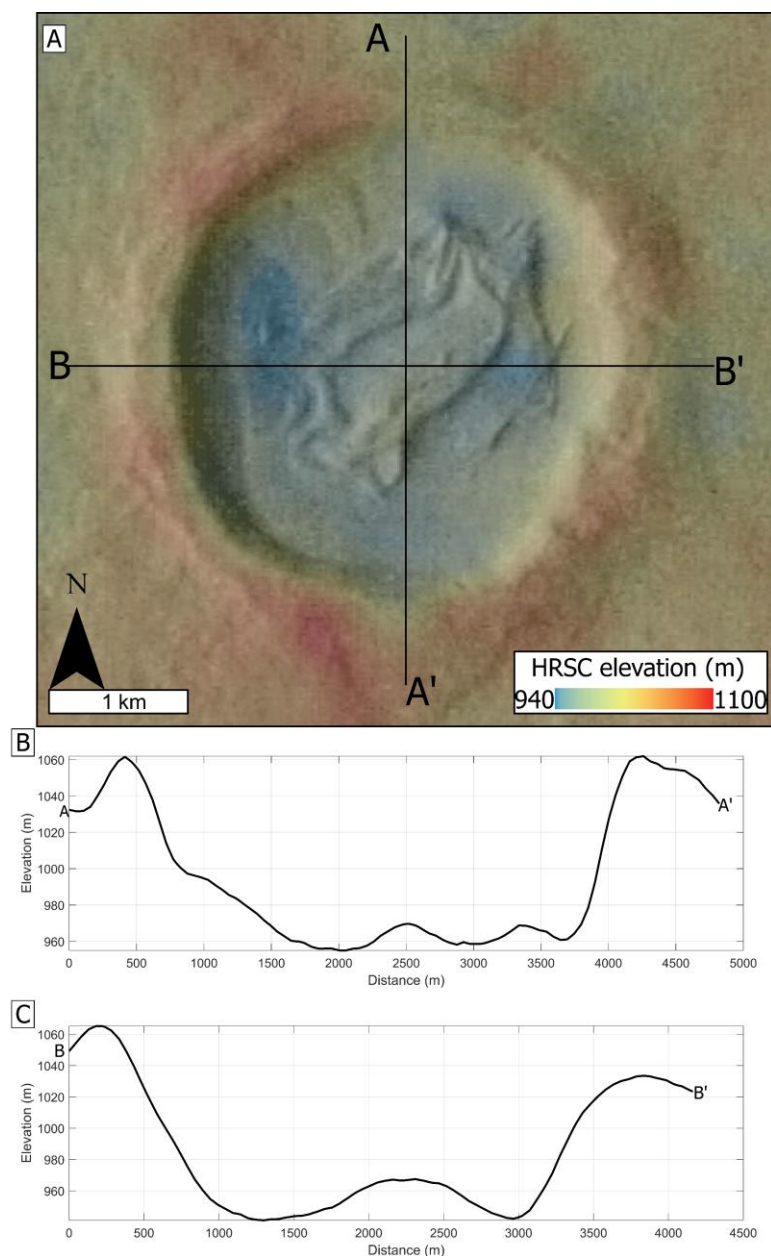


Figure 3.15. Topography and elevation transects of a ring-shaped marginal mantle surrounding a non-classic CCF. (A) HiRISE image H2694_0001 overlain on a HRSC DTM (H4097_0000_DA4) showing a CCF (37.89° S, 18.15; Figure 3.14B) encircled by a smooth marginal mantle developed along the outer edge of the deposit toward the crater wall. (B) North-south and (C) east-west elevation transects across the crater interior showing a distinct topographic depression at the outer margin of the central CCF deposit, followed by an increase in elevation toward the crater wall. The mantle forms a shallow, annular zone of lower elevation separating the central CCF from the crater wall and is typically ~400 m wide, although it broadens and becomes more gently sloping toward the northern sector of the crater where the mantle extends further upslope onto the crater wall.

In other instances, preferential degradation occurs at the centre of the non-classic CCF, with traceline structure usually restricted to the outer margins of the surface nearest the crater wall. The surfaces of these degraded central deposit ridges commonly exhibit a pattern of multiple curvilinear ridges, variably oriented, with some arcuate and others more linear (Figure 3.16). This suggests preserved internal layering/stratification, albeit in a degraded state. These ridge-and-furrow pattern and complex degraded topography have also previously been identified on debris tongue of LDAs (Van Gasselt et al., 2007). The central deposit often comprises irregular, ridge-forming units that exhibit distinct layering, interpreted as relict central mounds or exposed CCF-like materials.

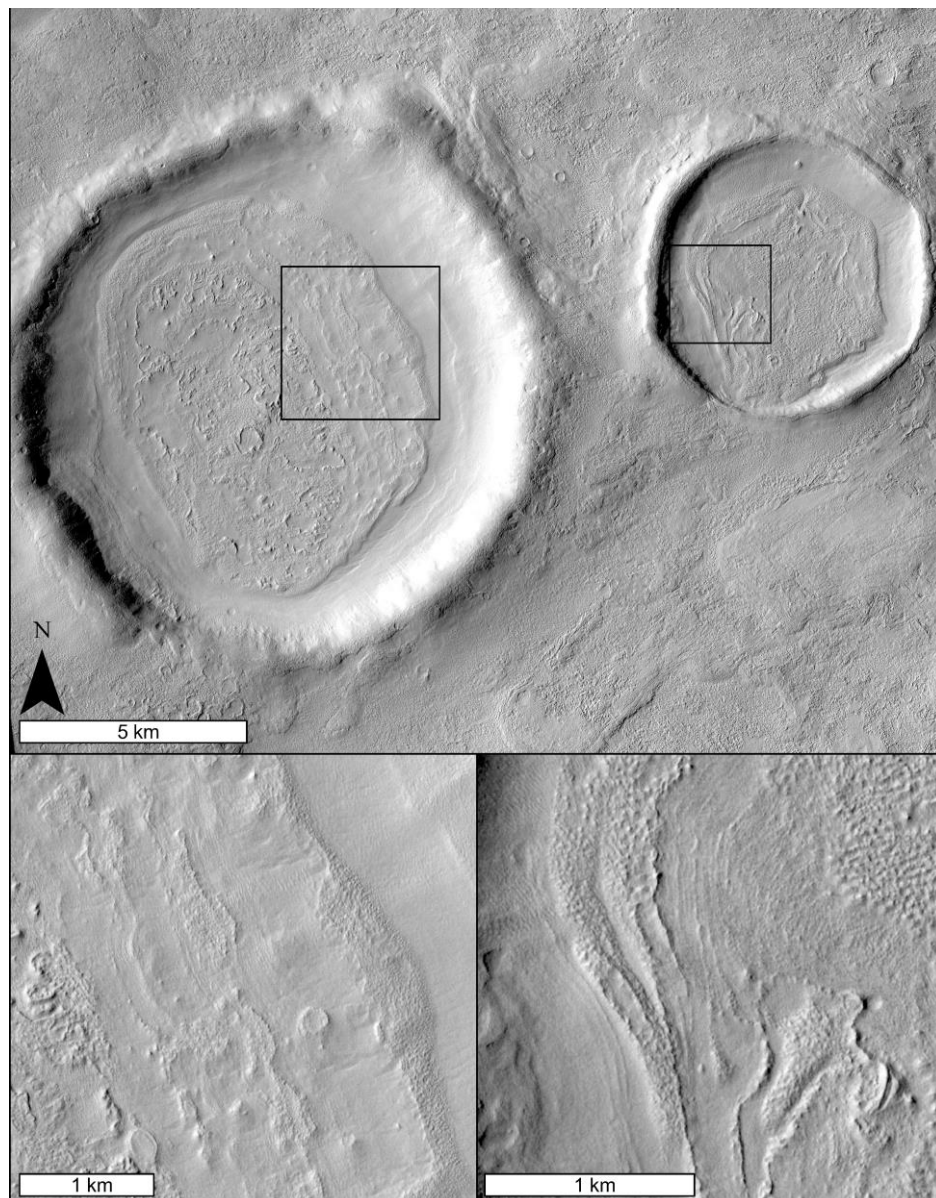


Figure 3.16. Examples of degraded non-classic CCFs. In both CCFs the central parts of the deposits are strongly modified. Both CCFs show central degradation and traceline expression preferentially preserved near the crater margins. Both CCFs also display marginal mantling, illustrating that individual non-classic CCFs commonly combine multiple modification pathways. (42.64° W, 114.25° E; CTX image: G06_020622_1385).

38 non-classic CCFs in the inventory display distinctive divergent traceline patterns (Figure 3.17), defined by curvilinear ridges that appear to emanate from one sector of the surface and radiate outward until terminating at the crater wall or an overlying mantle. Unlike the concentric loops typical of most CCFs, these tracelines form spreading, fan-like arrangements. Divergent examples are most common in the lower mid-latitudes ($\sim 30^\circ$ – 40° N and S) and are generally restricted to smaller craters, typically less than 7 km in diameter.

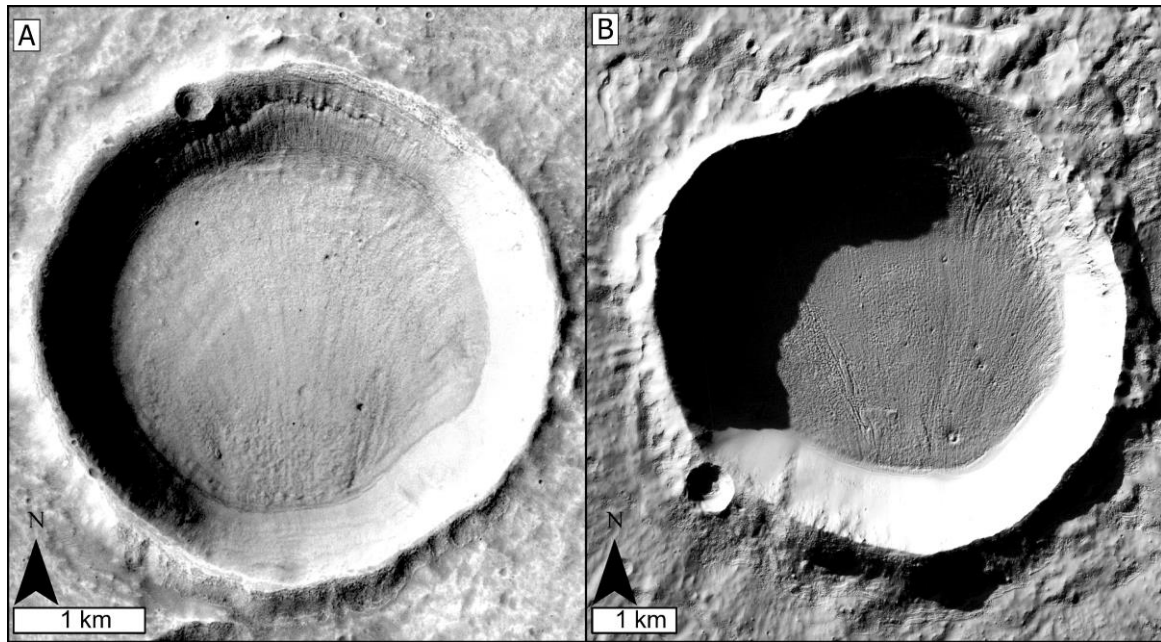


Figure 3.17. Examples of non-classic CCFs displaying divergent (fan-like) traceline patterns. In both cases, curvilinear ridges originate from the equator-facing crater wall and spread outward across the crater floor rather than forming concentric loops. (A) 30.96° S, 93.70° E; CTX image: P13_006237_1475. (B) 34.42° S, 0.19° W; CTX image: P21_009286_1430.

Morphologically, these cases resemble crater-confined LDA discussed in Section 3.2.1. Similar features have been described in previous literature as lobate deposits within craters, termed *pole-facing flow features* (Dickson et al., 2012; Sinha and Vijayan, 2017), interpreted as younger, small-scale ice flows that typically drape over portions of the crater wall and floor rather than completely filling the basin. In contrast, the divergent CCFs identified here extend fully across the crater floor, meeting the opposite wall and thus satisfying the criteria for CCFs (Table 3.1). Their surface morphology is visually analogous to terrestrial piedmont glaciers, where ice accumulates along a confined margin before spreading outward across flatter terrain. Interestingly, the divergent CCFs identified in Figure 3.17 are not pole facing and instead their tracelines are consistently more tightly packed at the equator-facing wall and diverge laterally as they extend toward the northern end of the crater. These examples are therefore included in the inventory on the basis of their crater-filling extent and likely icy origin, though their classification remains tentative.

A related morphological but notably rare subset (78 cases) comprises non-classic CCFs, which exhibit anomalous, linear surface morphologies (Figure 3.18) defined by linear, subparallel tracelines which cross the CCF surface. These stand in marked contrast to the concentric, looped traceline patterns characteristic of most (both classic and non-classic) CCFs. To the author's knowledge, linear CCFs have not been specifically discussed in prior literature.

Their divergence from the classic concentric morphology raises questions about their true origin. One possible explanation is that these features also represent remnants of LDA-like flow, confined within crater interiors. Similar processes were recently described by Wueller et al. (2025), where LDA development in confined environments results in material flowing downgradient in parallel lobes, with surface lineations oriented perpendicular to the direction of flow and parallel to lobe fronts. Although their traceline orientations differ from both classic and non-classic CCFs, these features satisfy the morphological criteria defined in Table 3.1 and are therefore included in the inventory.

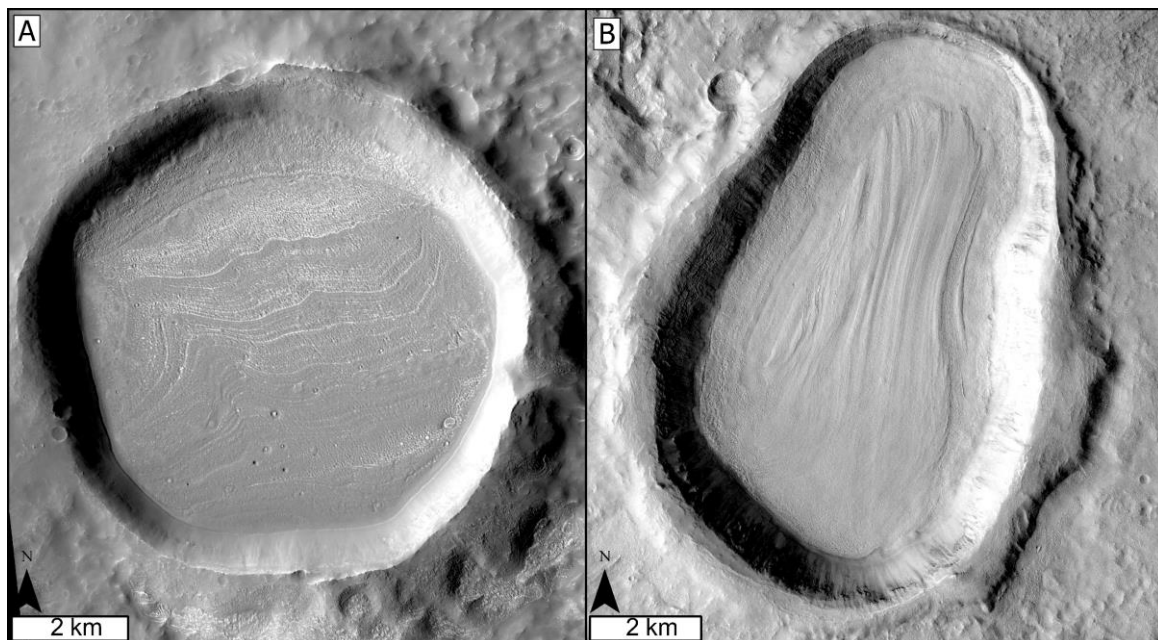


Figure 3.18. Examples of linear CCF surface morphologies characterised by transverse, sub-parallel traceline patterns. (A) Example of a linear CCF within a typical circular impact crater, where tracelines extend coherently across the CCF surface (30.35° S, 67.85° W; CTX image: B05_011596_1484). (B) Example of a linear CCF within an elongated (oval) host crater, in which tracelines are organised parallel to the crater's major axis, producing a strongly directional, north–south–oriented surface fabric (32.39° S, 26.91° E; CTX image: B08_012898_1476).

26 non-classic CCFs possess prominent, elevated outcrops of fractured terrain emerging from an otherwise smoother CCF surface (Figure 3.19). These outcrops, generally less than 2 km in diameter and hosted within craters typically under 6 km wide, are not interpreted as native to the crater basin. HiRISE images (Figure 3.19) confirm that they rise above the surrounding CCF surface—and in some cases even above the crater rim—indicating that they are superimposed rather than integral to the original crater floor. The outcrops typically exhibit heavily fractured, blocky textures, often with subtle

flow-like structures on one side and finer-scale linear textures (typically <100 m in length) on the other. They are rarely centrally located, instead occurring asymmetrically within the crater interior, frequently dominating one sector of the CCF and generating a distinct local contrast in surface morphology.

The origin and composition of these elevated blocks remain uncertain. They could represent remnants of moraine-like accumulations or stagnant ice blocks preserved beneath a sublimation lag, analogous to features observed at terrestrial glacial margins, though an alternative interpretation as exogenic ejecta material cannot be excluded. Notably, the surrounding concentric traceline patterns frequently follow the perimeter of these outcrops, appearing to conform to their margins. This alignment suggests that the development of surface patterning may have been influenced by the presence of the elevated material.

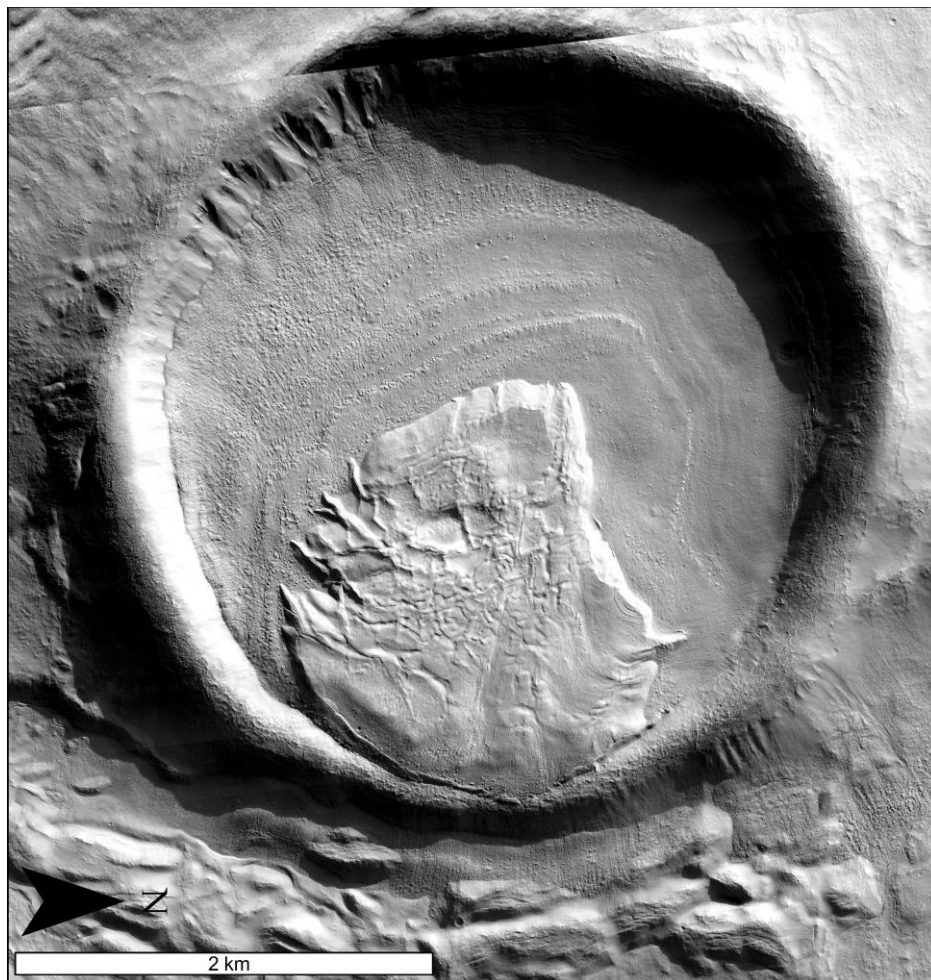


Figure 3.19. CCF with an outcrop rising above the crater rim. CCF containing a topographic outcrop that protrudes above the surrounding crater rim. Notably, the surrounding concentric traceline patterns follow the perimeter of this outcrop and appear to conform to its margins, indicating that the development of surface lineations was influenced by the presence of this elevated material (32.31° S, 50.39° E; HiRISE image: ESP_078265_1475; CTX image: G19_025701).

The majority of non-classic CCFs exhibit combinations of the characteristics described above, representing a continuum of modification states rather than distinct morphological types. Most display

some degree of mantling, surface degradation, or irregular traceline preservation, highlighting the progressive nature of post-emplacement alteration across the sample. Representative examples of these overlapping morphologies—encompassing degraded, mantled, and morphologically complex surfaces—are shown in Figure 3.20, illustrating only a small subset of the diversity expressed within the broader non-classic CCF population.

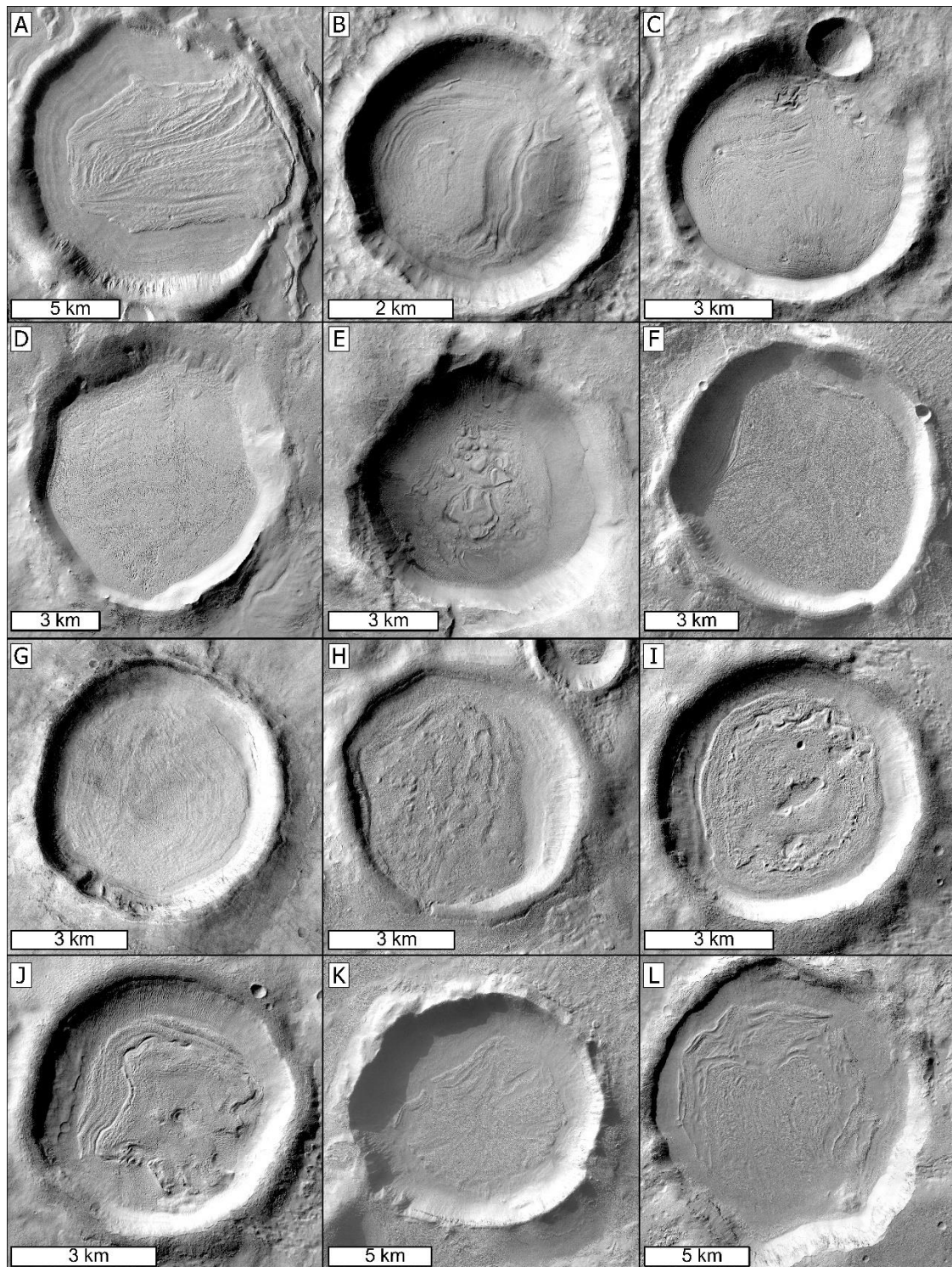


Figure 3.20. Representative examples of composite non-classic CCFs exhibiting mixed states of modification. All panels show atypical non-classic CCFs displaying various combinations of surface mantling, degradation, and irregular or partially preserved traceline morphologies, illustrating the continuum of post-emplacement modification states within this class. All images are oriented with north at the top.

(A) 33.15° S, 50.6° E; CTX image: G04_019899_1467. (B) 36.88° S, 87.25° E; CTX image: P15_007002_1442. (C) 34.21° S, 87.05° E; CTX image: P15_007002_1442. (D) 36.13° S, 76.48° W; CTX image: B16_016027_1445. (E) 42.03° S, 55.47° W; CTX image: P15_007060_1371. (F) 43.52° S, 84.14° W; CTX image: B19_016924_1387. (G) 45.81° S, 162.25° W; CTX image: G09_021687_1339; (H) 39.24° S, 120.1° E; CTX image: G09_021687_1339. (I) 39.29° S, 115.82° E; CTX image: D04_028626_1412. (J) 115.59° E, 38.57° S; CTX image: P17_007792_1397. (K) 42.29° S, 116.29° E; P17_007792_1397. (L) 38.46° S, 17.92° E; CTX image: B18_016551_1413.

In summary, this chapter established a strict morphological framework for identifying CCFs and applied it to the global inventory of Levy et al. (2014). The workflow (Figure 3.21) began with 9,533 entries and involved systematic re-examination in CTX images, classification based on crater-floor coverage and tracelines. This classification yielded a final population of 3,028 high-confidence CCFs, with the remainder classified as uncertain or non-CCFs. Of the retained CCFs, 360 are classed as “classic” examples displaying sharp, well-preserved lineations, while 2,668 represent “non-classic” morphologies that span a continuum from lightly modified to heavily mantled forms. This refined CCF sample forms the foundation for future analyses presented in subsequent chapters.

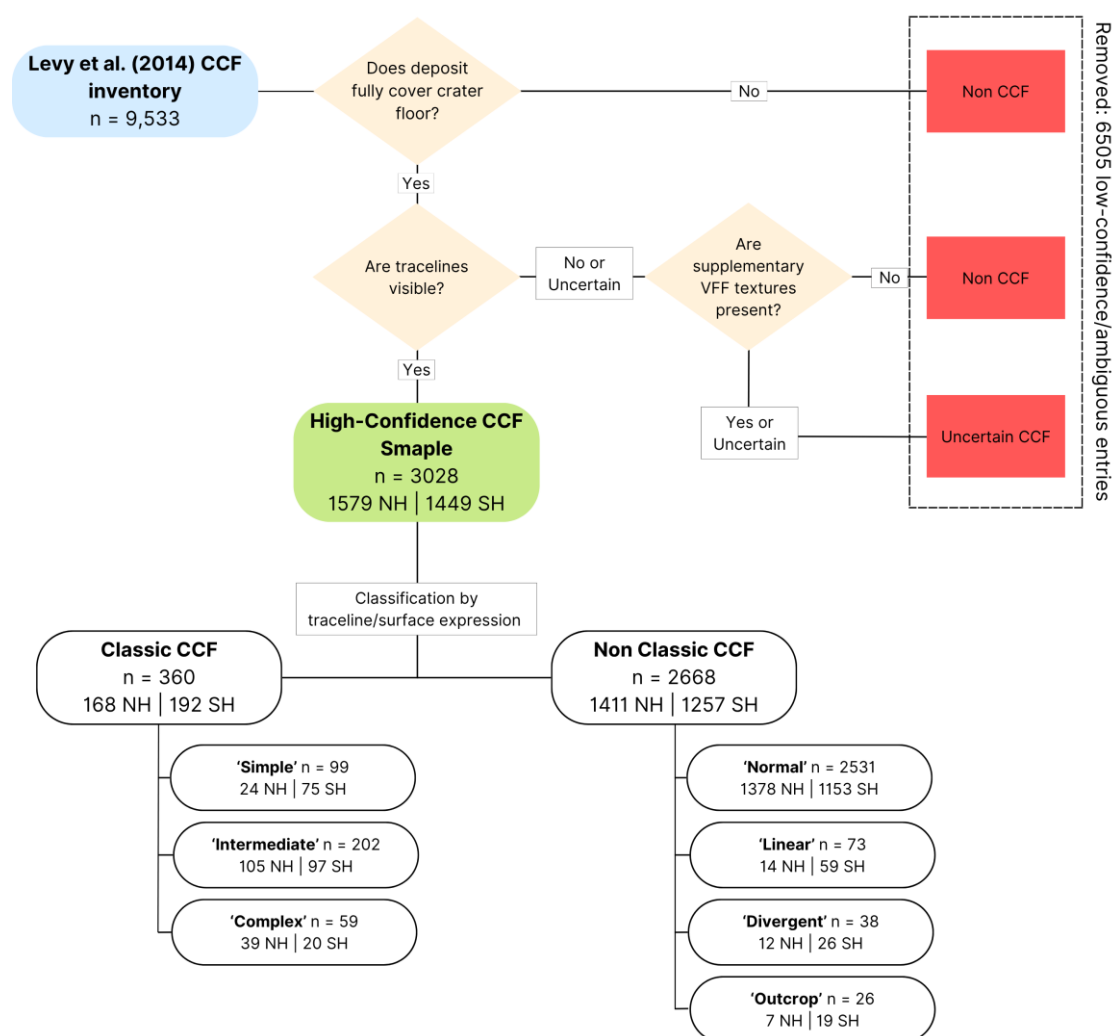


Figure 3.21. Workflow for refining the global CCF inventory of Levy et al. (2014) using the morphological classification scheme developed in this chapter. The process begins with 9,533 candidate CCFs and applies successive screening based on crater-floor coverage and the presence of tracelines, excluding uncertain and non-CCF cases. The final dataset comprises 3,038 high-confidence CCFs which form the basis for all subsequent analyses in this Thesis, subdivided into 290 “classic” and 2,738 “non-classic” examples. For non-classic CCFs, “Normal” denotes the broader continuum of modified non-classic morphologies not assigned to the linear, divergent, or outcrop-associated subgroups.

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Chapter 4: Global Distribution of CCFs

4.1 Introduction

Since their initial identification and classification, VFFs have consistently been documented as mid-latitude landforms concentrated within the 30–50° latitude belts of either hemisphere (e.g., Squyres, 1978, 1979; Milliken et al., 2003; Levy et al., 2007, 2014; Dickson et al., 2012; Souness et al., 2012). The mid-latitude spatial distribution of LDA, LVF and CCF is well established within the literature, largely due to the substantial effort of Levy et al. (2014) in creating the first global inventory of VFFs (Figure 4.1). Later inventories expanded this work, either through feature-specific catalogues, such as those targeting GLFs (e.g., Souness et al., 2012; Brough et al., 2019; Figure 4.1), or through regional surveys of individual VFF subtypes (e.g., Pedersen and Head, 2010; Levy et al., 2010; Berman et al., 2015; Berman et al., 2021). In many of these studies, different VFF subtypes have been regarded as forming an integrated glacial landscape that reflect long-term ice accumulation and modification under varying climatic regimes (e.g., Head et al., 2010; Souness et al., 2012).

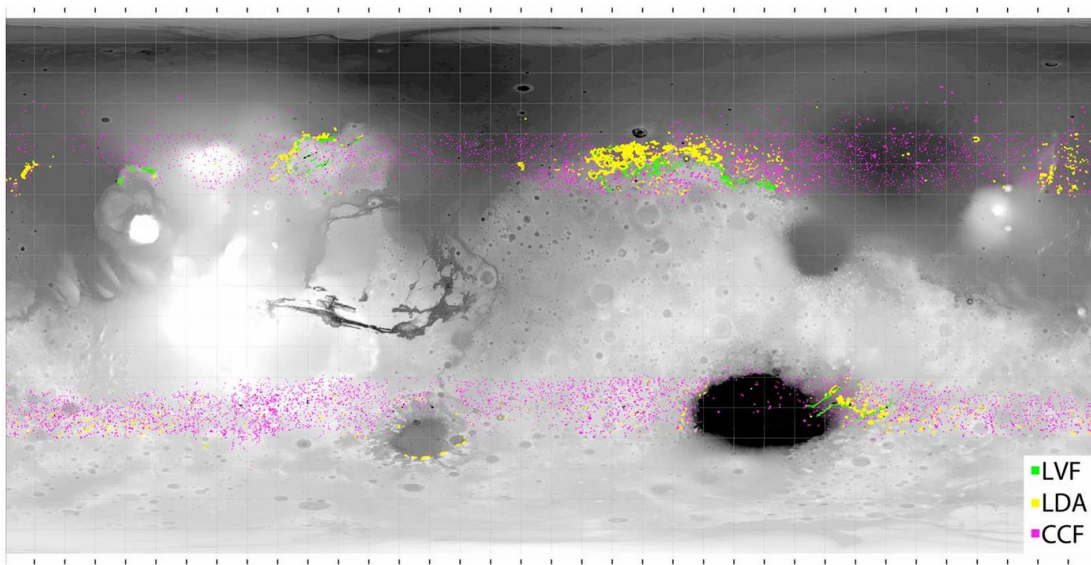


Figure 4.1. Global distribution of CCFs, LVFs and LDAs over MOLA topography, redrawn from Levy et al. (2014).

CCFs constitute the most numerous VFF subtype (Levy et al., 2014); however, as discussed in Chapter 3, studies and inventories of CCFs commonly include degraded or “low-definition” CCFs that lack clearly expressed concentric tracelines or where the crater fill do not extend across the entire crater interior/floor (e.g., Levy et al., 2010; Dickson et al., 2012; Levy et al., 2014). Within this Thesis, crater-filling deposits showing obscured concentric tracelines or not showing them, while potentially ice-rich, do not meet the criteria for high-confidence CCF identification and are therefore excluded from the analysed inventory, as detailed in Chapter 3. Reducing/Limiting the Levy et al. (2014) inventory to high

confidence CCF (Chapter 3) yields 1,579 CCFs in the northern hemisphere and 1449 in the southern hemisphere, spanning 26.5°–65.4°N and 25.7°–50.4°S and occupying craters between 3 and 40 km in diameter. Yet, many morphometrically comparable craters within the same latitude bands—12,756 in the north and 16,026 in the south—do not host CCFs. Considering the near-ubiquitous mid-latitude distribution of high confidence CCFs (Figure 3.9), this stark disparity prompts a central question: why are CCFs observed in some craters but absent in many others?

Addressing this question first requires considering CCFs within the broader context of mid-latitude VFF distribution. Whilst CCFs are described as spatially ubiquitous across the mid-latitudes (Levy et al., 2014), the distribution of non-CCF VFFs (LDA, LVF, GLF) exhibits pronounced zonal clustering, as documented by global and regional surveys (Squyres, 1978, 1979; Milliken et al., 2003; Souness et al., 2012; Levy et al., 2014; Berman et al., 2021; Figure 4.1). In the northern hemisphere, non-CCF VFFs are strongly concentrated along the steep slopes of the dichotomy boundary (e.g., Milliken et al., 2003; Levy et al., 2007; Dickson et al., 2008; Souness et al., 2012; Brough et al., 2019), where parts of the kilometre-scale elevation step coincide with steep slopes with elevation differences of 2–6 km (Guest and Smrekar, 2005). This spatial association implies a dominant topographic control on non-CCF VFF spatial distribution, whereby steep relief provides preferential sites for non-CCF VFF accumulation and/or long-term preservation (Souness et al., 2012). In the southern hemisphere, by contrast, the relationship between non-CCF VFF occurrence and steep relief is less systematic. Despite the widespread availability of steep highland terrain and steeper slopes compared to the northern hemisphere (Tanka et al., 2001; Kreslavsky and Head, 2000), LDA and LVF are distributed more sporadically, interpreted as reflecting the combined influence of slope and thermal constraints (Milliken et al., 2003). In this setting, lower annual mean surface temperatures are thought to restrict LDA and LVF formation and/or preservation to only the steepest, pole-facing slopes (Milliken et al., 2003). GLFs similarly exhibit a preference for relief in both hemispheres, but southern GLFs typically initiate part-way down interior crater walls rather than at plateau edges or cliff tops, contrasting with their northern counterparts (Souness et al., 2012). Collectively, these hemispheric differences demonstrate that while background relief exerts a fundamental control on non-CCF VFF spatial distribution, latitude-dependent thermal conditions modulate where these landforms can accumulate and/or persist (Milliken et al., 2003; Souness et al., 2012). However, it remains unresolved whether current non-CCF VFF spatial distribution reflects conditions necessary for their initial accumulation during mid-latitude glaciation, or conditions favouring their preservation during post glacial periods (Milliken et al., 2003; Souness et al., 2012).

CCFs, by nature of being crater confined, differ fundamentally from other VFFs in that CCFs are entirely bounded by their host topographic constraint. However, if steep slopes are indeed a prerequisite for *all* VFF formation then crater interior wall slopes inherently provide a key condition for/allowing CCF onset. Global topographic analyses show that crater rims and interior walls represent some of the

steepest surfaces on Mars, commonly exceeding 10° and locally reaching 30° in pristine craters (Kreslavsky and Head, 1999; Kreslavsky and Head, 2018), implying that prior to any palaeo-CCF occupation these craters readily supplied gradients adequate for the onset of ice flow. Fastook and Head (2014) utilised a 1D flowband shallow-ice model to demonstrate that slopes greater than $\sim 5^\circ$ are sufficient to initiate ice flow—for ice sheet thicknesses of 200 m—under Late Amazonian climatic conditions. By applying the shallow-ice approximation to integrate strain rates through the vertical ice column, they showed that crater interiors are among the most likely settings to sustain viscous flow under such environments. This combination of steep crater-interior slopes and the widespread availability of impact craters across the mid-latitudes provides abundant sites capable of sustaining viscous ice flow, offering a likely explanation for why CCFs are the most numerous VFF subtype. Nonetheless, their distribution is far from uniform, with both the Levy et al. (2014) inventory and the dataset employed here showing strong spatial variability, from regions of dense clustering to areas largely devoid of CCFs.

These considerations raise another currently unresolved question: if slope availability enables the occurrence of not only LDAs, LVFs and GLFs (Milliken et al., 2003) but also CCFs, is the current CCF distribution simply a by-product of crater availability, such that regions with more craters inevitably host more CCFs? Or do other environmental, geomorphological, or climatic factors govern their spatial distribution? To date, no study has directly compared the spatial pattern of CCF-hosting craters against the spatial pattern of the total crater population, and thus any relationship between crater density and CCF density remains unknown. Addressing this knowledge gap is essential for establishing whether the present-day spatial distribution of CCFs essentially reflects crater availability. Where departures from background crater density occur, it becomes possible to evaluate whether CCF occurrence is associated with systematic, measurable influences—such as latitude, elevation, or host-crater diameter—that may encode aspects of CCF glacial history. Conversely, if no coherent geographic or morphometric associations emerge, it would imply a dominance of localised, crater-scale variables (e.g., microclimate or debris supply) that are not resolved by global analysis.

Using a global, high-confidence CCF dataset and a matched mid-latitude crater population, this chapter investigates whether the present-day distribution of ice-fill reflects the spatial availability of suitable host craters. Specifically, does their present-day distribution simply reflect the spatial availability of suitable host craters, or does it depart from crater density in systematic ways? If departures exist, are they associated with geographic or morphometric variables such as latitude, elevation, or host-crater diameter? How does CCF distribution compare with that of other mid-latitude VFFs, and do regions of high CCF density or absence coincide with broader glacial landscapes? Ultimately, can spatial variations in CCF spatial distribution be attributed to differential accumulation, differential preservation, or a combination of both? The results establish that CCF occurrence is not governed primarily by crater availability but instead requires the invocation of additional environmental and geomorphological

controls, providing a baseline for subsequent chapters that examine specific mechanisms of accumulation and ablation.

Section 4.2 sets out the datasets and methods. Section 4.3 presents the empirical distributional characteristics of CCFs, describing how their occurrence varies with crater diameter, latitude, and elevation, and how the spatial distribution of CCF-hosting craters differs from that of the background crater population. Section 4.4 then evaluates the roles of crater availability, host-crater geometry, latitude, and regional topography, and compares CCF density with other VFFs and ice-related landforms, alongside predictions from mid-latitude glaciation GCMs. The chapter concludes by examining regions of high, low, and absent CCF occurrence to constrain the environmental conditions governing where crater-confined glaciers form and persist.

4.2. Datasets and methods

4.2.1. Datasets

The primary dataset used here is the set of high confidence CCFs. As noted before, it consists of 1579 CCFs in the northern hemisphere (spanning 26.5° to 65.4°N) and 1449 in the southern hemisphere (spanning -25.7° to -50.4°S), occupying craters ranging from 3 to 40 km in diameter. It is a filtered sample of the CCF inventory of Levy et al. (2014), which included ‘classic’ and ‘low-definition’ CCFs. This high-confidence CCF sample is also the focus of later chapters in this Thesis. See Chapter 3 for details on its derivation.

The other key input dataset, a global crater inventory, is essential for both normalisation and contextual analysis across this and subsequent chapters. For this purpose, I utilise the Robbins and Hynek (2012) global crater inventory, which catalogues impact craters ≥ 1 km in diameter. For craters with diameters ≥ 3 km, the Robbins and Hynek crater inventory additionally records parameters such as depth, morphological classifications (e.g. crater morphometrics, ejecta), infill state, and a relative assessment of pristineness. Although the Robbins and Hynek crater inventory was constructed from an incomplete image base—relying primarily on THEMIS Daytime IR mosaics from 2008–2009, which were ~10% incomplete and produced at 256 pixels/degree (~232 m/pixel)—it remains the most widely utilised global crater inventory within the literature.

Although Chapter 3 identifies 3028 high-confidence CCFs, the Robbins and Hynek (2012) global crater inventory provides rim and floor elevation data for only 2495 of their host craters. Later chapters therefore use this smaller subset of 2495 CCFs where host-crater rim elevation is required. This reduction reflects metadata availability within the crater database rather than further filtering of the high-confidence CCF inventory.

To enable comparison between craters that host CCFs and those that do not, a subset of mid-latitude impact craters was extracted from the global Robbins and Hynek (2012) inventory to represent the

population of craters geomorphically and geographically capable of hosting CCFs. The selection retained only craters between 3 and 40 km in diameter and located within -55° to -20° and 20° to 70° latitude, matching the established size and latitudinal bounds of CCFs (Chapter 3). Applying these constraints yields a set of 31810 craters (14335 in the northern hemisphere and 17475 in the southern hemisphere), hereafter referred to as the “filtered crater population”. Note that the CCF-hosting craters (1579 in the north and 1449 in the south) fall within this population. Crater diameter, latitude, longitude, and mean rim elevation were extracted directly from the Robbins and Hynek (2012) database, with crater rim elevation treated as a proxy for background topographic elevation. Geographic coordinates of crater centres were then transformed into a Lambert Conformal Conic (LCC) projection, with standard parallels set at $\pm 35^{\circ}$ and $\pm 55^{\circ}$ for each hemisphere, which preserves local angles, shapes, and azimuthal orientations along the selected parallels while minimising distortion across the mapped extent. However, the linear scale factor k —defined as the ratio of projected (map) distance to true (ground) distance—varies as a function of latitude. While k remains in unity ($k = 1$) exactly at the standard parallels, it decreases slightly between them ($k = 0.98$ at 45° latitude) and increases toward the projection limits $k = 1.04$ at 25° latitude and $k = 1.11$ at 70° latitude). To account for this, subsequent spatial analyses (Section 4.2.2) utilise k to locally adjust search radii, ensuring a constant physical sampling area on the Martian surface.

For comparison, several non-CCF VFF inventories with complete mid-latitude coverage are used also. These include the global inventory of LDA and LVF compiled by Levy et al. (2014), mapped using CTX image mosaics (down sampled to ~ 18 m/pixel) (Malin et al., 2007) and supplemented by THEMIS-VIS mosaics (Christensen et al., 2004) digitised at $\sim 1:250,000$ scale, and the GLF inventory of Brough et al. (2019), digitised from CTX images at $\sim 1:25,000$ scale. All datasets were processed and analysed in ArcGIS Pro 3.3.0, with the CCF and global crater inventories additionally examined in MATLAB R2024a.

4.2.2. CCF and crater heatmaps

To calculate CCF and crater spatial density, heatmaps were constructed for both hemispheres using complementary metrics. *Crater abundance* counts all craters from the filtered crater population within a fixed search radius at each grid node, whereas *CCF abundance* counts only those that host CCFs. From these, *CCF occupation* is calculated as the number proportion of CCF-hosting craters to the filtered crater population count within the same radius, expressed as a percentage. Unlike CCF abundance, CCF occupation is a normalised metric that accounts for the background crater population.

For mapping these quantities, a latitude–longitude meshgrid was generated for each hemisphere (20° – 70° N; -55° to -20° S) in MATLAB and projected into the hemispheric LCC projections detailed above. The search radius used at each grid node for CCF, and crater counting is based on Euclidean distance in the projected space (Figure 4.2). Tests with radii of 50, 75, 100, 150, and 200 km were undertaken.

A 75 km radius was selected as optimal: sufficiently broad to accommodate the spatial footprint of the largest craters (≤ 40 km diameter) yet small enough to resolve local density variations without over-smoothing distinct clusters. To ensure a consistent physical sampling area, the 75 km radius was locally adjusted at each node using the scale factor k (as detailed in Section 4.2.1) to prevent projection-induced bias in the resulting heatmaps. This produced smoothed density maps (i.e. “heat maps”) for both CCF abundance and crater abundance, from which a third heatmap was derived expressing the CCF occupation. For clarity, abundance maps presented in this chapter show raw counts of craters or CCF-hosting craters within each 75 km search radius, not number densities normalised by area. Because the search radius is fixed everywhere, the sampling area is constant, and abundance and areal density differ only by a uniform scaling factor. Grid spacing was evaluated in projected space to ensure appropriate sampling. In the LCC projection, a 1° node separation corresponds to ~ 56 – 62 km in the north–south direction and ~ 22 – 56 km in the east–west direction across the latitude–longitude mesh grid domain. As these values are smaller than the 75 km search-window radius, adjacent sampling windows overlap across the grid, ensuring continuous spatial coverage with no omission of regions due to node spacing, except locally at the projection margins where truncation occurs. Accordingly, working with raw abundance preserves all spatial information while retaining more intuitive values. By contrast, the CCF occupation map is explicitly a normalised metric, expressed as the percentage of craters within each search window that host CCFs.

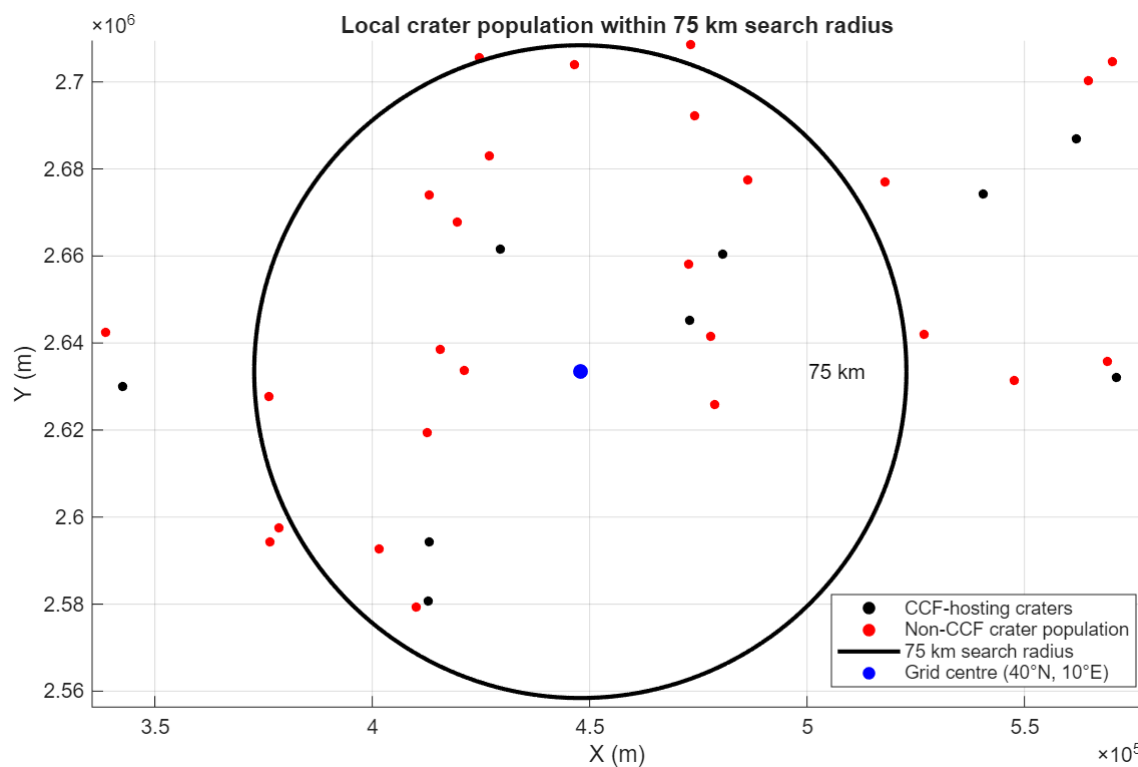


Figure 4.2. Search-radius method for calculating CCF and non-CCF crater densities. Example of the 75 km Euclidean search radius applied at each grid node in the Lambert Conformal Conic projection centred at 40°N, 10°E. CCF-hosting craters appear in black and non-CCF hosting craters in red.

The prime meridian is a discontinuity in LCC space, artificially separating CCFs straddling the boundary. To produce hemispheric heatmaps without associated projection-edge artefacts, I modified their derivation to ensure accurate counting (of CCFs and craters) near 0°–360° transition as follows. Longitudes were recentred on 180° and crater densities recomputed within three overlapping longitudinal windows: (A) –180° to –90° and (C) 90° to 180° in the shifted reference frame (Figure 4.3A), and (B) –90° to 90° in the original 0°-centred system (Figure 4.3b). These outputs were then mosaicked to produce a single seamless distribution spanning –180° to 180° (Figure 4.3c) for export into ArcGIS Pro.

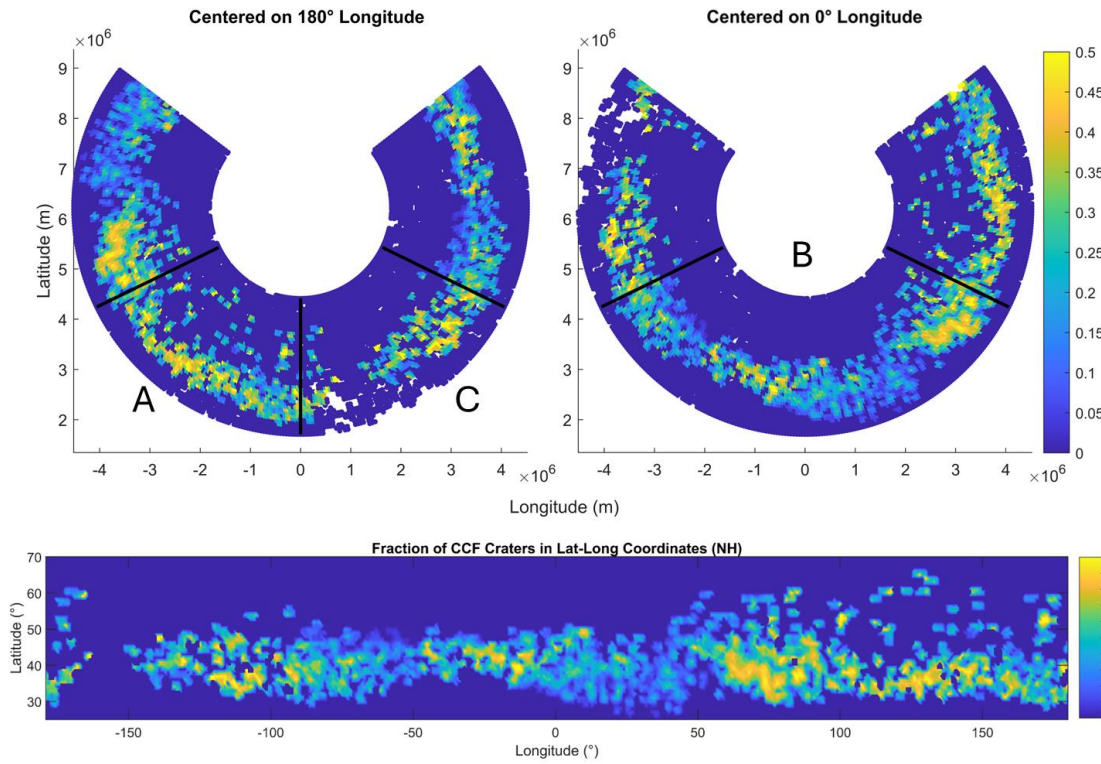


Figure 4.3. Construction of a seamless hemispheric heatmap in Lambert Conformal Conic projection.

(A) Northern Hemisphere CCF percentage occupation map computed in an LCC projection recentred on 180°, providing coverage for the longitudinal windows A: -180° to -90° and C: 90° to 180° . (B) Corresponding map computed in the standard 0° -centred LCC projection, providing coverage for the window B: -90° to 90° . (C) Mosaicked product formed by combining the three overlapping longitudinal windows from panels A and B to produce a single continuous distribution spanning -180° to 180° , thereby eliminating projection-edge artefacts at the prime meridian.

This corrected framework enables both visualisation and quantitative assessment of CCF occurrence relative to the wider mid-latitude crater population. The resulting maps enable comparison between CCF abundance and relative occupation, isolating spatial variability in CCF concentration from background variations in overall crater density.

4.2.3. Heatmap statistical analysis

To assess whether the spatial distribution of CCFs merely reflects the underlying availability of craters or reflects additional environmental controls, the spatial abundancies of CCF-hosting craters were compared against the spatial abundance of the filtered mid-latitude crater population using a scatterplot and quantifying correlation. Each grid cell provides a paired value (CCF-hosting crater abundance; filtered crater abundance), which is plotted to test whether CCF occurrence scales proportionally with crater availability. Association is quantified using Spearman's ρ as CCF distribution is non-normal, with associated p -values used to evaluate statistical significance. Under proportional scaling, grid cells with

higher crater abundance should also exhibit higher CCF abundance, producing a strong positive correlation; departures from this proportionality—manifesting as high scatter or low correlation coefficients—indicate that factors beyond crater availability influence where CCFs occur.

Kullback–Leibler divergence (D_{KL}) is used as a secondary, distribution-level measure to compare the normalised abundance distributions of CCF-hosting craters and filtered craters across the analysis grid. This can be applied to gridded heatmap data (Krakov, 2013) using:

$$D_{KL} = \sum_{i=1}^n P(i) \log \frac{P(i)}{Q(i)}, \quad \text{Equation 4.1}$$

where each grid cell defines a discrete bin i (with n total cells), and crater abundances are normalised to form probability distributions $P(i)$ for CCF-hosting craters and $Q(i)$ for all craters. The summation therefore evaluates the weighted log-ratio of probabilities across all grid cells, such that D_{KL} quantifies differences between the two abundance distributions rather than spatial co-location or neighbourhood structure.

Because D_{KL} operates on normalised frequencies, it does not encode spatial adjacency, direction, or co-location between cells. It therefore cannot test whether CCF-hosting craters and all craters occur in the same places. Instead, it evaluates whether the overall distribution of abundance across grid cells is similar between the two populations. In this context, low D_{KL} values ($D_{KL} \approx 0$) indicate that CCFs and craters are distributed across cells in comparable proportions, whereas higher values indicate departures from proportional abundance. D_{KL} is therefore used here only as a supplementary descriptor of distributional similarity, alongside correlation-based analyses that explicitly test spatial co-variation.

4.3. CCF distributional characteristics

Although the heat maps will form the core of the analysis, before presenting their results, first I investigate several non-geographical statistics of the CCF and (filtered) crater distributions. Specifically, I evaluate the physical and environmental attributes of the CCF population, assessing first how the diameter distribution of host craters compares with that of craters that do not contain CCFs (Section 4.3.1), then analysing how CCF occurrence varies with latitude and elevation in order to identify spatial preferences that may record present or past climate controls on CCF formation and/or preservation (Sections 4.3.2 & 4.3.3). This approach is grounded in terrestrial understanding that glacier distribution varies predictably with altitude, latitude (e.g., Evans and Cox, 2005; 2017), and glacier size (e.g., Carrivick and Brewer, 2004). Examining how host-crater diameter, latitude and elevation influence CCF distribution may provide a means to imply the climatic gradients that governed their accumulation and preservation, whilst also testing whether CCF-hosting craters reflect the broader mid-latitude crater population or form a distinctive subset.

After these parametric analyses, I turn to the heatmaps and use them to study the spatial organisation of the CCF population and available crater population (Sec 4.3.4.). By mapping CCF occupation as well as CCF abundance, this analysis separates genuine clustering from variations in background crater density and identifies regions of high CCF occurrence (e.g. where CCFs number few but are disproportionately high in occurrence, due to sparse cratering). Here, I use Spearman's rank correlation (ρ) to quantify the proportionality between these spatial distributions, testing whether CCF patterns scale directly with crater availability or are driven by independent environmental factors. Supplementary to this, D_{KL} divergence is used as a secondary, non-spatial measure to compare the similarity of their probability distributions. Unlike the scatterplot, this metric is location-agnostic and evaluates only the statistical "shape" of the populations rather than their geographical overlap. Hemispheric comparisons in CCF distributions are presented throughout to evaluate contrasts between the northern and southern mid-latitudes.

4.3.1. Diameter distributions of craters with and without CCFs

Figure 4.4 presents diameter-frequency histograms binned by host-crater diameter for the filtered crater population (Figure 4.4 A–C) and for CCF-hosting craters (Figure 4.4 D–F), both globally and for each hemisphere. The filtered crater population displays a non-normal diameter distribution that is strongly right-skewed and leptokurtic, confirming that small craters (diameter < 4 km) dominate the population, and that crater frequency declines rapidly with increasing diameter. The CCF-hosting crater population exhibits a similar distribution shape globally and within either hemisphere, suggesting that CCF counts primarily reflect the underlying crater size distribution.

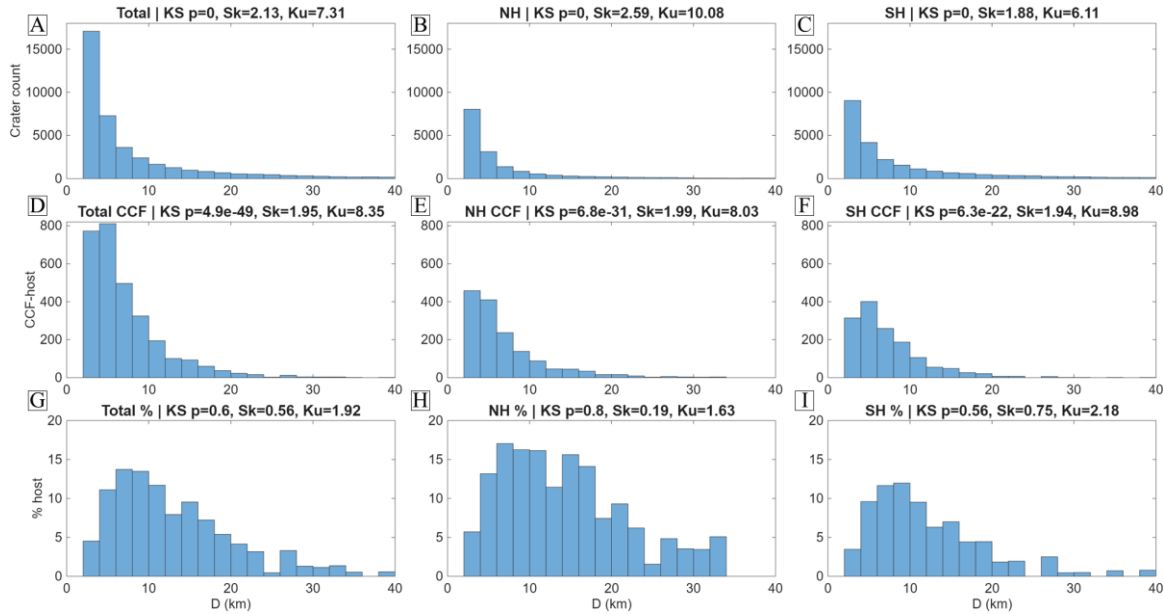


Figure 4.4. Diameter–frequency distributions of filtered craters and CCF-hosting craters, and corresponding CCF occupation by diameter. (A–C) Diameter-frequency histograms of the filtered crater population for (A) all craters, (B) the Northern Hemisphere, and (C) the Southern Hemisphere. (D–F) Diameter-frequency histograms of CCF-hosting craters for (D) all craters, (E) the Northern Hemisphere, and (F) the Southern Hemisphere. (G–I) Percentage of craters hosting CCFs within each diameter bin for (G) the global dataset, (H) the Northern Hemisphere, and (I) the Southern Hemisphere. Kolmogorov–Smirnov test p -values, skewness (sk), and kurtosis (ku) are reported in the titles of each panel.

The percentage of filtered craters that host CCFs provides a normalised perspective. The percentage occupation distributions (Figure 4.4 G–I) differ distinctly from the frequency distributions of the filtered crater population and the CCF-hosting subset (Figure 4.4A–F). They remain weakly to moderately right-skewed, but look closer to log-normal in shape, showing an initial rise in % up to about 10 km in diameter. In the north, percentage occupation remains above ~13% from the 4–6 km bin through to 16–18 km, whereas in the south it exceeds ~9.5% only between 4–6 km and 10–12 km, producing a narrower interval of elevated occupation. Crucially, because craters smaller than ~4 km show low percentage occupation in both hemispheres (<5%), the high absolute abundance of CCFs in small craters reflects their numerical dominance rather than a preferential association with small diameters. These results identify that CCF occupation of (available) craters peaks at intermediate crater diameters, with a broader peak in the northern hemisphere than in the southern hemisphere.

4.3.2. Latitudinal and elevation distribution of CCFs

The distributions of CCFs with latitude and elevation help evaluate any climate controls (or bounds/limits) under which mid-latitude CCFs could form and/or persist. In terrestrial settings, spatial variations in equilibrium-line altitude (ELA)—the altitude at which annual accumulation balances annual ablation—record first-order climatic and mass-balance gradients (e.g., Kuhle, 1988; Nesje, 1992;

Carrivick and Brewer, 2004), and reconstructions of palaeo-ELAs have long been used to infer former glacier environments and mass-balance histories (e.g., Ballantyne, 1990; Benn and Lehmkuhl, 2000; Rea and Evans, 2007).

Terrestrial ELA reconstruction is not applicable to CCFs since their crater-confined evolution precludes the development of distinct accumulation and ablation zones. However, at any given latitude, the elevation range of the surrounding terrain defines the climatic envelope within which ice stability is physically possible. The minimum and maximum elevations at which CCFs occur within a latitude band can therefore be treated as empirical bounds on palaeo-CCF formation and/or preservation. Where CCFs occupy a restricted portion of the available elevation range, this could indicate mass-balance forcing that limited either initial accumulation or long-term preservation at certain elevations. Conversely, where CCFs span the full elevation range permitted by local topography, their distribution is consistent with crater-confined ice being viable across that entire elevational interval. Variation in CCF elevational occupancy with latitude may therefore reflect the combined influence of latitude- and elevation-dependent climatic gradients—such as insolation, temperature, atmospheric pressure, and humidity—acting on CCF mass balance under different orbital configurations.

Figure 3.9 (Section 3.3) locates the high-confidence CCFs, and Figures 4.5 and 4.6 plot the latitudinal frequency of CCFs and the background filtered crater population, respectively, against absolute latitude for both the northern and southern hemispheres. Both hemispheres exhibit broadly comparable CCF latitudinal distributions (Figure 4.5), characterised by a rapid increase in abundance from $\sim 30^\circ$ to a maximum between $\sim 35^\circ$ and 40° , followed by a decline toward higher latitudes. This correspondence is statistically robust: the difference in median absolute latitude is negligible ($\sim 0.2^\circ$), and a Mann–Whitney U test confirms that the hemispheric distributions are indistinguishable ($p = 0.158$). Despite this statistical alignment in central tendency, subtle differences persist in distributional form. The northern CCF population extends $\sim 15^\circ$ further poleward than the southern population, although this high-latitude subset ($\sim 50^\circ\text{N}$) comprises only 56 CCFs concentrated in Arcadia and Utopia Planitiae, its presence results in a northern dataset that is right-skewed and leptokurtic. In contrast, the southern dataset, which lacks this secondary high-latitude concentration, remains approximately symmetric and platykurtic.

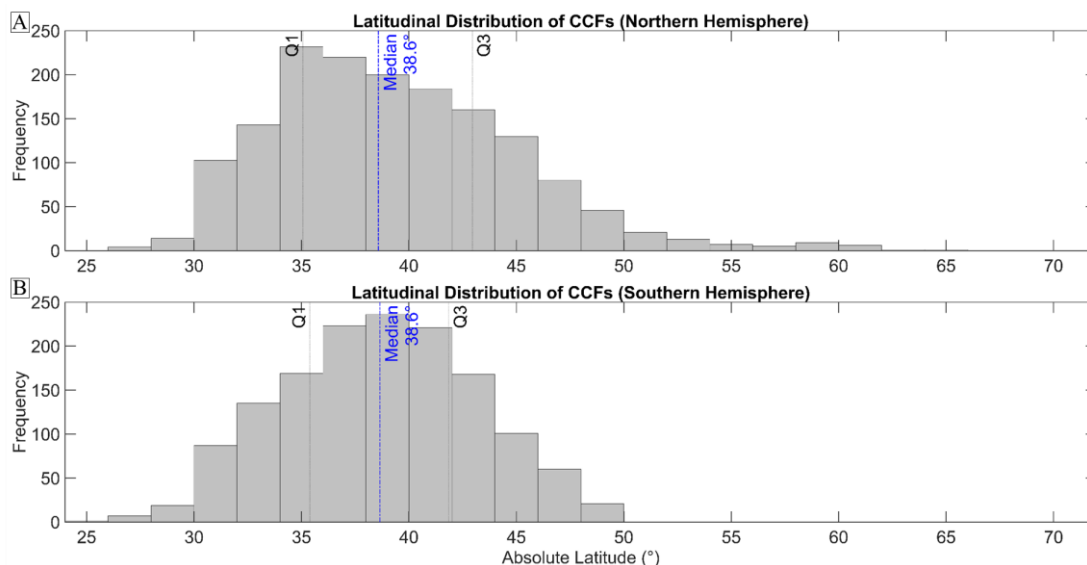


Figure 4.5. Latitudinal distribution of CCF-hosting craters. Latitudinal frequency histograms of CCF-hosting craters for (A) the Northern Hemisphere and (B) the Southern Hemisphere. The median latitude and the first and third quartiles (Q1 and Q3) are indicated on each panel.

By contrast, the filtered crater population exhibits a strong hemispheric contrast in absolute crater abundance (Figure 4.6). Although crater frequency decreases with increasing latitude in both hemispheres, the southern hemisphere consistently contains more craters per latitude bin: no northern hemisphere latitude bin exceeds ~ 1000 craters per 2° interval, whereas all southern hemisphere bins exceed ~ 1050 craters. Within the latitude ranges over which CCFs occur, the southern hemisphere contains 16026 craters compared with 12756 in the north, despite spanning a $\sim 15^\circ$ narrower latitude band. If CCF occurrence reflected host-crater availability alone, a substantially larger southern CCF population would be expected. The absence of such a correspondence indicates that factors beyond crater availability influence present-day CCF distribution.

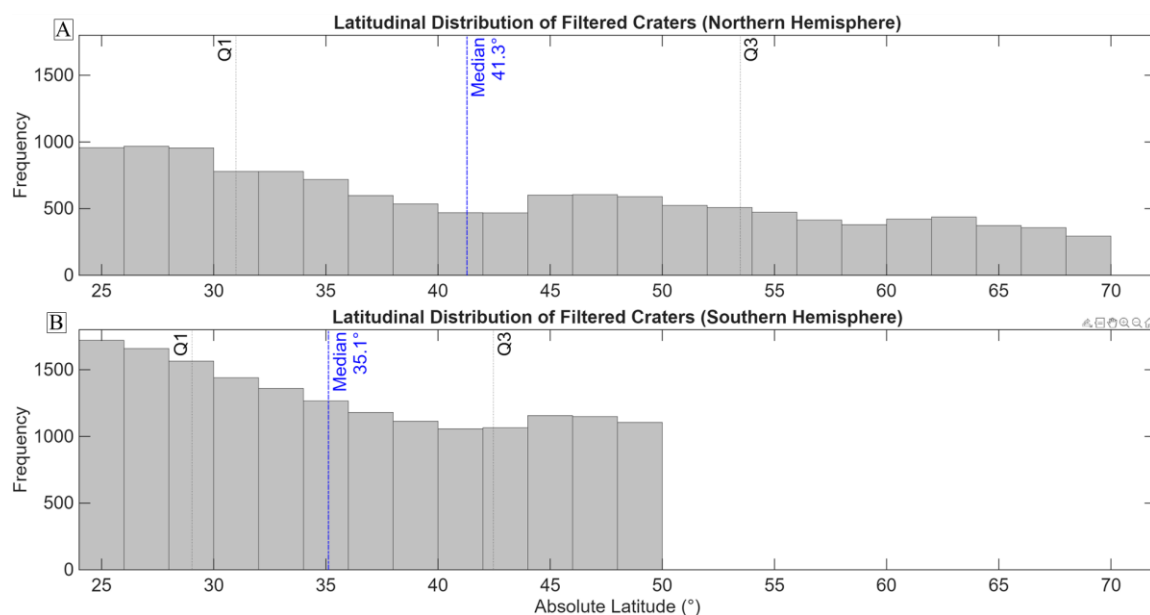


Figure 4.6. Latitudinal distribution of the filtered crater population. Latitudinal frequency histograms of the filtered crater population for (A) the Northern Hemisphere and (B) the Southern Hemisphere. The median latitude and the first and third quartiles (Q1 and Q3) are indicated on each panel.

Figure 4.7 plots CCF elevations against latitude, compared with the latitude-dependent elevation range of the filtered crater population, adding a second dimension to their geographic distribution and revealing a strong hemispheric contrast in elevation values. This offset reflects the primary topographic signal of the Martian crustal dichotomy (Figure 1.1). In the northern hemisphere, both CCFs and the filtered non-CCF crater population occur mainly between -5 and 1 km, corresponding to the elevation range of the northern lowlands and adjacent dichotomy-boundary terrain. Conversely, both populations occur predominantly between 0 and 3.5 km in the southern hemisphere, within the dissected Noachian highlands.

From their equatorward limits at $\sim 25\text{--}28^\circ$ S and $\sim 25\text{--}32^\circ$ N, CCFs in either hemisphere are restricted to elevations below 0 km. Between $\sim 32\text{--}53^\circ$ N and $\sim 28\text{--}42^\circ$ S, CCFs span the full elevation range permitted by local topography, indicating that mid-latitude conditions within these central bands were compatible with palaeo-CCF formation and persistence across all elevations. Poleward of these intervals, CCF elevational occupancy becomes progressively restricted. In the north (Figure 4.7A), CCF elevations converge toward ~ -4 km as occurrences become confined to the northern lowlands. In the south (Figure 4.7B), the occupied elevation range narrows as minimum elevation increases with latitude, while maximum elevation remains broadly constant. Collectively, these observations indicate that each hemisphere contains a central mid-latitude zone in which CCF formation and preservation were viable across the full topographic envelope, whereas both equatorward and poleward margins exhibit

constrained elevational occupancy consistent with either reduced palaeo-accumulation or diminished post-glacial preservation at certain elevations.

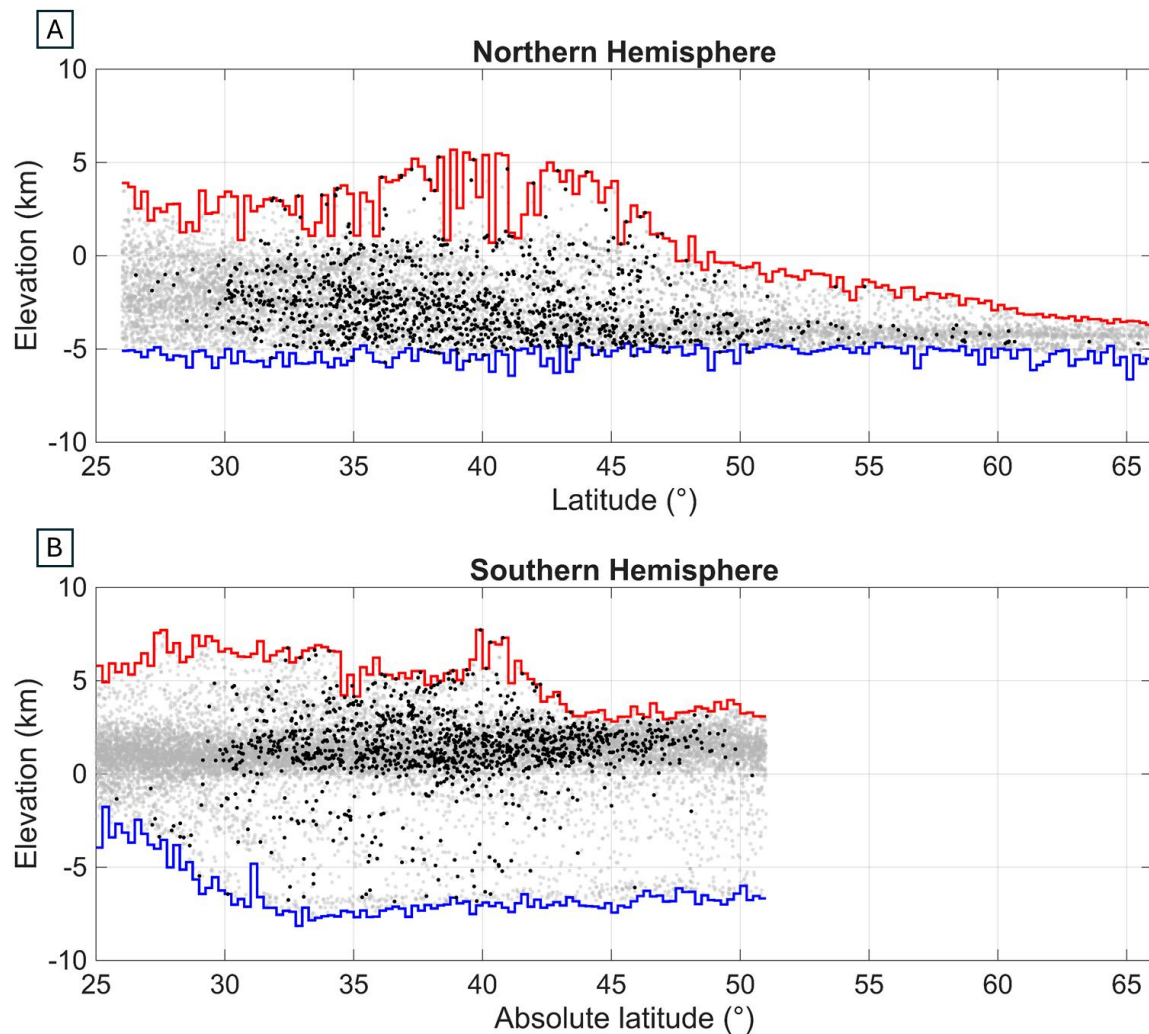


Figure 4.7. CCF surface elevation as a function of latitude compared with the elevation range of the filtered crater population in (A) the northern hemisphere and (B) the southern hemisphere. Black points show individual CCF surface elevations plotted against absolute latitude. Grey points show the elevation distribution of the filtered non-CCF crater population. Red and blue stepped lines show the maximum and minimum crater elevations, respectively, within 0.25° latitude bins. Northern hemisphere CCFs span 26–66°N, whereas southern hemisphere CCFs span approximately 25–52°S.

4.3.3. Spatial abundance of CCFs

Heatmaps of CCF abundance (Figure 4.8) illustrate marked spatial variability in CCF density across either hemisphere. Localised zones of high (>6 CCFs per cell) and low (<2 CCFs per cell) CCF abundance occur alongside extensive areas of complete CCF absence, notably within Hellas Basin and Arcadia Planitia. In Figure 4.8, the black outlined regions delineate regions of CCF abundances above 6, in order to highlight abundance “hotspots”, which occur as spatially discontinuous clusters, with individual cells containing between 6 and 15 CCFs within a 75 km radius.

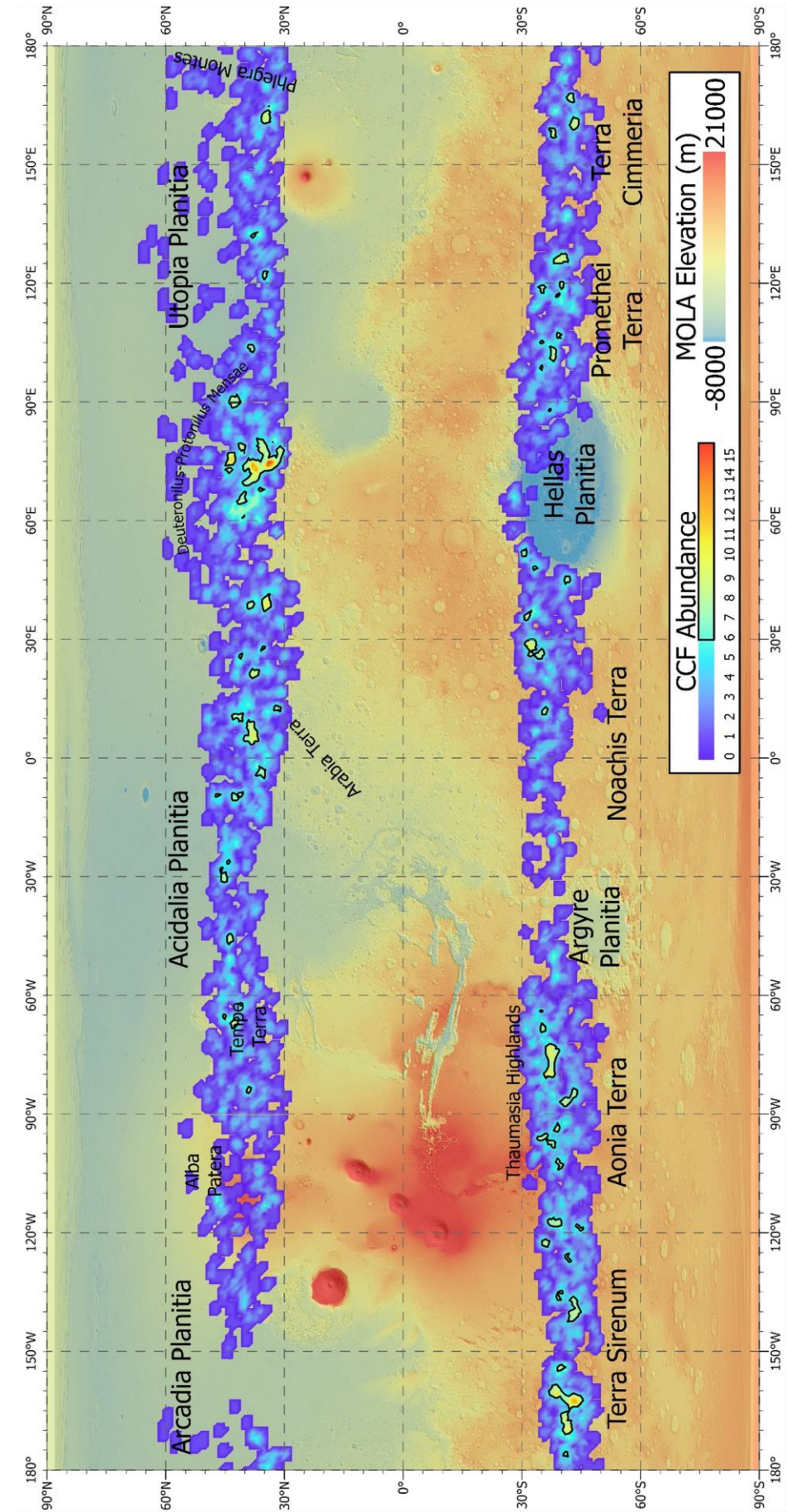


Figure 4.8. Heatmap of CCF abundance. Heatmap showing CCF abundance in each hemisphere overlaid on MOLA gridded topography and MOLA-derived hillshade. Black overlay represents clusters of ≥ 6 CCFs.

The heatmap of the filtered crater population (Figure 4.9) shows variable background crater density, with a clear contrast in mean density between the hemispheres. The northern mid-latitudes exhibit a low hemispheric mean of ~6 craters per cell, punctuated by discrete high-density concentrations in Tempe Terra, the high latitudes of Acidalia Planitia, Deuteronilus Mensae, and the highlands of Arabia Terra, alongside localised zones of crater scarcity in Arcadia Planitia and Utopia Planitia. The southern mid-latitudes display a higher hemispheric mean of ~11 craters per cell and a more heterogeneous spatial distribution, with distinct low-density regions (<3 craters per cell) across Daedalia Planum, the Thaumasia highlands, Neridum Montes in the northeastern Argyre Basin, and within Hellas Basin and Hesperia Planum. The hemispheric contrast in crater density reflects the underlying crustal dichotomy: ancient Noachian terrains, which dominate the southern highlands and are also preserved in northern regions such as Arabia Terra, retain systematically higher crater densities than younger resurfaced provinces, including much of the northern lowlands. Local departures from this framework include anomalously high crater densities within Acidalia Planitia, and low crater densities within the interiors of large impact basins such as Hellas and Argyre, the latter likely reflecting extensive degradation and resurfacing (Tanaka et al., 2001).

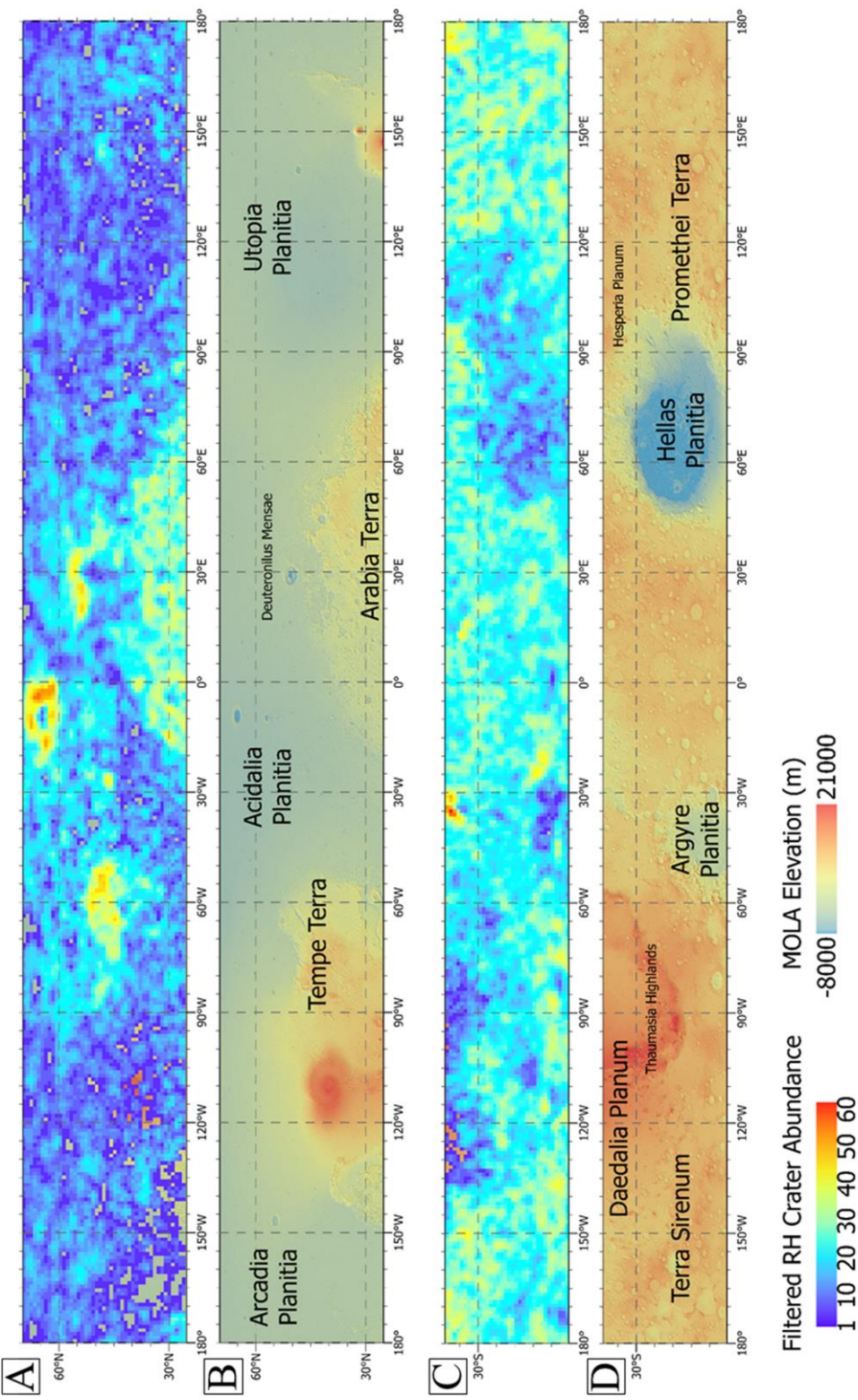


Figure 4.9. Spatial distribution of filtered Robbins-Hynek crater inventory abundance. Heatmaps show the abundance of filtered RH craters overlaid on MOLA gridded topography with MOLA-derived hillshade. (A) Northern hemisphere filtered RH crater abundance. (B) Northern hemisphere MOLA reference map with labels for context. (C) Southern hemisphere filtered RH crater abundance. (D) Southern hemisphere MOLA reference map with labels for context.

Comparison of the heatmaps in Figures 4.8 and 4.9 suggests that CCF abundance does not scale simply or proportionally with the abundance of available host craters. This relationship is examined directly by plotting, for each grid cell, CCF abundance against filtered crater abundance (Figure 4.11). In both hemispheres, the two quantities are positively correlated (Spearman's $\rho = 0.47$ in the north and $\rho = 0.42$ in the south; $p < 0.001$ in both cases), indicating that regions with more craters do, on average, host more CCFs. However, the substantial scatter in these relationships demonstrates that crater availability explains only a moderate fraction of the spatial variability in CCF abundance. Many cells with similar crater abundances host markedly different numbers of CCFs, and conversely, comparable CCF abundances occur across a wide range of crater densities.

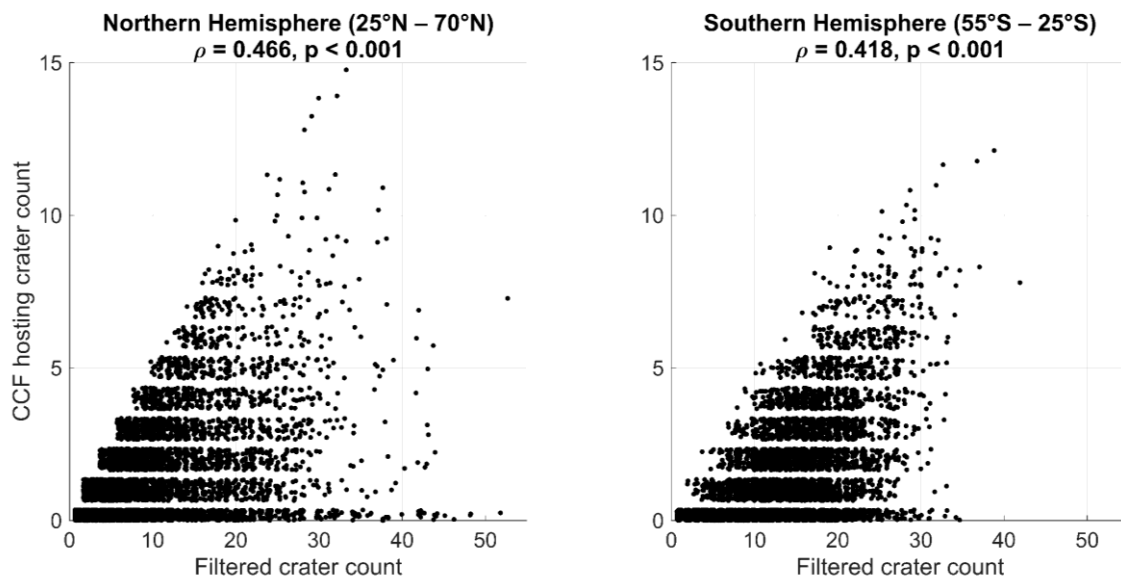


Figure 4.10. Proportionality between CCF abundance and background crater abundance. Scatterplots of CCF abundance versus filtered background crater abundance per $1^\circ \times 1^\circ$ grid cell for (A) the Northern Hemisphere ($25^\circ\text{--}70^\circ\text{N}$) and (B) the Southern Hemisphere ($25^\circ\text{--}55^\circ\text{S}$). ρ and associated p -values are annotated. Data points include a small random positional jitter to visualise the density of discrete integer counts and to reduce overplotting.

The distributions also differ in a statistical sense ($D_{KL} = 2.09$ in the north and 1.24 in the south), indicating substantial divergence between the *abundance distributions* of CCF-hosting craters and the underlying crater population. As this metric does not encode spatial correspondence, it is not used further here.

Thus, while crater availability exerts a first-order control on where CCFs can occur, it is not the dominant factor governing their spatial distribution. Instead, additional controls related to regional or local environmental conditions must play a major role in determining whether a given crater ultimately hosts a preserved CCF. This inference is consistent with the strong hemispheric and regional contrasts

observed in the heatmaps and points toward differences in glacial accumulation, preservation, and post-glacial modification history as key drivers of the observed CCF distribution. This result motivates an examination of CCF percentage occupation, which quantifies the proportion of craters hosting CCFs independent of local crater abundance.

4.3.4. CCF percentage occupancy

CCF percentage occupancy measures CCF density relative to the spatial-varying density of the underlying crater population. Since it controls for crater abundance at each grid cell, it can highlight regions where CCF formation or preservation was proportionally efficient despite low counts. Figure 4.11 portrays its heatmap. Percentage occupancy values reach ~50% in both hemispheres, although high-percentage cells are more common in the north, a consequence of normalisation against the substantially larger crater population in the south.

In the northern hemisphere, regions exceeding 40% occupancy are widespread (Figure 4.11, black outlines), with values approaching 50% across Nilosyrtis Mensae, Alba Patera, and along the southern and western perimeter of Utopia Planitia. Additional high-occupancy clusters occur across the northern lowlands, including central Utopia, Acidalia and Arcadia Planitiae. These clusters typically correspond to areas with few craters but a high proportion of CCF-hosting craters and include several zones that are not apparent in the raw abundance maps. The southern flank of Alba Patera is a clear example, where high occupation remains inconspicuous when considering CCF counts alone. By contrast, regions such as Arabia Terra exhibit high CCF abundance but low percentage occupation due to dense crater backgrounds. In the southern hemisphere, high-occupancy regions are rare (Figure 4.11, black outlines), reflecting the increased crater population. Isolated high occupation clusters occur in the Thaumasia Highlands and in Promethei Terra, East of Hellas, yet many southern high-abundance regions identified in the raw heatmaps (Figure 4.8, black outlines) do not remain prominent once normalised to crater density (Figure 4.11, black outlines).

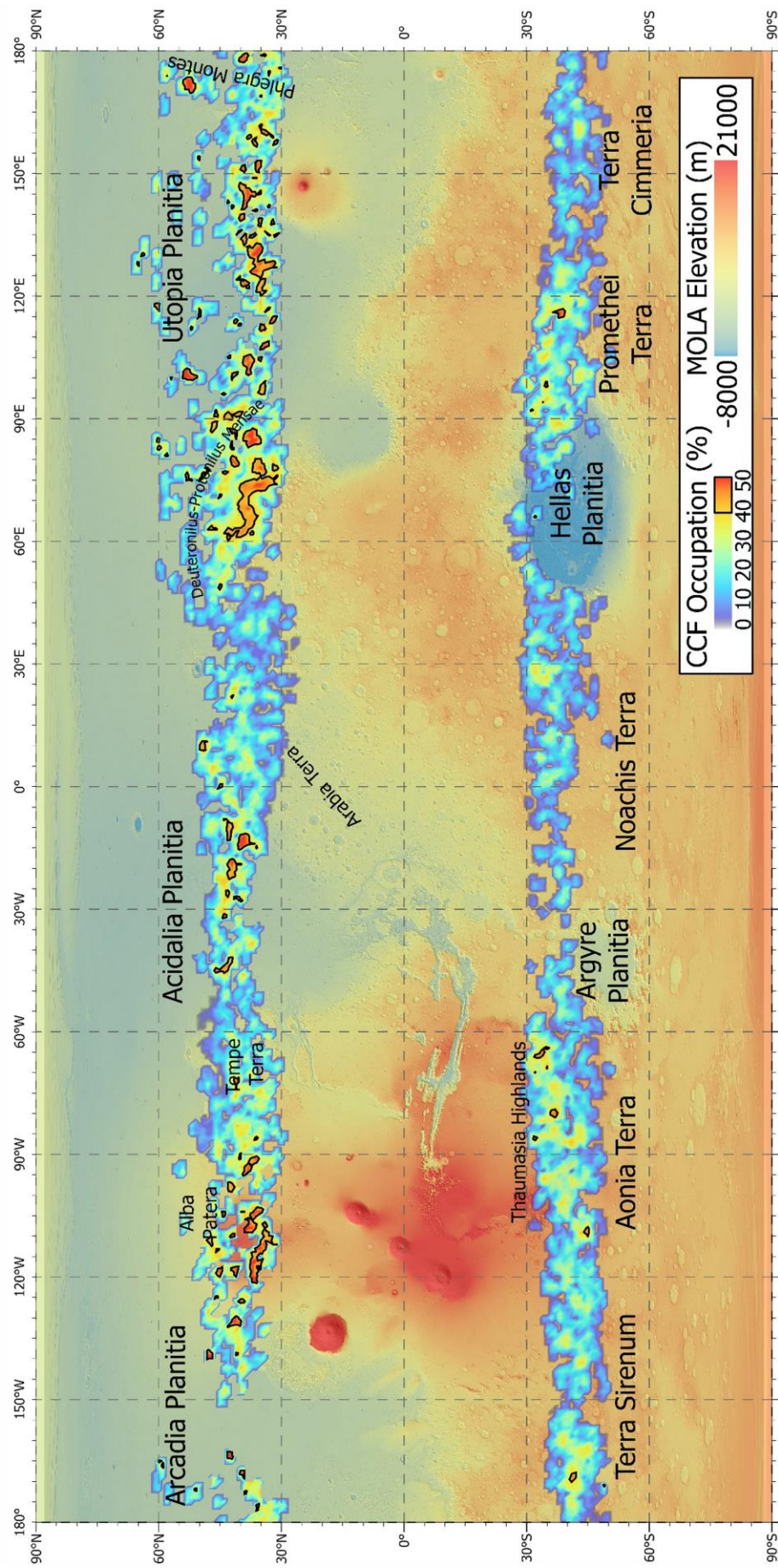


Figure 4.11. Heatmap of CCF crater occupation. Heatmap showing CCF percentage occupation in each hemisphere overlaid on MOLA gridded topography and MOLA-derived hillshade. Black overlay represents clusters of $\geq 40\%$ CCFs.

4.4. Evaluating controls on CCF spatial distribution

The evidence presented above demonstrates that CCF occurrence cannot be explained as a simple by-product of crater availability. While this relationship may appear intuitive, its finding is crucial, and this conclusion is supported by three independent lines of analysis. First, direct cell-by-cell comparison of CCF abundance against filtered crater abundance (Figure 4.11) shows only a moderate positive correlation (Spearman's $\rho = 0.46$ in the north and 0.42 in the south; $p < 0.001$), with substantial scatter, demonstrating that crater availability explains only a limited fraction of the spatial variability in CCF abundance and that proportional scaling does not hold. Second, a comparison of the normalised abundance distributions reveals a statistically significant divergence, with D_{KL} values of 2.09 in the northern hemisphere and 1.24 in the southern hemisphere. This divergence confirms that the frequency of CCF-hosting craters does not scale proportionally with the overall crater population; the two populations represent distinct probability distributions rather than one being a representative subset of the other. Third, the non-CCF crater population exhibits strong hemispheric asymmetry, with the southern hemisphere containing approximately 50% more craters at equivalent latitudes, yet CCF abundance remains broadly comparable between hemispheres. If CCF development were directly proportional to crater frequency, a corresponding southern excess would be expected.

This constitutes the first major result of the Thesis: that CCF distribution is not governed by crater availability alone, and explaining it requires considering other controls, e.g., plausibly controls linked to host-crater properties and geographic setting. The next sections proceed to examine how key host-crater parameters—diameter, latitude, and elevation—may have regulated the accumulation and preservation of CCFs under alternating glacial and interglacial conditions. Establishing the result is essential, as it means that the findings below can be interpreted in terms of glacial and climatic influences rather than artefacts of the underlying crater population. The result also motivates the consideration of host-crater and geographic factors in the work of subsequent chapters.

4.4.1 Host crater properties and CCF spatial distribution

Host-crater diameter exerts the strongest first-order influence on CCF spatial distribution. In both hemispheres, CCF occurrence is highest within intermediate-sized craters (4–12 km in the south; 4–18 km in the north) (Figure 4.4G–I), despite CCFs being numerically most abundant in craters smaller than ~6 km (Figure 4.4D–F). This distinction indicates that small craters dominate CCF counts primarily because they dominate the crater population, whereas the probability of hosting a CCF increases with crater diameter up to ~18 km. The observed hemispheric differences therefore do not reflect simple scaling with crater frequency but instead identify diameter ranges within which conditions were more favourable for CCF formation and/or long-term preservation. This diameter dependence is consistent with terrestrial observations that larger glaciers respond more slowly to climatic perturbations than smaller ones (e.g., Bahr and Pfeffer, 1998; Raper and Braithwaite, 2009). By analogy, CCFs in small

(<6 km) craters may have been more sensitive to orbitally driven insolation variability and post-glacial mass loss, whereas intermediate-sized craters may have provided greater volumetric and thermal buffering across glacial–interglacial cycles. The absence of a continued increase in CCF occupation within larger craters therefore implies that additional size-dependent constraints—beyond simple volumetric stability—limit the formation or long-term persistence of crater-confined ice in large cavities.

Methodological effects related to the Chapter 3 classification criteria (Section 3.2) likely contribute to the apparent preference of CCFs for intermediate-sized host craters. Because CCFs are identified only where the deposit completely covers the crater floor and preserves visible tracelines, detectability depends directly on surface area and the susceptibility of surface textures to mantling. Larger CCF surfaces provide more extensive exposures over which tracelines can be expressed and resolved in CTX images, increasing the probability of correct classification. By contrast, the small surface areas of CCFs in small craters require only limited mantling to obscure diagnostic morphology, rendering many such deposits undetectable within the Chapter 3 mapping workflow. Intermediate-sized craters therefore occupy an observational optimum: they are large enough to preserve and expose resolvable tracelines, yet not so heavily mantled that surface expression is commonly masked. This detectability bias likely contributes to the reduced representation of small craters in the inventory, but it cannot by itself explain the systematic diameter dependence observed in the dataset.

Both CCF abundance and percentage occupation decline beyond ~18 km diameter (Figure 4.4G–I), indicating that increasing crater size introduces constraints that progressively inhibit the persistence of crater-confined ice. No CCFs occur in craters larger than ~40 km, although several large craters contain substantial ice-rich fills that preserve tracelines while leaving central peaks exposed (Chapter 3). These fills demonstrate that significant ice accumulation can occur within large craters without producing CCF morphology. Two end-member explanations remain unresolved. Either such fills never attained sufficient thickness to overtop the central peak, implying that the ice volumes required in craters >40 km exceeded Amazonian mid-latitude accumulation potential, or they once exceeded the peak and subsequently thinned during prolonged post-glacial ablation. The latter interpretation implies an upper stability limit in which the large surface area and volume of palaeo-CCFs may inhibit development of a sufficiently thick protective debris lag, promoting sustained mass loss until ice surfaces retreat below peak elevation. Both scenarios indicate intrinsic size-dependent constraints on CCF growth and stability imposed by host-crater geometry rather than a simple absence of glaciation at larger scales.

While host-crater diameter defines the volumetric capacity and geometric boundary conditions, it cannot account for the broader spatial distribution of CCFs. Their occurrence is further restricted by latitude and elevation, which together define the climatic window within which crater-confined ice could accumulate and persist.

The latitudinal distribution of CCFs delineates this window clearly (Figure 4.5). CCFs cluster between $\sim 30^\circ$ and 45° absolute latitude in both hemispheres, consistent with zones of mid-latitude ice stability predicted under past obliquity regimes (Forget et al., 2006; Madeleine et al., 2009). At lower latitudes, elevated insolation likely limited long-term preservation, whereas poleward of this zone the scarcity of CCFs may reflect reduced net accumulation or subsequent modification that hindered long-term survival. The narrower southern distribution may represent a climatic accumulation margin, while the broader northern extent likely reflects locally favourable preservation conditions in regions such as Utopia and Arcadia Planitia. The small number of high-latitude northern examples are best interpreted as isolated occurrences rather than evidence of sustained poleward stability.

Elevation imposes a secondary constraint that modulates CCF occurrence within this latitudinal window. CCF elevations increase with absolute latitude, while the occupied elevation range narrows toward both equatorward and poleward margins (Figure 4.7; Section 4.3.3). Elevation therefore appears to influence the conditions under which CCFs formed or survived without defining the overall latitudinal limits of occurrence.

Overall, CCF distribution reflects the combined influence of host-crater diameter, latitude, and elevation operating within the context of orbital forcing. The departure from proportional scaling with crater abundance confirms that CCF occurrence is neither random nor purely crater-controlled. What remains unresolved is whether the preserved distribution primarily records environments of initial palaeo-CCF accumulation or environments that favoured survival through subsequent interglacial modification. This question motivates the work in Chapters 5, 6 and 7, which investigate CCF depth, volume, slope, and aspect to gain insight on how host-crater geometry and geographic gradients influenced CCF glacial evolution.

4.4.2 Distributions of CCFs and other VFFs

Comparing the spatial distributions of CCFs and non-CCF VFFs provides a means of assessing whether these landforms occupy comparable geographic domains under shared climatic and topographic constraints. Here, I evaluate whether CCFs coincide with regions of enhanced LDA, LVF, and GLF occurrence in the mid-latitude belts. The idea is that mismatch between these landforms' distributions should reflect differences in their glacial histories, although any apparent spatial agreement between CCFs and other VFFs requires careful interpretation. Today's extent of LDAs, LVFs, and GLFs is strongly conditioned by the availability of steep slopes that permit ice accumulation and/or preservation. By contrast, CCFs—being entirely crater-confined—could theoretically develop and/or preserve within any suitably sized and morphologically intact crater, independent of regional gradient.

The spatial distribution of high confidence CCFs shows broad, though not universal, overlap with regions of high non-CCF VFF occurrence (Figures 4.1; 4.12). Zones of elevated CCF abundance and percentage occupation (Figures 4.8; 4.11, black outlined regions) coincide with many VFF-rich regions

identified in previous studies, including Alba Patera, Tempe Terra, Arabia Terra, Nilosyrtris Mensae, Phlegra Montes, the northern flanks of Acheron Fossae, and Arcadia Planitia in the north, and Terra Sirenum, the Thaumasia Highlands, Noachis Terra, Hellespontus Montes, Promethei Terra, and Terra Cimmeria in the south (e.g., Head et al., 2006; Levy et al., 2007, 2009, 2014; Morgan et al., 2009; Pedersen et al., 2010; Head and Marchant, 2010; Hubbard et al., 2011; Souness et al., 2012; Orgel et al., 2019; Gallagher et al., 2021; Driver et al., 2024). This spatial coincidence is consistent with regions that experienced favourable conditions for ice emplacement and/or preservation during mid-latitude glaciation (e.g., Forget et al., 2006; Madeleine et al., 2009) and supports a shared glacial context among VFF subtypes. In these settings, the climatic and orbital regimes that enabled ice stability on open slopes hosting LDAs, LVFs, and GLFs would also have supported the retention of crater-confined ice within steep crater interiors.

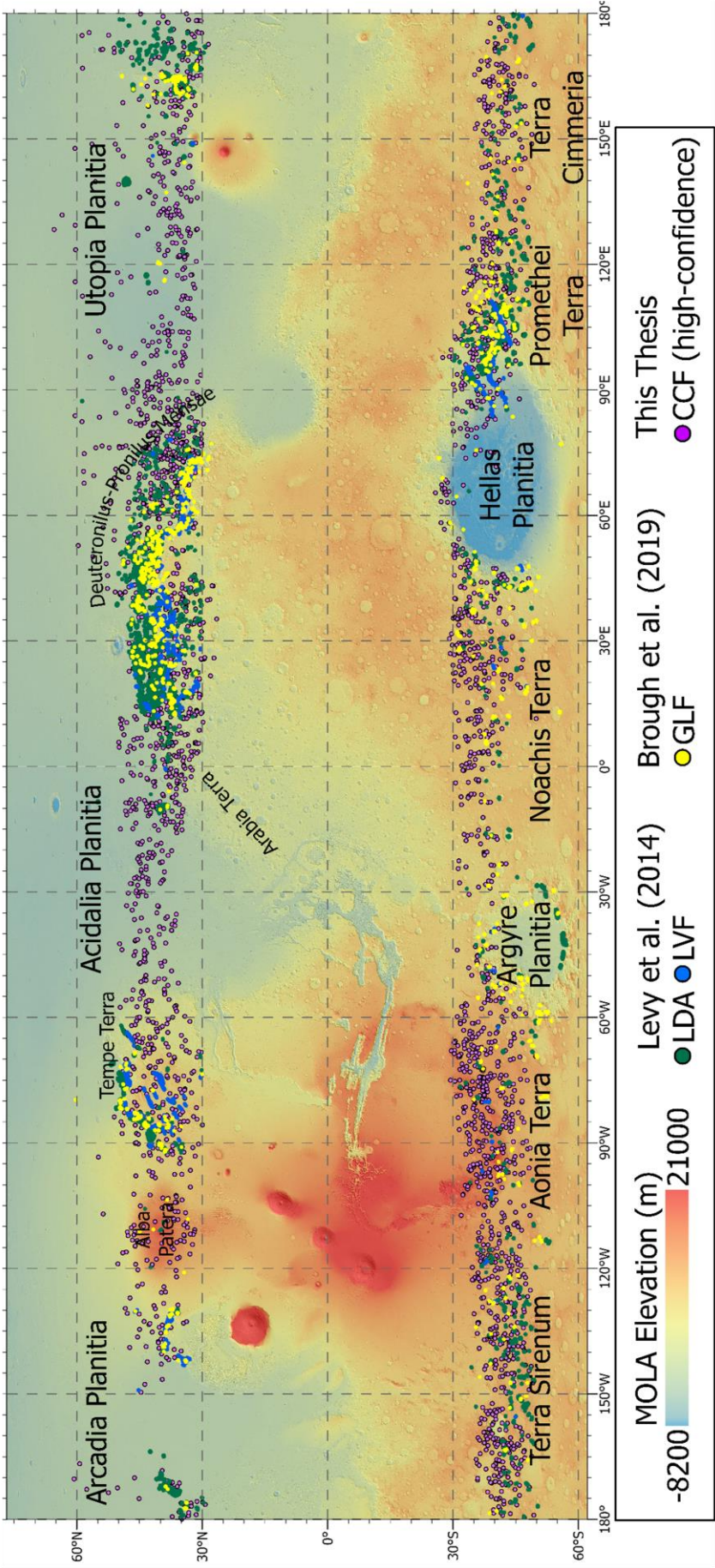


Figure 4.12. Global distribution of VFFs over MOLA topography. Map showing CCFs (this chapter), together with LVFs and LDAs from Levy et al. (2014) and GLFs from Souness et al. (2012), overlaid on MOLA gridded topography and MOLA-derived hillshade.

Despite the broad overlap, several regions show pronounced mismatch between CCF occurrence and non-CCF VFF abundance. The eastern and northern margins of Argyre Planitia and the Deuteronilus–Protonilus Mensae region host dense assemblages of LDAs and LVFs but contain few or no CCFs. In Deuteronilus–Protonilus Mensae, many filled craters are incorporated into integrated LDA and LVF systems (e.g., Morgan et al., 2009), where valley glaciers extend into or through crater interiors, limiting the development or preservation of discrete crater-confined deposits. Two such examples in Arabia Terra illustrate this ambiguity (Figure 4.13): both craters exhibit CCF-like surface textures, yet both are intersected by an LVF that continues beyond the crater walls, indicating that they form part of a connected flow system rather than discrete crater-confined deposits. Consistent with this, neither feature is included in the CCF inventory of Levy et al. (2014), and the western crater fill is mapped in their LVF inventory. Genetically, it remains unclear whether such fills formed by inflow from the regional VFF system or whether they initially accumulated as independent crater-confined bodies and were later captured and overrun. Although this distinction is potentially important for reconstructing local glacial histories, it likely cannot be resolved from morphology alone and is not pursued further in this study.

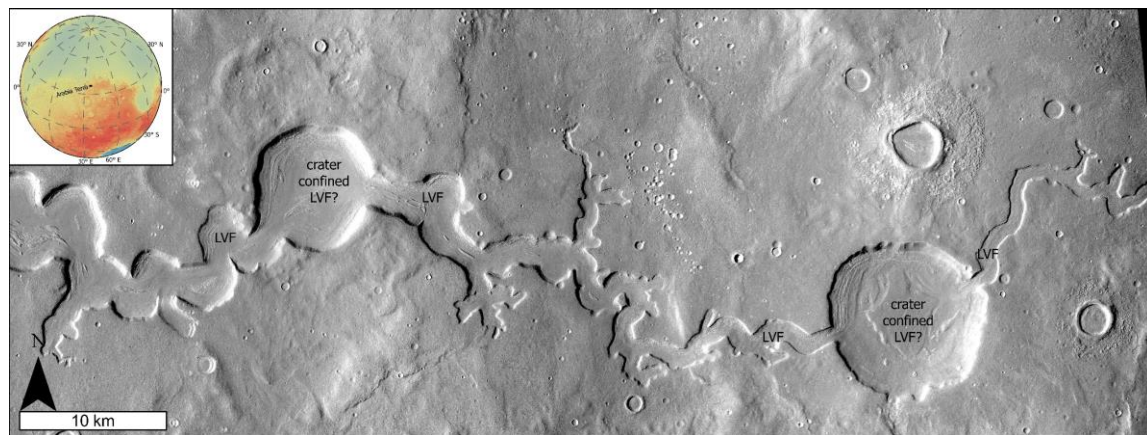


Figure 4.13. Integrated LVF-crater systems in Arabia Terra. CTX mosaic showing a system of LVF intersecting and breaching two filled impact craters. While the fill within these craters meets the morphological criteria for CCFs (e.g., complete floor coverage surface tracelines, Chapter 3), the clear connectivity with the broader LVF system suggests a shared evolutionary history rather than independent accumulation. Inset shows regional context within Arabia Terra (36.62° S, 36.24° E; CTX images: P22_009825_2173, D22_035815_2170, P18_008111_2165).

The southern and western margins of Argyre similarly contain abundant non-CCF VFFs but sparse CCF occurrence. Conversely, several regions display high CCF abundance despite sparse representation of other VFF subtypes, most notably across the northern lowlands, including Arcadia Planitia, Alba Patera, and Utopia Planitia. These contrasts are consistent with the strong slope dependence of non-CCF VFFs, which restricts their formation and preservation to topographic settings rarely found across the lowlands.

4.4.3. CCF density and mid-latitude glaciation GCMs

The spatial variability in CCF abundance and percentage occupation provides a way to assess whether contrasts in predicted mid-latitude net ice-accumulation zones from GCMs are reflected in equivalent contrasts in CCF abundance, specifically do areas of high predicted ice accumulation match areas of high densities of CCF? Here, I address this question by comparing global CCF abundance patterns with the intermediate-obliquity simulations of Madeleine et al. (2009), generated for 35° obliquity, 0.1 eccentricity, and perihelion at $L_p = 90^\circ$ and 270° . CCF abundance is used rather than percentage occupation because the higher southern-hemisphere crater density reduces the normalised signal to a few small, isolated high-occupation areas (Section 4.3.4). The broader regional CCF concentrations retained in the abundance maps therefore provide a more consistent basis for comparison with the spatial extent of GCM-predicted accumulation zones in both hemispheres.

These simulations produce spatial fields of net accumulation under one plausible Amazonian climate state, offering a reference frame for evaluating whether dense CCF populations coincide with modelled ice-rich regions. For each hemisphere I compare CCF spatial distribution and areas of high abundance (>6) to the GCM. Agreement in location between areas of high CCF abundance and simulated accumulation maxima would indicate that CCF distribution primarily records palaeo-accumulation patterns. Mismatch between those areas may point to a stronger influence of post-glacial ablation and differential preservation, or formation under orbital configurations not represented by the GCM simulation. Since the Madeleine et al. (2009) simulations represent a single (past) snapshot within a broad obliquity–eccentricity parameter space, whereas the mapped CCF distribution is itself a present-day snapshot of a time-integrated and preservation-filtered population that formed under multiple Amazonian orbital configurations (Fastook and Head, 2014), only partial concordance between modelled accumulation maxima and observed CCF abundance is expected.

The northern hemisphere shows a mixed but structured correspondence between modelled accumulation and observed CCF occurrence (Figure 4.14). Several broad zones of CCF distribution overlap regions of enhanced ($>10 \text{ mm yr}^{-1}$) simulated accumulation, most notably in Tempe Terra and Nilosyrtris Mensae, including the eastward extension of the accumulation maximum from Nilosyrtris into western Utopia Planitia between $\sim 60^\circ\text{E}$ and $\sim 100^\circ\text{E}$, where CCFs also extend to more poleward latitudes. This sector therefore represents one of the clearest cases of first-order spatial agreement between predicted accumulation and observed crater glaciation. By contrast, substantial CCF populations also occur in regions where the GCM predicts little or no net accumulation, particularly around the southern margin of Utopia Planitia, Alba Patera, and Acidalia Planitia, indicating that the overall CCF footprint is broader than the principal modelled accumulation zones. When abundance is considered, the correspondence becomes more selective (Figure 4.14B). Several of the highest-abundance CCF clusters—most notably Tempe Terra, Nilosyrtris Mensae, Coloe Fossae, and Phlegra

Montes—coincide closely with simulated accumulation maxima. In contrast, high CCF abundance in Alba Patera, Arabia Terra, and Arcadia Planitia occurs largely outside the main accumulation cores. Thus, in the north, while the gross CCF distribution only partly matches the GCM fields, the strongest concentration centres show a clearer, though still incomplete, spatial association with predicted accumulation.

The southern hemisphere exhibits a substantially weaker correspondence between GCM-predicted accumulation and both the distribution and abundance of CCFs (Figure 4.15). Some overlap is present, particularly in the Thaumasia Highlands, Noachis Terra, and parts of Terra Cimmeria and Promethei Terra, where CCFs coincide with simulated accumulation maxima. However, the dominant feature of the southern hemisphere comparison is a systematic spatial mismatch. This mismatch is most clearly expressed as a latitudinal offset of approximately $\sim 10^\circ$, best illustrated in Terra Sirenum. There, the GCM predicts peak net accumulation of $\sim 14\text{--}16\text{ mm yr}^{-1}$ between $\sim 40^\circ\text{S}$ and 60°S , whereas CCFs are largely restricted to $\sim 35\text{--}50^\circ\text{S}$, with the main abundance maxima centred closer to $\sim 40^\circ\text{S}$. More generally, the GCM predicts its strongest accumulation between $\sim 50^\circ$ and 60°S , yet CCFs are almost entirely absent poleward of $\sim 50^\circ\text{S}$ and instead cluster equatorward of this limit. This offset is visible both in the raw CCF distribution (Figure 4.15A) and in the abundance field (Figure 4.15B).

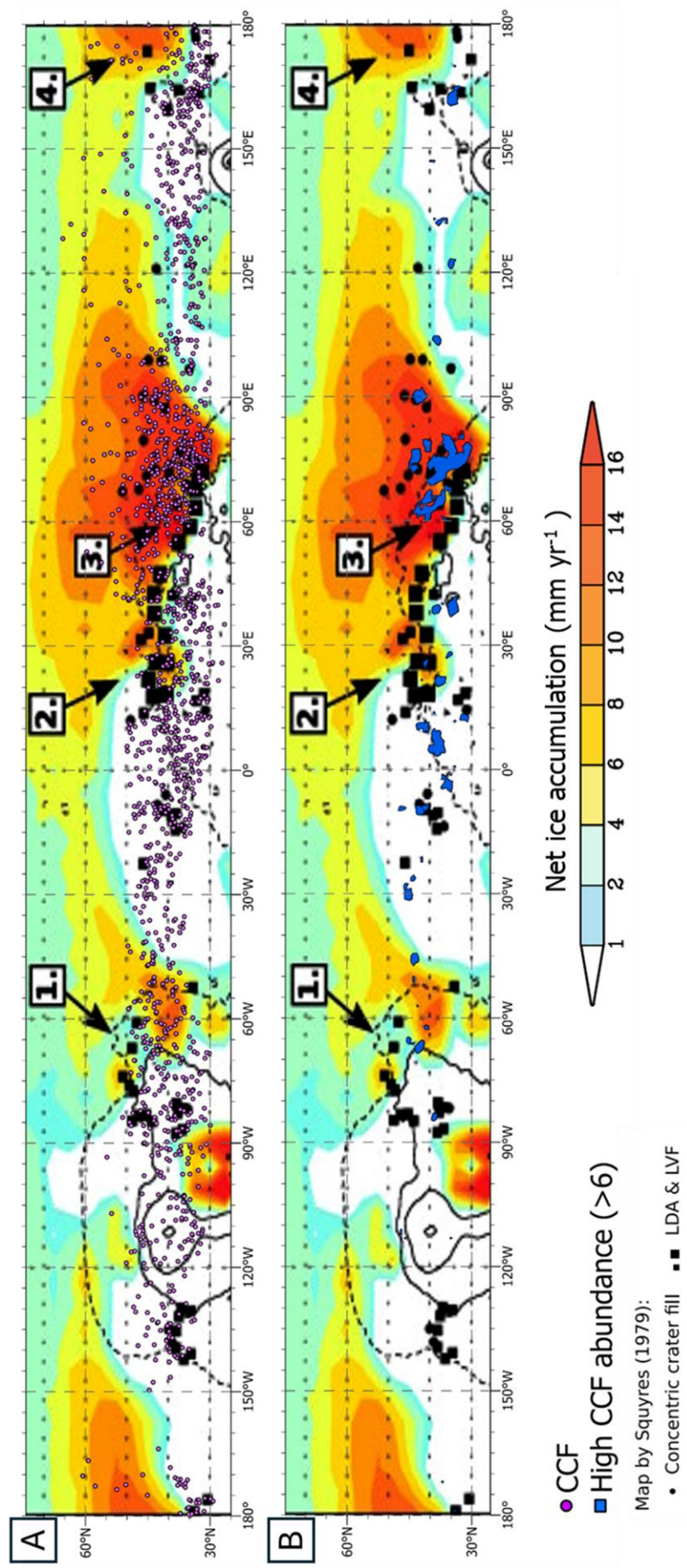


Figure 4.14. Northern hemisphere CCF distribution compared with GCM-predicted net ice accumulation. (A) Distribution of CCFs (purple) superposed on the mid-latitude net ice accumulation field (mm yr⁻¹) from the GCM simulations of Madeleine et al. (2009). (B) As in (A), but showing only grid cells with high CCF abundance (>6). The climate simulation uses an orbital configuration with obliquity $\theta = 35^\circ$ and eccentricity $\varepsilon = 0.1$, with perihelion at $L_s = 270^\circ$ for the northern hemisphere, dust opacity of 2.5, and an ice source derived from sublimation of tropical mountain glaciers. The background field shows the modelled net ice accumulation rate.

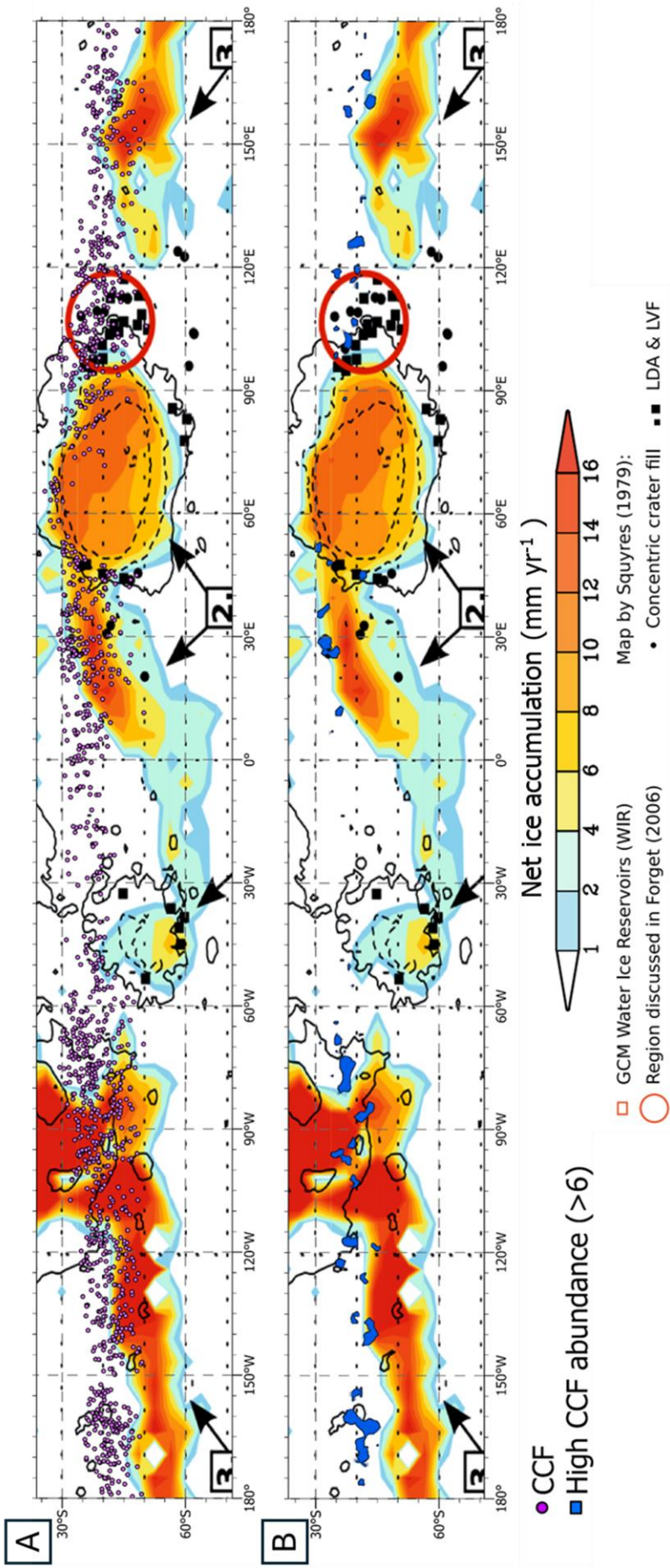


Figure 4.15. Southern hemisphere CCF distribution compared with GCM-predicted net ice accumulation. (A) Distribution of CCFs (purple) superposed on the mid-latitude net ice accumulation field (mm yr⁻¹) from the GCM simulations of Madeleine et al. (2009). (B) As in (A) but showing only grid cells with high CCF abundance (>6). The climate simulation uses the same orbital configuration as in Figure 4.14 ($\theta = 35^\circ$, $\varepsilon = 0.1$), but with perihelion at $L_s = 90^\circ$ for the southern hemisphere, dust opacity of 2.5, and an ice source derived from sublimation of tropical mountain glaciers. The background field shows the modelled net ice accumulation rate.

The southern hemisphere therefore shows the strongest disagreement between observed high-confidence CCF distribution and abundance with modelled net accumulation, implying that favourable accumulation conditions at high southern latitudes were not, by themselves, sufficient to ensure long-term preservation of crater-confined ice. One possibility is that many craters within these high-accumulation zones formerly hosted CCFs that were subsequently removed by interglacial ablation or buried beneath younger latitude-dependent mantle deposits (e.g., Head et al., 2003). Alternatively, the offset may reflect limitations in the GCM representation of accumulation patterns under Amazonian climate states. More generally, however, such mismatches most plausibly reflect the difficulty of interpreting a multimillion-year glacial record using a single orbital configuration. CCFs are widely thought to be hundreds of Myr old (Kreslavsky and Head, 2006; Levy et al., 2010; Butcher et al., 2025) and therefore formed and evolved under a broad range of obliquity states and orbital conditions (Forget et al., 2006; Mischna et al., 2013; Fastook and Head, 2014), such that their present distribution is best interpreted as the time-integrated outcome of ice emplacement—either during single or multiple episodes (Section 2.3.2)—and subsequent modification across multiple orbital states.

Recent GCM developments incorporating radiatively active clouds, cloud microphysics, and latent-heat exchange (e.g., Vos et al., 2024; Clément et al., 2024) indicate that these processes can substantially alter simulated vapour transport and surface energy balance and may therefore help resolve some of the spatial mismatches identified here. Radiatively active clouds and latent-heat exchanges, in particular, are predicted to modify surface temperatures, enhance northward vapour transport and strengthen albedo–thermal-inertia feedback, collectively shifting the spatial pattern of simulated net ice accumulation (Vos et al., 2024; Clément et al., 2024). It remains to be seen whether GCMs fully incorporating these refinements can reconcile the discrepancies between present model outputs and the observed CCF distribution.

Overall, while many CCF clusters coincide with regions of modelled mid-latitude ice accumulation, the incomplete correspondence—most evident in the southern hemisphere—shows that present-day CCF distribution and density are not directly predicted by accumulation potential under a single GCM climate state. Instead, CCF occurrence reflects the combined influence of orbital variability, resurfacing history, and differential preservation superimposed on their underlying accumulation signal.

4.4.4. CCF abundance and absence

The spatial distribution of CCFs provides constraints on the climatic and geomorphic conditions that governed mid-latitude ice accumulation and/or preservation on Mars. The results presented above demonstrate that variations in the underlying crater population weakly control the observed spatial variability in CCF occurrence. Regional differences in CCF abundance and percentage occupation are therefore most plausibly attributed to differences in glacial accumulation, post-glacial modification, or an interaction between these processes, rather than to impact crater availability alone.

Interpreting these distributions requires consideration of where CCFs occur frequently, where they occur infrequently, and where they are apparently absent, as each case places different constraints on mid-latitude glacial history. Regions of high CCF abundance or high percentage occupation likely correspond to settings that supported favourable mass-balance conditions, either through enhanced net accumulation during glacial intervals or reduced ablation during subsequent retreat. Regions characterised by low CCF abundance or low occupation may reflect environments where initial accumulation was limited, where post-glacial mass loss was more effective, or where CCFs exist but are obscured beneath younger mantling deposits that inhibit confident identification within the present dataset. Regions of apparent CCF absence require separate consideration. Absence may indicate genuinely unfavourable conditions for CCF formation or preservation, but may also arise from a scarcity of suitable host craters or from classification limitations introduced by extensive surface mantling masking underlying CCF deposits and preventing their inclusion in the inventory. The following sections adopt this framework by examining regions of CCF absence, regions of low CCF abundance, and regions of high CCF abundance in turn, and situating each within its broader regional glacial and geomorphic context.

4.4.4.1. Areas of CCF absence

Although CCFs occur widely across the mid-latitudes, several extensive regions exhibit persistent absence or strongly reduced occurrence. These zones are concentrated primarily within areas of very low elevation, including the northern lowlands (Utopia, Acidalia, and Arcadia Planitia) and the major southern impact basins of Argyre and Hellas (Figure 4.16). This section examines these regions to assess whether CCF absence reflects genuine constraints on palaeo-CCF formation or subsequent preservation, or instead arises from resurfacing, mantling, or classification limitations.

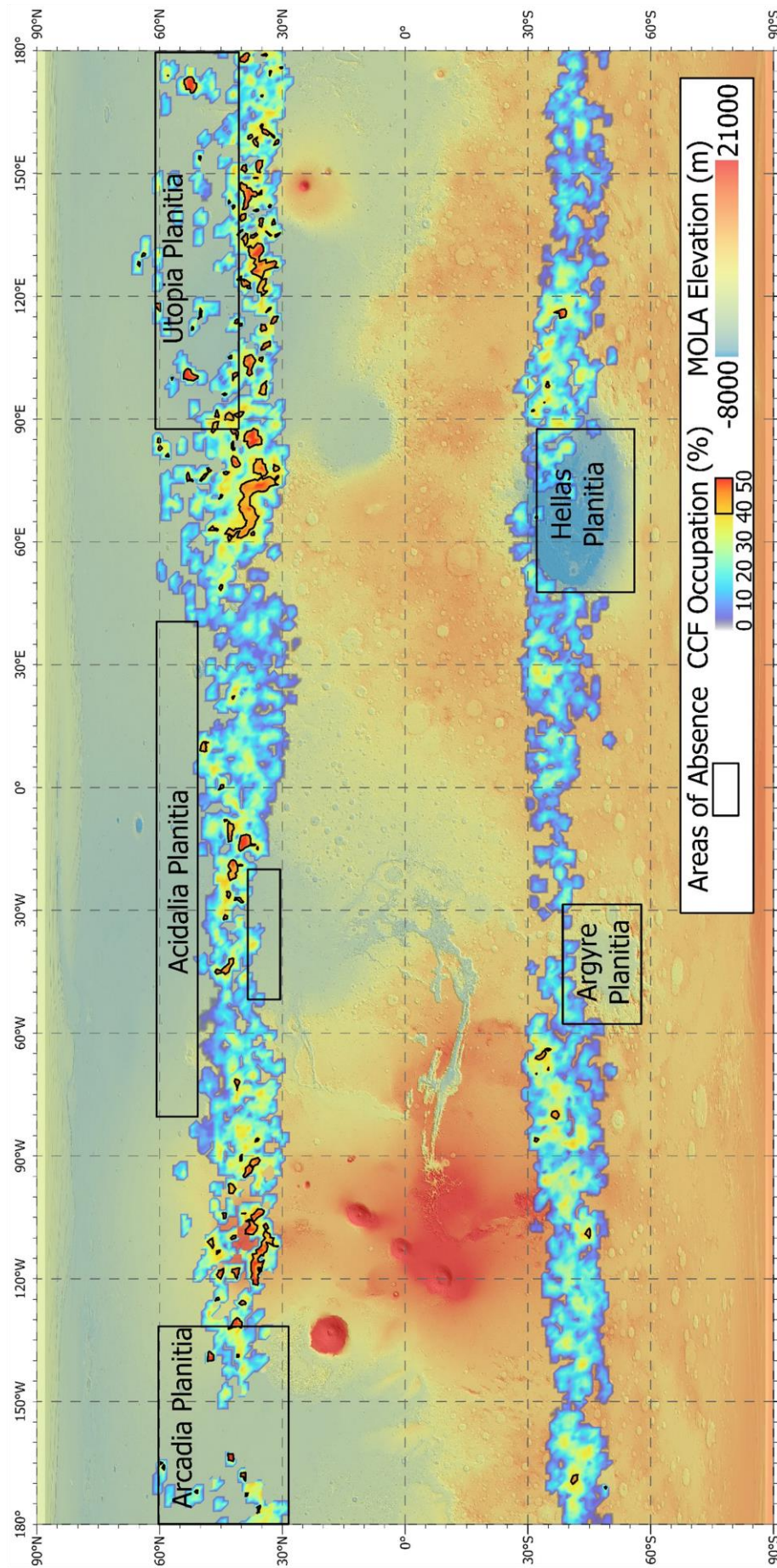


Figure 4.16. Regions of CCF absence. Heatmaps showing CCF occupation (%) in the northern and southern hemispheres, overlaid on MOLA gridded topography with MOLA-derived hillshade. Black boxes highlight regions characterised by a notable absence of CCFs.

In the northern hemisphere, the most extensive CCF absences occur across the northern lowlands—Utopia, Arcadia, and Acidalia Planitia—where CCF occurrence becomes spatially sparse and locally restricted (Figure 4.16). In central Utopia Planitia, reduced CCF identification coincides with widespread infilling of crater interiors by ice-rich deposits. SHARAD observations reveal subsurface layers 80–100 m thick containing up to ~85 vol.% water ice along the western basin margin (Stuurman et al., 2016; Séjourné et al., 2019), while central Utopia hosts thicker and more laterally extensive mantle deposits (Xu et al., 2025). Textured terrains attributed to Late Amazonian ice-rich mantling are near-continuous, and many craters exhibit “*concentric crater hump*” morphologies—heavily infilled crater forms with elevated interiors rising above degraded rims (Xu et al., 2025). Such modification would obscure any underlying CCF morphologies, plausibly accounting for their limited recognition across central Utopia. By contrast, elevated CCF occurrence along the outer margins of Utopia, where mantling is thinner and surface modification is reduced, is consistent with preferential preservation in less resurfaced terrain.

In Acidalia Planitia, CCFs occur only between ~35° and 50° N, a distribution inherited from the source inventory (Levy et al., 2014). Their absence at higher latitudes coincides with a continuous expanse of layered and textured terrain extending from ~40° to 78° N (Orgel et al., 2019), surfaces that likely obscure crater-confined deposits. The absence of CCFs at the equatorward margin of Acidalia (<35° N) remains unresolved.

In the southern mid-latitudes, CCF occurrence is more spatially continuous than in the north but is interrupted by major impact basins, most notably Argyre and Hellas, by western Noachis Terra, and by highland terrain poleward of ~55° S (Figure 4.16). This distribution aligns with regional heterogeneity reported by Levy et al. (2014), who identified pronounced scarcity within central Argyre and a decline toward higher southern latitudes. The present dataset reproduces these broad gaps but resolves a sharper southern limit, with no CCFs identified beyond ~55° S, where earlier inventories reported only a small number of isolated examples.

The high southern mid-latitudes therefore represent extensive terrain where CCFs are absent, yet non-CCF VFFs remain widespread. LDAs and LVFs are well documented along the margins of Argyre Basin, across western Noachis Terra, within Aonia Terra (Levy et al., 2014), and as isolated GLFs along the eastern interior wall of Lowell crater (Souness et al., 2012). These occurrences confirm that mid-latitude glaciation operated within these regions, raising the question of why CCFs are not preserved. Perhaps initial glaciation in these high southern latitudes did not promote the development of palaeo-CCFs if accumulation was too limited to sustain crater-confined ice or so extensive that palaeo-CCFs became embedded within larger LDA–LVF systems rather than isolated ice bodies. Alternatively, subsequent

resurfacing and mantling may have modified or buried earlier CCFs, leaving only non-crater-confined expressions of viscous flow visible today.

This contrast between widespread non-CCF VFFs and the absence of CCFs is particularly clear around Argyre Basin. Along Nereidum Montes, the mountainous northern rim hosts abundant VFFs and thick mantling deposits identified as ice-rich materials (Milliken et al., 2003; Berman et al., 2021; Butcher et al., 2024). The north-western sector of Nereidum contains some of the highest concentrations of well-preserved southern mid-latitude VFFs (Berman et al., 2021), many exhibiting surface textures indicative of post-depositional mantling. Despite these ice-rich conditions, CCFs occur only along the western portion of Nereidum (Figure 3.9), whereas non-CCF VFFs dominate the eastern rim and adjacent Charitum Montes. This west–east asymmetry is consistent with spatial variation in mantling thickness, with heavier depositional cover across the eastern basin rim likely obscuring or eliminating crater-confined morphologies. Examples from western Nereidum Montes illustrate how crater interiors in this region can be completely infilled yet retain no diagnostic CCF surface expression. Some craters contain lobate deposits superposed on more subdued, older interior fills, while others are entirely reworked by aeolian bedforms that obscure any underlying glacial textures if they exist at all (Figure 4.17). Perhaps these surface processes have either overprinted or buried palaeo-CCFs, rendering them morphologically subdued and unrecognisable in CTX images alone. These observations suggest that within heavily mantled and degraded southern high-mid-latitude regions that host abundant non-CCF VFFs, CCF absence could be more plausibly linked to this partial burial and resurfacing than to a fundamental absence of crater-confined glacial activity.

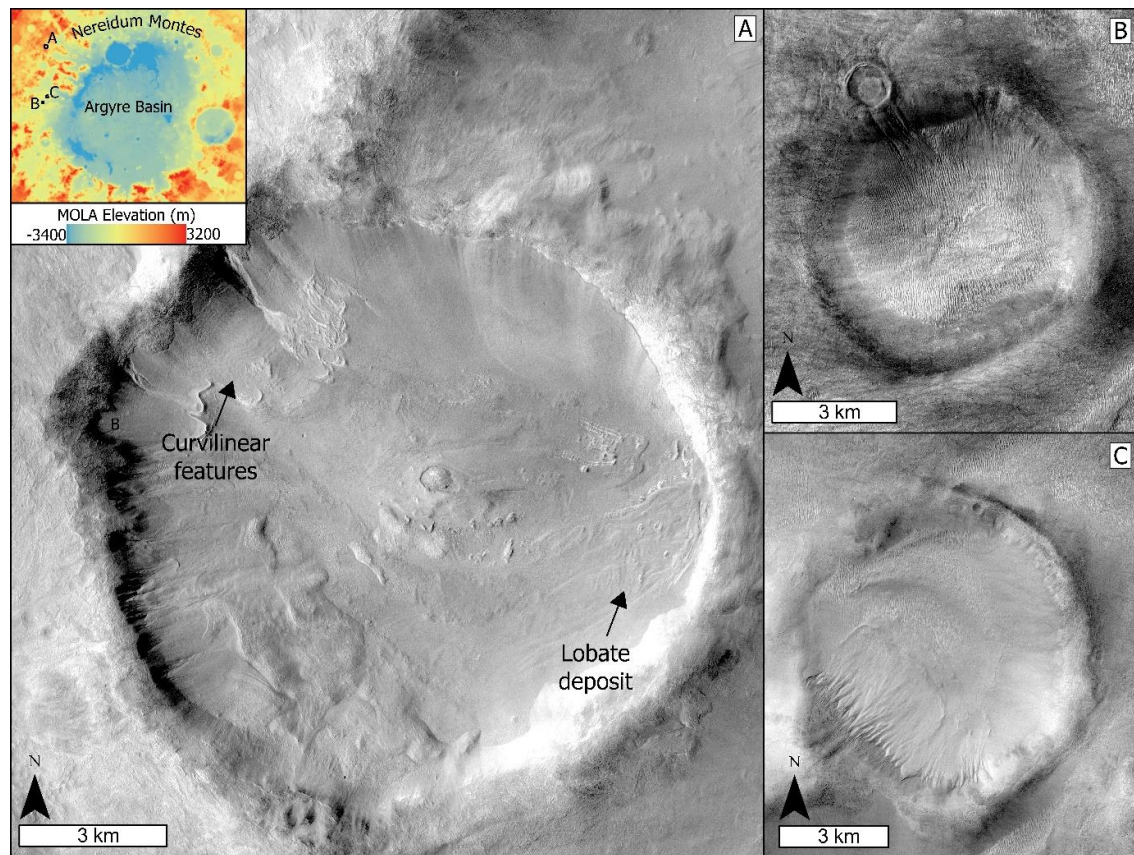


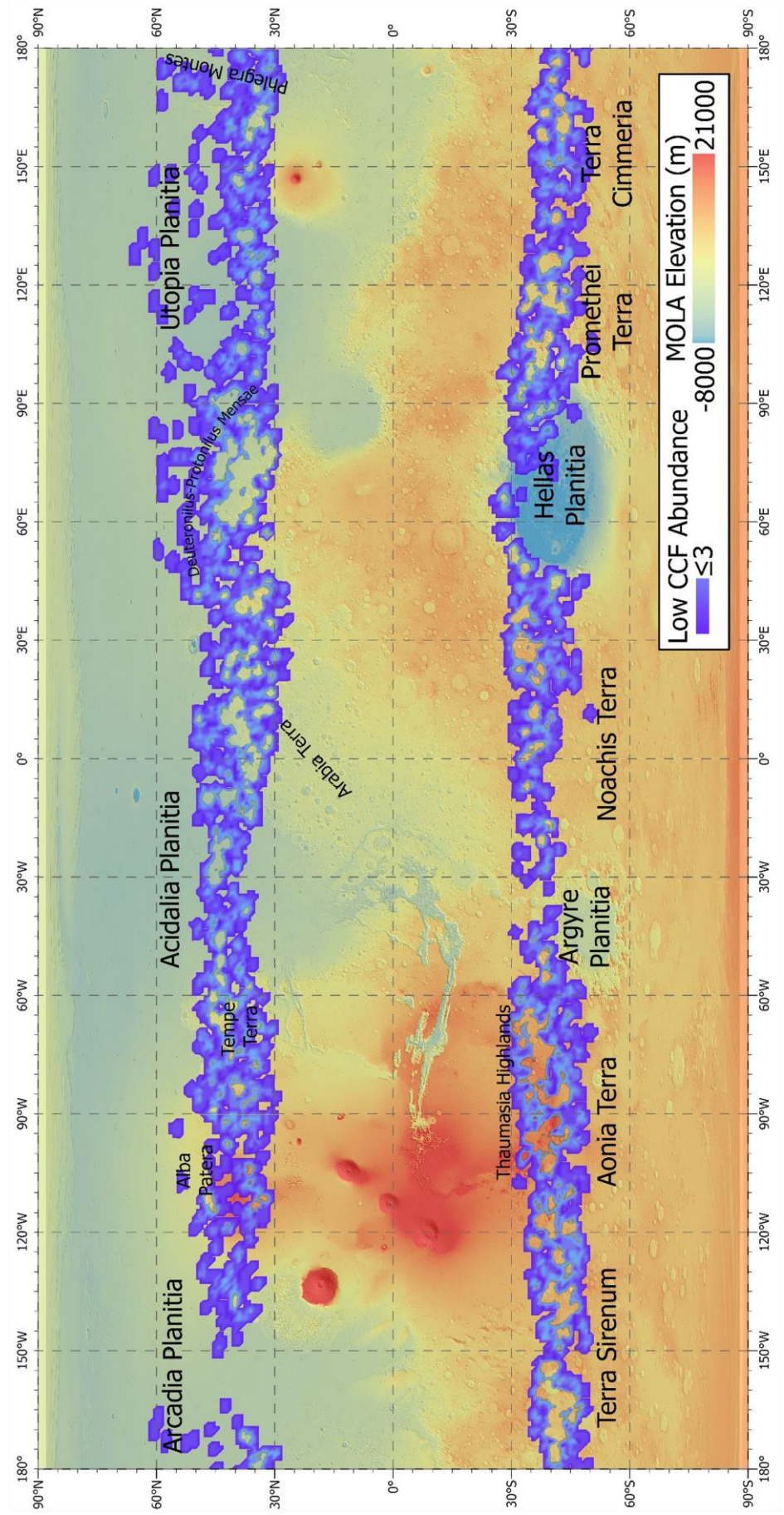
Figure 4.17. Examples of infilled and mantled craters in western Nereidum Montes. Three neighbouring craters showing complete interior infill but strong surface modification. (A) Crater filled by a lobate debris apron superposed on an older, more subdued interior deposit, perhaps indicating multiple phases of emplacement or modification (44.02° S, 53.13° W; CTX images: B12_014180_1351, B10_013468_1357, B16_016013_1366). (B–C) Two filled craters whose surfaces are extensively reworked by aeolian bedforms, obscuring any inference of their buried fill (B: 48.88° S, 54.43° W; CTX images: G11_022527_1309 C: 48.39° S, 53.73° W; CTX image: D13_032219_1308)

4.4.4.2. Areas of low CCF abundance and occupancy

Areas of low CCF abundance and occupancy are difficult to isolate due to methodology choices masking genuinely sparse regions. Apparent zones of low CCF occupancy across the high-latitude northern lowlands of Utopia, Arcadia, and Acidalia Planitia largely reflect this effect, however, these patches also possess low crater count (Figure 4.18). Beyond these methodological artefacts, the northern hemisphere contains limited low-CCF occupancy and abundance zones in Arabia Terra and parts of Tempe Terra. The southern hemisphere shows comparable pockets of low values in Terra Sirenum, Noachis Terra, and Terra Cimmeria.

Regions characterised by persistently low CCF abundance and occupancy are uncommon. Instead, where CCFs occur in isolation from major clusters (e.g. central Utopia Planitia), they typically form small clusters, occupying multiple neighbouring craters rather than as single, isolated occurrences. This

spatial organisation indicates that CCF formation and/or preservation was rarely governed by highly localised crater-scale microenvironments alone. Rather, the co-occurrence of multiple CCFs within the same geographic setting implies control by regional-scale glacial or mass-balance conditions capable of sustaining palaeo-CCF accumulation across several craters simultaneously. If crater-scale factors alone dominated CCF viability, widespread isolated occurrences or uniformly sparse clusters would be expected. Their limited presence instead supports the interpretation that CCF distribution reflects broader environmental regimes, motivating further examination of latitudinal and elevational controls in subsequent chapters.



4.18 Heatmap of low CCF abundance. Heatmap highlighting notable areas of low CCF abundance (≤ 3) overlaid on MOLA gridded topography and MOLA-derived hillshade.

4.4.4.3. CCF high occupation

High CCF percentage occupation identifies regions where palaeo-CCFs either experienced greater net ice accumulation—through greater net supply or greater duration—or experienced comparatively reduced post-glacial mass loss, resulting in enhanced long-term preservation. These regions therefore provide a basis for interrogating the local conditions under which CCFs were preferentially retained. Several high-occupation CCF clusters identified in Section 4.3.4 coincide with landscapes known to host extensive ice-rich LDM, such as Arcadia Planitia (e.g., Bramson et al., 2015; Ramsdale et al., 2019), Nilosyrtris Mensae (Head et al., 2015), and the outer margins of Utopia Planitia (Stuurman et al., 2016). Furthermore, these clusters coincide with occurrences of younger superimposed flow features/LDAs such as in Alba Patera (Sinha and Murty, 2013; Sinha and Vijayan, 2017; Schiff et al., 2022). These settings contrast with other northern regions of central Utopia, northern Arcadia, and high-latitude Acidalia, where thick Late Amazonian mantling obscures crater interiors and corresponds to zones of CCF absence (Section 4.4.4.1).

The coexistence of high CCF occupation with widespread LDM and superimposed small-scale LDAs complicates the expected anticorrelation between extensive younger superimposing deposits and the preservation of older CCF surface morphology needed for their identification. While a systematic evaluation of all such regions lies beyond the scope of this chapter, Alba Patera provides a representative example that illustrates how high-confidence CCF populations can inform regional glacial histories.

Alba Patera (Figure 4.19A) is widely interpreted as an ice-rich volcanic edifice shaped by relatively young Late Amazonian glacial activity (e.g., Sinha and Murty, 2013; Sinha and Vijayan, 2017). Numerous pole-facing flow features, dated to ~10–100 Ma, have been attributed to localised snow and ice accumulation under microclimatic conditions on crater walls and alcoves (Levy et al., 2007; Dickson et al., 2008; Morgan et al., 2009; Sinha and Vijayan, 2017). Despite this extensive younger glacial deposition, the southern flanks of Alba Patera retain a high abundance and occupation of CCFs, identified both in this chapter (Figure 4.8 & 4.11, black outlined regions; Figure 4.19A) and in previous mapping by Sinha and Vijayan (2017), who documented 137 CCFs alongside 346 pole-facing flow features. This juxtaposition raises a key question: why did some craters maintain stable, near-rim-full CCFs while adjacent craters—which clearly possess the climate-driven ability to accumulate ice—host only small-scale flows atop deep, degraded floor deposits?

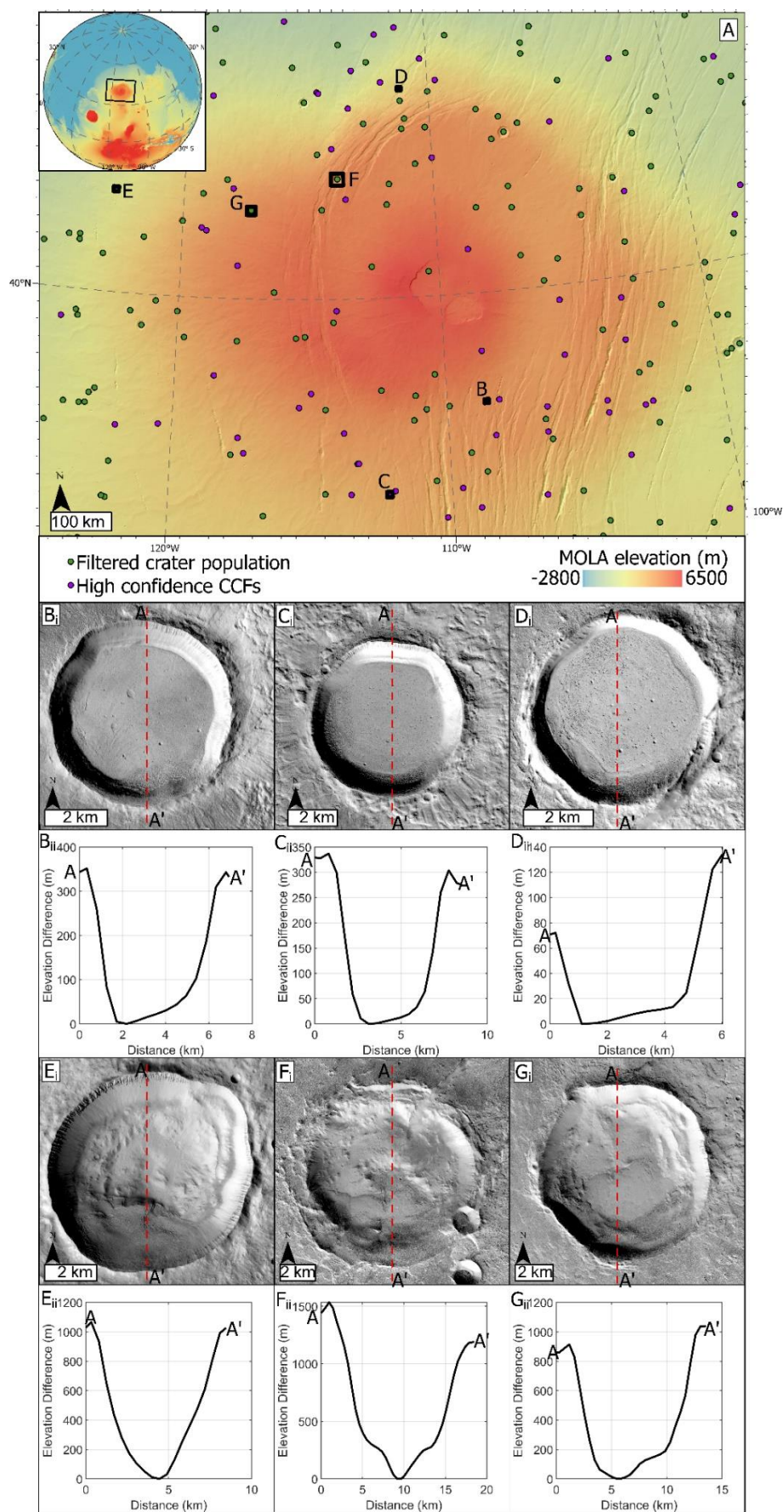


Figure 4.19. Comparison of elevation transects across CCFs and pole-facing flow features in Alba Patera
(Caption continued on next page).

(A) MOLA topography draped on hillshade showing the study area in Alba Patera. Green points mark the filtered crater population (Robbins and Hynek, 2012), and purple points indicate the subset hosting high-confidence CCFs. (B_i–D_i) CTX images of representative CCF-hosting craters within Alba Patera. Red dashed lines mark the locations of the elevation transects, which are shown in the corresponding profiles in panels B_{ii}–D_{ii}. (E_i–G_i) CTX images of pole-facing flow features mapped by Sinha and Vijayan (2017). Red dashed lines mark the locations of the elevation transects, shown in panels E_{ii}–G_{ii}. Image locations and sources: (B) 36.82° N, 108.57° W; CTX image: B16_016041_2171, (C) 34.28° N, 112.23° W; CTX image: B11_013984_2147, (D) 46.11° N, 111.17° W; CTX image: P16_007233_2267, (E) 43° N, 122.7° W; CTX image: B18_016503_2233, (F) 43.51° N, 113.94° W; CTX image: G14_023821_2238, (G) 39.92° N, 135.55° W; CTX image: P17_007669_2196.

Addressing this question requires consideration of the nature of the crater-floor deposits underlying the younger pole-facing flow features. The origin of these older crater fills is not definitively established; however, several morphological indicators—including faint curvilinear surface textures and polygonal cracking—are consistent with a glacial or periglacial origin (Sinha and Vijayan, 2017). If these underlying deposits are glacial, they may represent degraded palaeo-CCF ice bodies that either accumulated to smaller thicknesses or experienced substantially greater post-glacial mass loss than neighbouring, near-rim-full CCFs. Under this interpretation, the present juxtaposition reflects not a failure of some craters to ever accumulate ice, but strong heterogeneity in preservation and ablation histories at the crater scale.

A comparison of CCF surface depths and termini depths of flow features is consistent with this possibility (Figure 4.19). CCFs on the Alba Patera flanks are commonly near rim-full, with present-day surfaces ≤ 0.32 km below the rim in craters 6–18 km in diameter (Figure 4.19B–D). In contrast, pole-facing flows mapped by Sinha and Vijayan (2017) exhibit markedly deeper interior surfaces, with flow termini positioned ~ 0.9 –1.5 km below the crater rim in craters 8–18 km in diameter (Figure 4.19E–G). This contrast may partly reflect the influence of the host crater (diameter and depth) on the surface elevation of the ice deposit, examined in Chapter 5, but it more plausibly indicates differences in accumulation and/or post-glacial modification histories among otherwise comparable crater settings.

If the superimposed crater-floor deposits beneath the younger pole-facing flows are indeed degraded palaeo-CCFs, then the observed depth contrast implies strongly heterogeneous ablation histories operating at the scale of individual craters. Under this scenario, some palaeo-CCFs must have experienced substantially greater post-glacial mass loss—potentially driven by unresolved microclimatic or topographic controls—allowing crater interiors to deepen sufficiently for later slope-fed flow features to form, while neighbouring CCFs remained near-rim-full. This scenario further suggests that the emplacement or preservation of small-scale pole-facing flows may depend on crater interiors having thinned beyond a threshold depth not reached in well-preserved CCF-hosting craters. However, this scenario assumes all craters began with a similar initial ice volume. An alternative end-member interpretation is that this spatial heterogeneity is primary rather than secondary: some craters may never have accumulated substantial crater-filling ice during earlier glacial episodes, and only later developed flow features under more recent glacial episodes.

Distinguishing between these two histories—differential ablation versus differential initial accumulation—is complicated by the fact that variability in CCF surface depths has been recognised previously across Mars (e.g., Pedersen and Head, 2010; Levy et al., 2014; Figure 4.19B–D). However, whether such differences primarily reflect host-crater geometry or record spatial variability in past mass-balance forcing remains unresolved. The ambiguity of these local results motivates the analysis in Chapter 5, which expands the scope to the global high confidence CCF dataset to examine whether spatial variability in CCF surface depths records signatures of former mass-balance conditions rather than purely geometric controls.

4.5 Conclusion

This chapter presents the first global evaluation of how high-confidence CCFs are distributed relative to the mid-latitude crater population on Mars. By comparing CCF-hosting craters with non-hosting craters, the chapter establishes whether (and to what extent) CCF occurrence reflects impact crater availability or records environmental and geomorphological controls on their accumulation and preservation.

I draw the following conclusions which address the specific questions posed in Section 4.1.

1. CCF occurrence departs from background crater availability

CCF-hosting craters form a statistically distinct subset of the mid-latitude crater population. Spatial proportionality is tested directly by comparing CCF abundance against filtered crater abundance (Figure 4.11) for each grid cell. Although the two quantities are positively correlated in both hemispheres (Spearman's $\rho = 0.46$ in the north and $\rho = 0.42$ in the south; $p < 0.001$), the large scatter demonstrates that crater availability explains only a moderate fraction of the spatial variability in CCF abundance. Many cells with similar crater densities host very different numbers of CCFs, and comparable CCF abundances occur across a wide range of crater densities, showing that CCF occurrence does not scale proportionally with crater availability. This conclusion is supported by divergence analysis of the abundance distributions ($D_{KL} = 2.09$ in the north; 1.24 in the south), which confirms that the CCF-hosting population is statistically distinct from the background crater population. Thus, spatial variability in CCF abundance cannot be interpreted as a simple population artefact and instead likely reflects additional environmental or geomorphological controls on where crater-confined ice is able to accumulate and persist.

2. Host-crater diameter influences the likelihood of CCF preservation

CCF counts closely follow the strongly skewed diameter distribution of the mid-latitude crater population, confirming that small craters dominate CCF numbers primarily because they dominate crater frequency. However, when normalised to crater availability,

percentage occupation shows that in either hemisphere CCFs more frequently occupy intermediate-sized host craters between diameters of 4–12 km in the south; 4–18 km in the north (Figure 4.4G–I).

This result establishes host-crater diameter as a significant control on the preserved CCF record. Diameter directly constrains crater cavity volume and attainable CCF thickness, but likely also implicitly captures a suite of crater-scale influences, including relief magnitude, interior slope geometry, shadowing from the crater rim of the CCF surface and debris supply. As a result, host crater diameter acts not only as a geometric parameter but as a proxy for multiple physical host crater processes that influence both CCF formation and preservation. However, this inferred diameter control should be interpreted cautiously, because the Chapter 3 selection criteria may partly condition the observed diameter range by favouring craters large enough for coherent concentric tracelines to be identified while excluding smaller, less well-resolved examples.

The broader diameter interval associated with elevated CCF occupation in the northern hemisphere suggests hemispheric differences in either accumulation magnitude, preservation efficacy, or the interaction between crater geometry and mass-balance forcing, although the specific mechanism cannot be isolated here. Crucially, this finding demonstrates that host-crater effects must be explicitly separated from geographic and climatic variables in subsequent analyses if genuine environmental signals in CCF morphology and evolution are to be resolved.

3. Latitude defines the primary climatic window for CCF occurrence, with elevation acting as a secondary modulator within that window

CCFs are concentrated predominantly between $\sim 30^\circ$ and 45° absolute latitude in either hemisphere (Figure 4.5), defining a latitudinal band within which crater-confined ice formation and/or long-term preservation was viable. This clustering broadly coincides with mid-latitude net ice accumulation zones predicted under past orbital configurations (e.g., Madeleine et al., 2009; Figures 4.15, 4.16), indicating that latitude exerted a first-order climatic control on where CCFs could persist. Although the northern hemisphere distribution extends further poleward, occurrences beyond $\sim 50^\circ$ N are rare and geographically restricted, indicating that they represent isolated cases rather than a sustained poleward expansion of favourable conditions. The absence of a proportional southern excess of CCFs, despite substantially higher crater densities at equivalent latitudes, further confirms that CCF occurrence is not governed by crater availability alone but is conditioned by latitude-dependent environmental constraints.

Within this latitudinal window, elevation exerts a secondary, modulating influence. Across central mid-latitudes, CCFs occupy the full local elevation range (Figure 4.7), indicating that elevation did not limit crater-confined ice stability under optimal climatic conditions.

Toward both equatorward and poleward margins, however, the occupied elevation range contracts systematically, implying that elevation-dependent factors increasingly constrained either accumulation or preservation near the climatic limits of CCF stability. These results demonstrate a hierarchical climatic control on CCF distribution, in which latitude sets the broad window for crater-confined glaciation, while elevation seemingly governs where within that window CCFs could persist.

4. CCF distribution reflects distinct glacial constraints and preservation filters

High confidence CCF occurrence shows broad but incomplete spatial correspondence with regions rich in LDAs, LVFs, and GLFs (Figure 4.12), further indicating that these landforms share a common mid-latitude glacial context. Where CCFs and other VFFs coexist, it suggests that mid-latitude glaciation was sufficient to sustain slope-fed viscous flow, yet either failed to integrate CCFs into larger systems (e.g., Figure 4.13) or was followed by post-glacial mass loss that re-isolated them. Conversely, areas of high CCF abundance where other VFFs are sparse—particularly across the northern lowlands—highlight a reduced dependence on regional slope availability.

Regions lacking CCFs do not necessarily indicate glacial inactivity, but rather the limits of preservation. Areas of CCF absences are consistently associated with thick Late Amazonian mantling, resurfacing, or landscapes where CCFs were likely incorporated into larger, integrated flow systems. In several southern high mid-latitude regions where LDAs and LVFs remain abundant, the lack of CCFs implies that crater-confined morphologies were subsequently buried, degraded, or masked by diagnostic concentric structures (Figure 4.17). Ultimately, the present CCF inventory does not purely record where crater-confined ice accumulated; it represents the subset of crater-confined ice that has remained morphologically identifiable through subsequent modification, highlighting post glacial preservation as a key filter on the spatial signal retained by CCFs.

5. Regions of high CCF percentage occupation highlight preferential preservation and hemispheric asymmetry

Regions of high CCF percentage occupation identify locations where crater-confined ice was proportionally more likely to form and/or persist, independent of local crater abundance. Such regions occur in both hemispheres but are more extensive and spatially coherent in the northern hemisphere, where occupation frequently exceeds 40–50%, particularly across Alba Patera, Nilosyrtris Mensae, and along the margins of Utopia Planitia. In contrast, high-occupation regions in the southern hemisphere are sparse and isolated, reflecting the substantially larger crater population and suggesting lower proportional retention of CCFs. This hemispheric contrast indicates that conditions favouring the formation or preservation of crater-

confined ice were more widespread or sustained in the north than in the south. High percentage occupation therefore isolates regions where CCF retention was unusually efficient relative to the available crater population, providing a complementary perspective to raw abundance and reinforcing the inference that CCF distribution reflects hemispheric differences in glacial accumulation and/or post-glacial modification rather than crater availability alone.

The spatial distribution of CCFs demonstrates that their occurrence is governed by host-crater geometry and by systematic latitude- and elevation-dependent constraints, rather than by crater availability alone. This chapter establishes that crater diameter exerts a first-order control on the likelihood of CCF preservation, while latitude defines a restricted mid-latitude window within which CCFs occur, and elevation modulates their presence toward the margins of that window. These geographic and morphometric dependencies identify systematic latitude- and elevation-related signals within the CCF population that may reflect underlying environmental gradients influencing CCF formation and/or preservation. However, the processes responsible for these signals—such as variations in insolation, surface temperature, or atmospheric pressure—are not tested here. The present analysis demonstrates that CCF distribution is neither random nor purely crater-controlled, but it cannot yet determine whether the preserved population reflects selective formation within favourable environments, selective survival through interglacial modification, or a combination of both. This uncertainty motivates the analyses in subsequent chapters, which examine spatial variability in CCF surface depth, volume, slope, and aspect—after accounting for host-crater controls—in order to evaluate whether these properties record systematic geographic signals consistent with climate-driven accumulation and/or ablation, or instead reflect crater-scale influences.

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Chapter 5: CCF Apparent Depth

5.1. Introduction

One consistent characteristic of CCFs is that their surfaces lie below the elevation of their host crater rims, though the degree to which they do so varies considerably. This variability in elevation between the CCF surface and the crater rim, reflecting differing degrees of crater infill or ablation, remains a notable but understudied feature. In this chapter, I will study this elevation difference, which I define as ‘apparent depth’ (d_a); see Figure 5.1. Figure 5.2 exemplifies this variation with two large CCF-hosting craters (27 km and 20 km in diameter) in eastern Utopia Planitia. Despite their similar size and proximity, their CCF surfaces lie at different elevations relative to their rims. MOLA topographic transects show $d_a \approx 320$ m for one crater, $d_a \approx 210$ m for the other. Their current depths differ by ≈ 110 m. As we shall see, across Mars' mid-latitudes, d_a varies significantly, with some CCFs nearly rim-full, while others are positioned well below the rim.

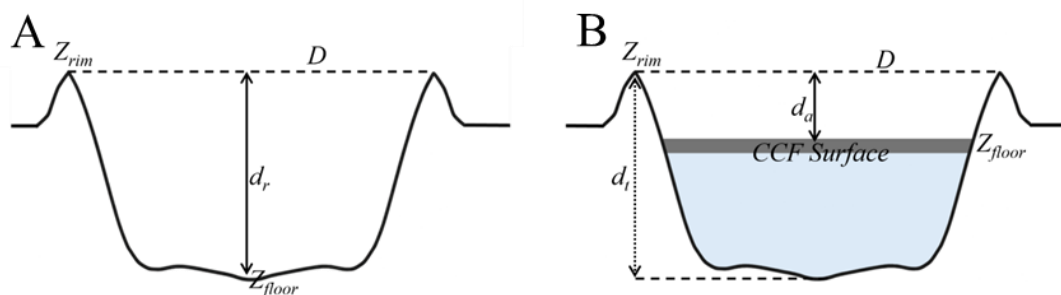


Figure 5.1. Cross section of (A) a pristine crater and (B) a CCF hosting crater. Symbols signify crater depth (d_r), rim elevation (Z_{rim}), crater floor elevation (Z_{floor}), CCF surface elevation (Z_{floor}), theoretical depth (d_t) and CCF apparent depth (d_a).

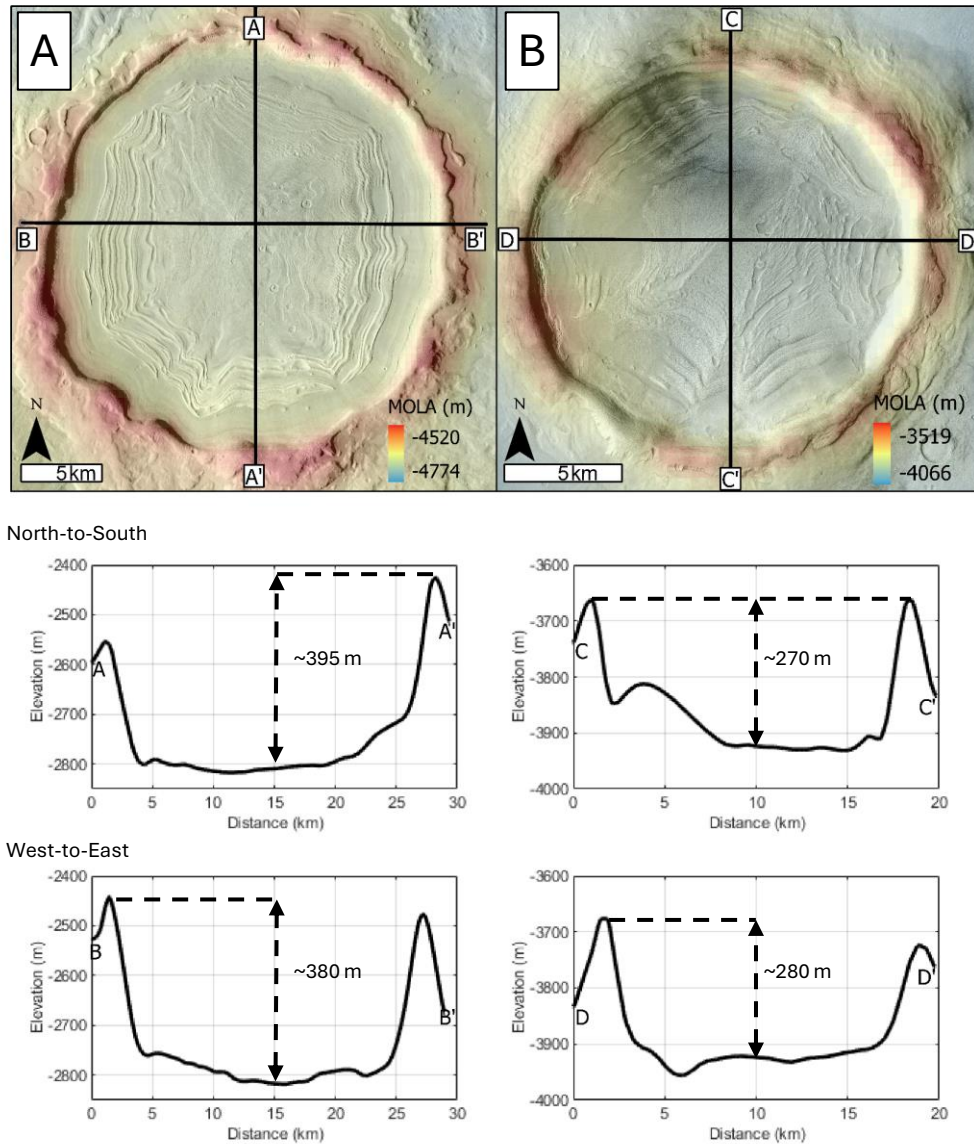


Figure 5.2 Topographic transects from gridded MOLA data across two CCFs found in Utopia Planitia. CTX images is overlaid on MOLA elevation data. (A) 37.89° N, 75.67° E; CTX images: P17_007714_2184 and B22_018144_2182. (B) 41.89°, 87.69°; CTX image: P22_009533_2219.

Since CCFs are interpreted as remnants of past mid-latitude glaciation (e.g. Forget et al., 2006; Fastook and Head, 2014; Levy et al., 2014), their present surface elevations record a history of ice accumulation and subsequent ablation. As discussed in Section 2.4.1, morphological evidence indicates that some CCFs formerly occupied substantially higher surface positions within their host craters. Upslope-directed ridges preserved on interior crater walls, at elevations above the present CCF surface and proximal to the crater rim, are interpreted as drop moraines and imply former ice surfaces several hundred metres higher than at present (Levy et al., 2010). In addition, lobate ridges preserved on exterior plains together with locally depressed crater rims indicate that, in some cases, palaeo-CCFs may have overtopped their host craters and extended onto adjacent terrain (Dickson et al., 2010). If CCF surfaces

once lay at or above rim level, why do all present-day CCFs now lie some distance below the crater rim?

Resolving this question requires determining whether this surface lowering is random, or if it systematically tracks the same environmental and geomorphological controls shown in Chapter 4 to govern CCF occurrence. Chapter 4 demonstrated that CCF distribution reflects the combined influence of host-crater diameter (D), latitude (δ), and elevation, operating within a constrained climatic window. If these variables exert a first-order control on CCF mass balance, they should also be expressed in the present-day elevation of CCF surfaces.

Systematic covariance in d_a —whether positive or negative—with environmental gradients (e.g., longitude, δ , or elevation), would therefore indicate that the processes governing CCF evolution and preservation are spatially structured rather than purely local. δ , for example, determines the insolation regime and thus surface ice stability (Ward, 1974; Madeleine et al., 2009). Under present obliquity conditions, higher- δ CCFs are expected to experience reduced insolation and lower ablation rates—favouring long-term preservation and smaller d_a —whereas lower- δ CCFs receive greater insolation, promoting enhanced surface lowering and larger d_a . Host-crater geometry may further modulate these effects, as larger craters provide greater relief and more extensive wall shading, while smaller craters offer limited protection. Elevation and regional setting may additionally influence d_a by modifying the local thermal regime.

Relevant questions are whether d_a is governed primarily by climate, crater properties (e.g., diameter and depth), or an interplay of both, and how these controls vary across spatial scales. This chapter addresses these questions by analysing the values of CCF d_a alongside their environmental and geomorphological parameters, keeping in mind that we don't know which of the two aforementioned hypotheses more accurately describes their current state.

Variations in d_a values between CCFs have been pointed out by previous studies but not fully explored. Pedersen and Head (2010), studying 212 degraded CCFs within Utopia Planitia, documented significant variability in surface elevation below the crater rim, finding cases ranging from deeply inset deposits to nearly rim-filled craters. Even among neighbouring craters of similar size, d_a differs by up to 250 m. Pedersen and Head (2010) attributed the cause to differences in crater age (i.e., different timespans of CCF accumulation), rather than local climate variations. Levy et al. (2014) similarly identified variations in CCF surface elevations from their original inventory (Chapter 3), but they did not investigate the spatial variations in d_a .

The aim of this chapter is to determine whether trends in d_a can be linked to these host-crater properties or instead record signals of past climate-driven processes. By analysing the global distribution of d_a alongside environmental parameters, I seek to identify systematic patterns that refine our understanding

of how and why CCF surfaces occupy their current elevations, and what this implies for their formation, modification, and long-term preservation.

5.1.1. Background on crater morphology and depth–diameter ratios

As introduced above, D may influence d_a through geometric effects such as wall shading. More fundamentally, crater size determines the original depth of the host crater prior to palaeo-CCF emplacement and therefore the maximum vertical space available for ice accumulation. The true rim-to-floor crater depth (d_r) imposes an upper bound on d_a , as CCF surfaces cannot lie below the original crater floor ($d_a \leq d_r$). Discriminating between a CCF surface that lies low because of climatic ablation and one that lies low simply because it formed within a naturally shallow crater therefore requires an independent estimate of d_r . Establishing this baseline—and estimating host crater d_r —is essential for separating climate-driven controls on d_a from constraints imposed by host-crater geometry.

A fundamental descriptor of impact crater morphology is the relationship between rim-to-floor depth (d_r) and D . This depth-to-diameter ($d_r:D$) ratio has been widely examined to establish baseline geometries for pristine craters—those retaining sharp rims, undegraded floors, and minimal infill—using a range of datasets, methodologies, and spatial scales (e.g., Garvin et al., 2003; Robbins and Hynek, 2012a; Robbins and Hynek, 2012b; Tornabene et al., 2018). Table 5.1 summarises a selection of these studies and their latitudinal coverage, with their associated scaling $d_r:D$ ratios illustrated in Figure 5.3. Among these, Garvin et al. (2003) established the first global $d_r:D$ ratios based on 2263 simple ($D < 7$ km) and 469 complex ($D > 7$ km) craters.

Subsequent work has demonstrated that $d_r:D$ ratios are not globally uniform. Even well-preserved craters exhibit systematic departures from global averages, reflecting regional variations in target properties and local geology (e.g., Garvin et al., 2003; Robbins and Hynek, 2012b). These departures are most pronounced at higher δ ($> \pm 40^\circ$), where fresh craters are consistently shallower than equatorial equivalents (e.g., Jankowski and Squyres, 1992; Stewart and Valiant, 2006; Robbins and Hynek, 2012b). This high- δ shallowing is widely attributed to viscous relaxation of ice-rich target crust or enhanced infilling rates in periglacial environments, both of which reduce the available accommodation space for subsequent CCF accumulation.

Table 5.1. Published depth-diameter ratios of pristine^a craters and their sources

Author	Types(s)/latitudinal range of craters studies	Equation	Crater Sample Size
Garvin et al., 2003	simple; global	$d_r = 0.21D^{0.81}$	2263
Garvin et al., 2003	complex; global	$d_r = 0.36D^{0.49}$	469
Robbins and Hynek 2012b	simple; global	$d_r = 0.179D^{1.012}$	37,091
Robbins and Hynek 2012b	complex; global	$d_r = 0.286D^{0.582}$	32,021
Robbins and Hynek 2012b	simple; $\leq -40^\circ, \geq +40^\circ$	$d_r = 0.177D^{0.742}$	12,210
Robbins and Hynek 2012b	complex; $\leq -40^\circ, \geq +40^\circ$	$d_r = 0.244D^{0.579}$	9742
Robbins and Hynek 2012b	simple; -40° to $+40^\circ$	$d_r = 0.175D^{1.022}$	24875
Robbins and Hynek 2012b	complex; -40° to $+40^\circ$	$d_r = 0.28D^{0.570}$	22290
Tornabene et al., 2018	global	$d_r = 0.347 \pm 0.021 D^{0.537 \pm 0.017}$	51

^aPristine crater interiors within the mid latitudes were identified by Robbins and Hynek 2012a (and updated 2014, 2021) crater dataset, in which their definition allows pristine craters to potentially have small craters superimposed or a small bit of modification on their interior.

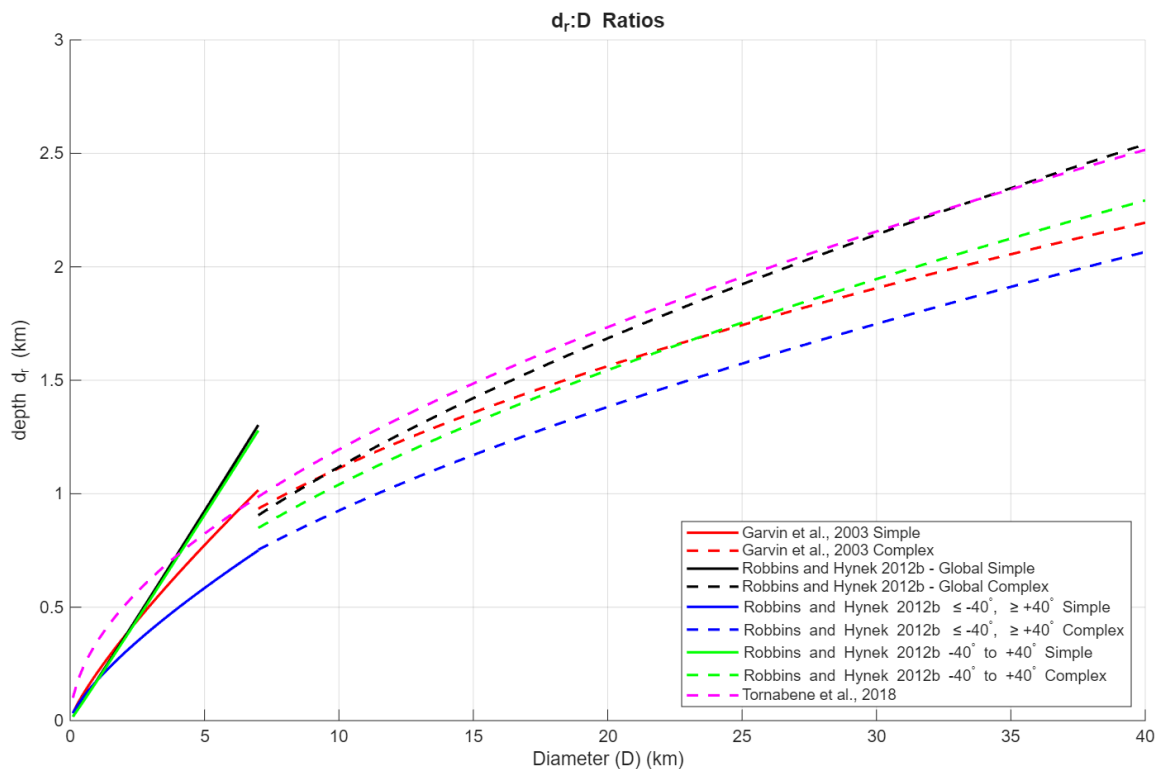


Figure 5.3. Pristine crater depth scaling on Mars. Global and regional crater depth: diameter ratios from Garvin et al. (2003), Robbins and Hynek (2012b), and Tornabene et al. (2018) for Martian craters with D between 0 and 40 km. The transition from simple to complex crater morphology is marked at $D = 7$ km.

To visualise these latitudinal trends in crater geometry, I utilise the global crater database of Robbins and Hynek (2012)—discussed later in Section 5.2—which provides morphometric parameters for Martian craters larger than 1 km in D . Figure 5.4 plots the $d_r:D$ ratios for both pristine (depicted in red)

and non-pristine craters with a certain degree of infill (depicted in blue) craters from this population against δ . The distribution reveals a clear spatial dependency affecting both populations: while equatorial craters reach high depth ratios, both pristine and non-pristine craters exhibit a marked decrease in d_r values poleward of $\pm 40^\circ$. Furthermore, the density of pristine craters notably diminishes at these higher latitudes, reflecting the enhanced modification rates described above. The latitudinal distribution of CCFs (grey bands) coincides precisely with this transition zone. Thus, applying a single global $d_r:D$ ratio would inaccurately estimate the initial depths of mid-latitude host craters as they sit within this transition. Consequently, accurate calculation of CCF d_a requires deriving a custom $d_r:D$ ratio specific to the mid-latitude bands to calculate CCF d_r .

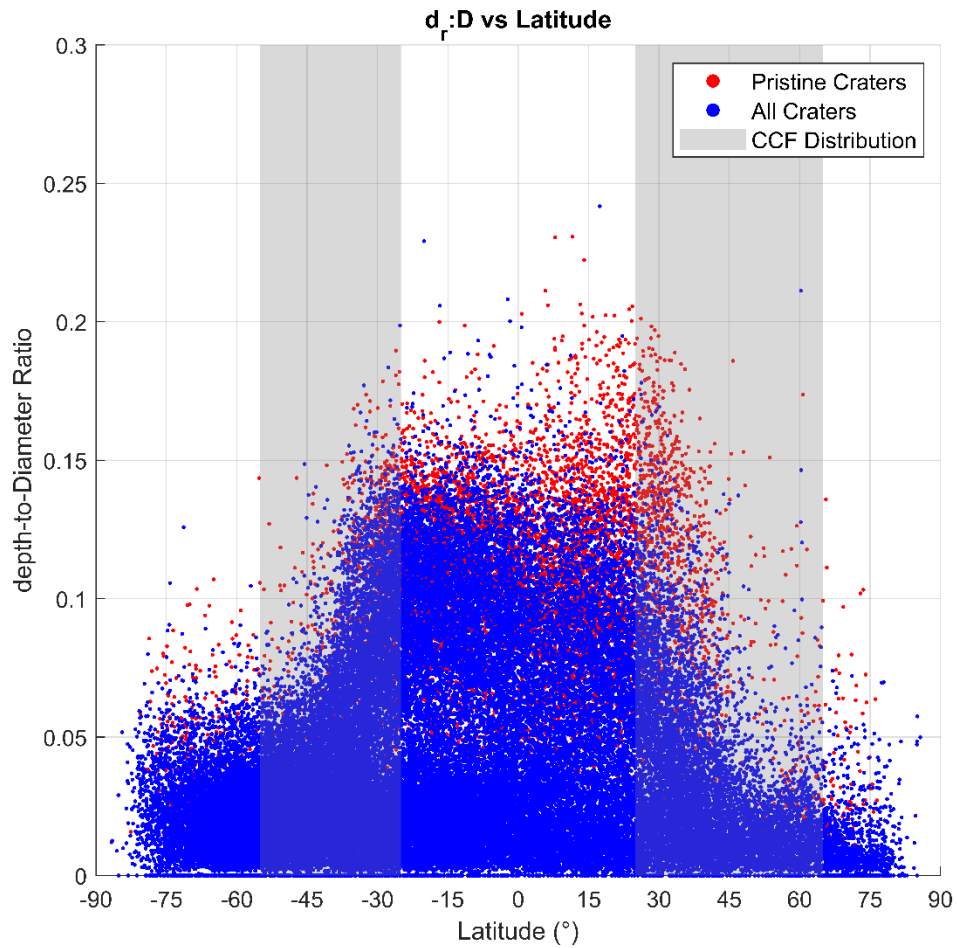


Figure 5.4. Latitude-dependent variations in depth: diameter ratios of pristine and infilled craters.

Pristine (red) and infilled craters (blue) as classified in the Robbins and Hynek (2012) global crater database exhibit a systematic shallowing trend beyond 40° latitude. The mid-latitude bands in which CCF occur (grey) align lie at the transition zones from deeper craters at the equator to shallower ones at higher latitudes. Data obtained from the Robbins and Hynek (2012) global crater dataset.

Even for the craters classified as pristine in the Robbins and Hynek (2012b) dataset, significant variability exists in the d_r/D ratios in the mid- to high latitudes (Figure 5.4). This suggests that those craters having fresh appearance in THEMIS Daytime IR images (thus classified as “pristine”), may still

contain substantial infill or mantling, resulting in lower d_r/D ratios. Tornabene et al. (2018) caution that scaling equations derived from such populations can be skewed, noting that visual ‘freshness’ in THEMIS images does not guarantee a lack of fill; their inspection of the 50 deepest craters revealed that some lacked diagnostic floor features (e.g., pitted materials), indicating that even the ‘fresh’ population is contaminated by partially infilled examples.

Given these uncertainties, relying on global d_r/D ratios is insufficient for the latitudinal extents of this chapter. To determine d_r , this chapter derives custom d_r/D ratios specific to the mid-latitudes ($\pm 30^\circ$ to $\pm 60^\circ$). The methodology (Section 5.2) follows the approach of Boyce and Garbeil (2007) and Robbins and Hynes (2012b), employing D size bins and a fitting power-law ratio using only the deepest craters within each bin (e.g., the single deepest or average of the two deepest). This ‘upper envelope’ approach should filter out the shallower, potentially modified craters as discussed by Tornabene et al. (2018). By establishing this robust local baseline, it becomes possible to interpret CCF d_a not just as a depth from the rim, but as a position relative to the floor. This allows us to investigate whether CCFs maintain a consistent vertical offset above the theoretical crater bottom, potentially revealing a threshold of required fill for CCF formation or preservation.

5.2. Datasets and methods

This chapter draws on two complementary datasets introduced earlier in the Thesis: the high-confidence CCF sample established in Chapter 3 (subset of the Levy et al., 2014 global inventory) and the filtered mid-latitude crater population introduced in Chapter 4 (derived from the Robbins and Hynes (2012) global crater inventory). The high-confidence CCF dataset provides measurements of CCF-surface elevation within host craters, while the crater inventory supplies the geometric and topographic properties of the CCF-hosting craters and of non-CCF craters required for comparison. Using both datasets is necessary to place CCF apparent depths within the broader context of crater morphology and modification state across the mid-latitudes. The following sections detail how crater and CCF parameters are obtained (Section 5.2.1), assess the consistency of crater measurements used in this study (Section 5.2.2), derive theoretical host-crater depths for CCF analysis (Section 5.2.3), and quantify the associated uncertainties (Section 5.2.4).

5.2.1. Obtaining CCF and crater parameters

This chapter requires consistent measurement of crater geometry and surface elevations in order to quantify d_a and to compare CCF-hosting craters with the wider mid-latitude crater population (Figure 5.1). For each crater, I require D , mean rim elevation (Z_{rim}), and a representative interior surface elevation ($Z_{surface}$). For non-CCF craters, this interior elevation corresponds to the lowest measured crater floor elevation recorded in the Filtered Robbins and Hynes (2012) database. For CCF-hosting craters, the interior elevation is instead defined by the mean elevation of the CCF surface. d_a is then

calculated consistently as $d_a = Z_{\text{rim}} - Z_{\text{surface}}$ —where Z_{surface} refers either to the crater floor/deposit elevation (non-CCFs) or the CCF surface elevation.

For CCF-hosting craters, I obtained host crater parameters (D , Z_{rim}) directly from the Robbins and Hynek (2012) global crater inventory, ensuring consistency with the non-CCF crater population. I did not use the Robbins and Hynek interior elevation for these craters, as it does not resolve the geometry of internal ice fills. Instead, I redrew CCF surface outlines from Levy et al. (2014) to isolate only the exposed CCF surface, excluding superimposed deposits where present. I traced these outlines manually in ArcGIS Pro using CTX mosaic images (Malin et al., 2007) at 6 m pixel⁻¹ resolution and projected them into Lambert Conformal Conic (LCC) space with standard parallels at 35° and 55° N for the northern hemisphere and -35° and -55° S for the southern hemisphere. I then extracted mean CCF surface elevation from the MOLA DEM using these revised outlines.

To verify D and Z_{rim} for CCF-hosting craters, I independently sampled crater rim topography using a custom MATLAB workflow. Around each revised CCF outline, I applied a 5 km buffer, which was used to clip the MOLA DTM prior to export into MATLAB. From a manually defined crater centre, the tool sampled radial elevation profiles at 1° azimuthal spacing. Along each profile, the maximum elevation was identified as the rim point, yielding a continuous rim-elevation trace (Figure 5.5). Mean rim elevation and crater diameter were then calculated geometrically from this trace. This procedure produces an independent set of D and Z_{rim} estimates for CCF-hosting craters, which are evaluated against the corresponding values reported in the Robbins and Hynek (2012) inventory in Section 5.2.2 to assess consistency and measurement uncertainty.

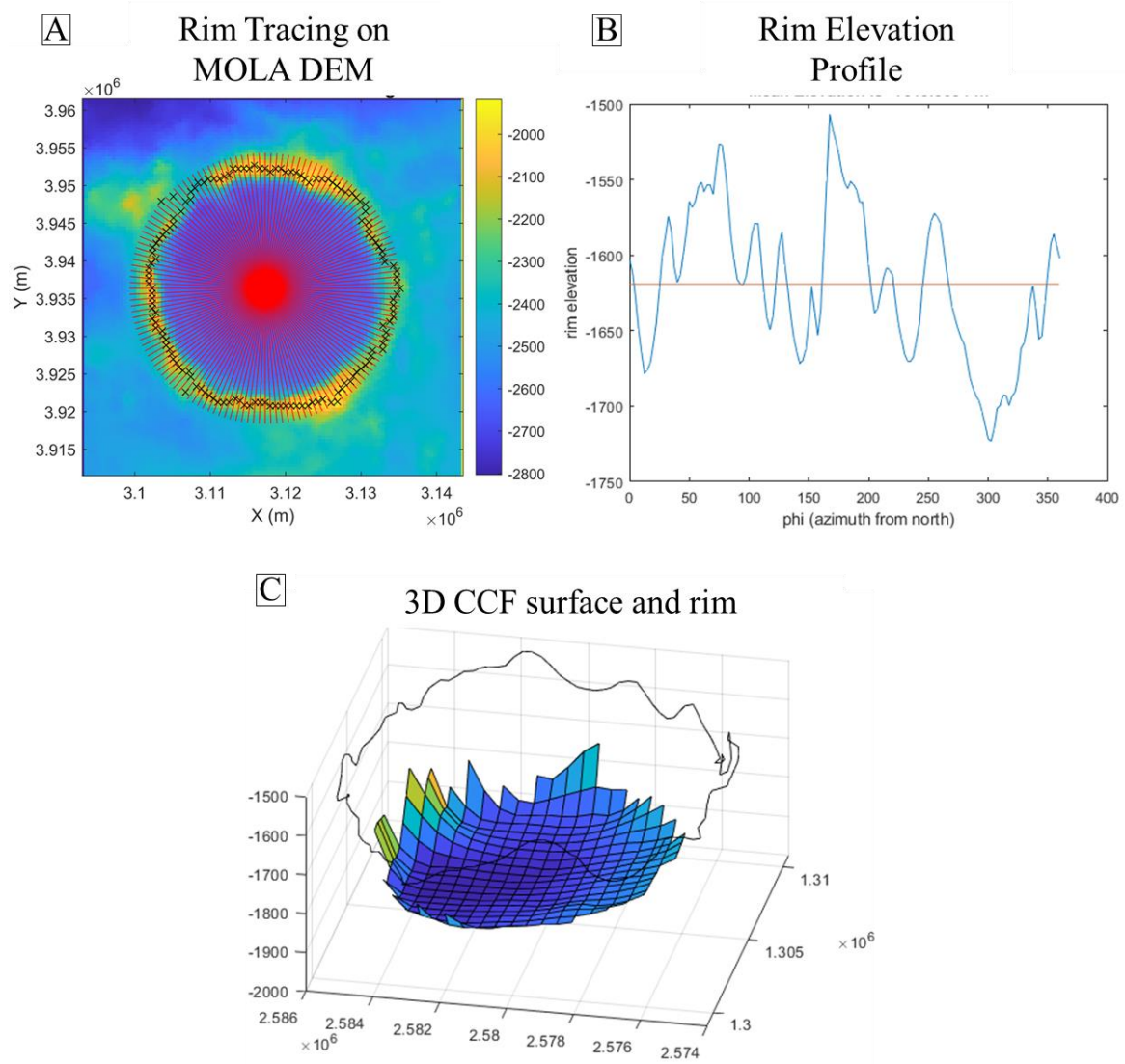


Figure 5.5. Crater rim elevation extraction methodology. Schematic illustration of the procedure used to determine mean crater rim elevation from MOLA topography. (A) MOLA DEM of a CCF-hosting crater, showing the radial sampling scheme in which lines are cast from the crater centre at 1° azimuthal intervals; the maximum elevation along each radial is taken to represent the local rim elevation. (B) Resulting crater rim elevation profile, with elevation (m) plotted against azimuth measured clockwise from north at 1° intervals; the mean rim elevation is shown in red. (C) Three-dimensional surface model of the crater derived from the gridded MOLA DEM, showing the extracted rim trace (black outline) superimposed on the crater topography.

The non-CCF-hosting crater population are utilised in two ways within this chapter. First, to calculate apparent depths for the non-CCF crater population as a comparative backdrop for CCF d_a vs D ; and second, to derive $dr:D$ ratios from pristine craters, which are required to estimate d_i for CCF-hosting craters (Section 5.2.3). All required parameters for these non-CCF hosting craters were obtained directly from the Robbins and Hynke (2012) global crater inventory. In this dataset, crater diameter and rim elevation are derived from manually digitised rims mapped primarily using THEMIS Daytime Infrared images, while interior elevation represents the lowest point within the crater cavity. Importantly, for

non-CCF craters, the interior elevation does not imply an unmodified crater floor; it simply records the lowest preserved surface within the cavity.

By combining Robbins and Hynek host-crater geometry with MOLA-derived CCF surface elevations, I ensure that apparent depth is defined consistently across the dataset while allowing CCF surfaces to be analysed relative to both their host-crater geometry and the broader population of mid-latitude craters.

5.2.2. Consistency of host-crater parameters between datasets

The analyses in this chapter depend directly on the accuracy of D and Z_{rim} because these parameters underpin all subsequent calculations of apparent depth (d_a) in this chapter and in later chapters. It is therefore essential to establish whether the values reported in the Robbins and Hynek (2012) crater inventory are consistent with those obtained using an independent measurement approach. Accordingly, in this section I compare D , Z_{rim} , $Z_{surface}$, and derived d_a values from the filtered Robbins and Hynek (2012) dataset with values measured directly from MOLA topography using the custom rim-tracing workflow developed in this chapter. The purpose is not to replace the Robbins and Hynek dataset, but to evaluate its suitability for CCF-focused analyses and to quantify the uncertainty introduced by different measurement approaches

Independent rim and diameter measurements

For all CCF-hosting craters, I independently traced crater rims using the MOLA-based workflow described in Section 5.2.1. This procedure successfully reconstructed rim elevation profiles for many isolated and well-preserved craters but proved unreliable in regions of complex topography, including overlapping craters, dissected rims, and steep background relief. In such cases, the assumption that the crater rim corresponds to the local elevation maximum breaks down, leading to erroneous rim identification.

To identify and exclude these cases, I assessed rim-trace consistency by measuring the radial distance from the crater centre to the traced rim at 1° azimuthal spacing. For each crater, I calculated the mean rim radius and its standard deviation (σ). Craters with a standard deviation exceeding 0.1 of the mean radius—indicating irregular or poorly resolved rims—were excluded from further analysis. After this filtering, 1386 CCF-hosting craters remained (1058 in the northern hemisphere and 328 in the southern hemisphere), spanning diameters of approximately 3–27 km. This subset represents cases where the elevation-based rim-tracing method performs reliably and is used solely for validation purposes in this section.

Rim elevation consistency

Both the Robbins and Hynek (2012) dataset and the MOLA-based measurements derived here provide estimates of crater rim elevation: Robbins and Hynek (2012) derive rim elevations from THEMIS

Daytime IR images combined with MOLA topography, whereas the present study samples rim elevations directly from MOLA along the full rim circumference. Residuals were calculated as the difference between rim elevations measured using the MOLA-based rim-tracing workflow—here after *MOLA-Traced*—and those reported by Robbins and Hynek (2012) ($\Delta Z_{rim} = Z_{rim, \text{MOLA-Traced}} - Z_{rim, \text{RH}}$). Across the refined subset, rim elevations show strong agreement (Figure 5.6A). For craters with $D < 7$ km, the mean residual is -0.015 km, indicating a very small systematic offset, whereas for craters with $D > 7$ km the mean residual is effectively zero. In total, 87.6% of craters have rim-elevation residuals within ± 0.05 km between the two methods. This close residual correspondence indicates that, where the rim-tracing workflow succeeds, elevation variation around crater rims is small. It also demonstrates that the Robbins and Hynek (2012) rim elevations provide a reliable representation of host-crater rim height for CCF analyses.

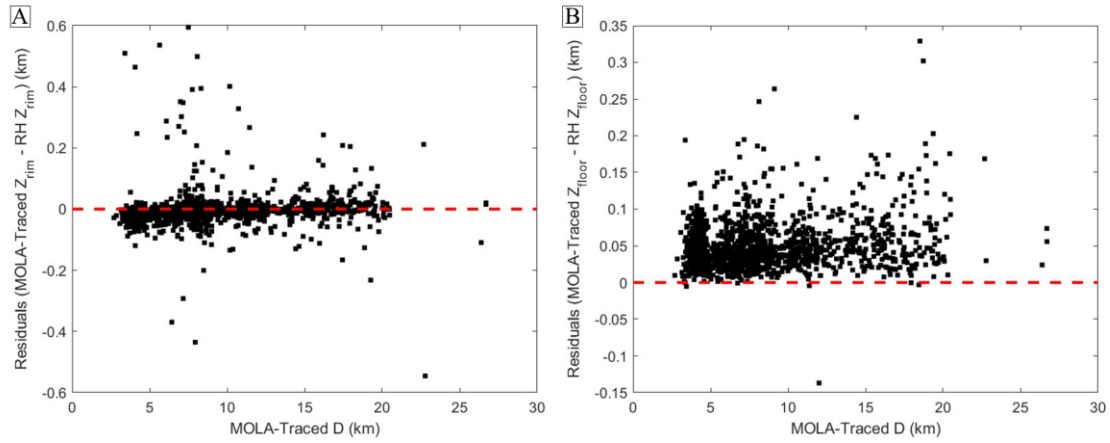


Figure 5.6. Residual comparisons of rim and floor elevations derived here and from Robbins and Hynek (2012). (A) Residual rim elevation ($\Delta Z_{rim} = Z_{rim, \text{MOLA-Traced}} - Z_{rim, \text{RH}}$) plotted against D . (B) Residual floor/surface elevation ($\Delta Z = Z_{floor, \text{MOLA-Traced}} - Z_{floor, \text{RH}}$) plotted against D . Both plots show strong overall agreement and indicate that residuals exhibit no systematic dependence on crater diameter.

Floor elevation and apparent depth consistency

For CCF-hosting craters, floor elevation differs fundamentally between the two approaches. In the Robbins and Hynek (2012) dataset, crater floor elevation represents the lowest measured point within the crater. In contrast, for CCFs I define floor elevation as the mean elevation of the CCF surface, measured from MOLA using the revised CCF outlines.

Residuals between CCF surface elevations derived here and the Robbins and Hynek (2012) floor elevations are predominantly positive (Figure 5.6B), as expected, because the Robbins and Hynek floor elevation corresponds to the single lowest point within the crater, whereas the present study defines floor elevation as the mean elevation of the CCF surface. Mean residuals are 0.044 km for craters smaller than 7 km and 0.055 km for larger craters. A small number of negative residuals arise—likely where the lowest point of the crater floor lies outside the mapped CCF outline—highlighting the

sensitivity of mean CCF elevation to outline definition rather than a systematic discrepancy between datasets.

Diameter and apparent depth consistency

D also differs slightly between the two approaches (Figure 5.7A). D derived from the MOLA-Traced are systematically larger than those reported by Robbins and Hynek (2012), although the correlation remains strong ($R^2=0.89$). This offset reflects methodological differences: the MOLA-Traced measures distances between sampled rim high points, whereas Robbins and Hynek (2012) employ a non-linear least-squares circle fit to manually mapped rim vertices. The magnitude of this difference is small relative to the D bins used later in this chapter and thus does not affect the interpretation of diameter-dependent behaviour.

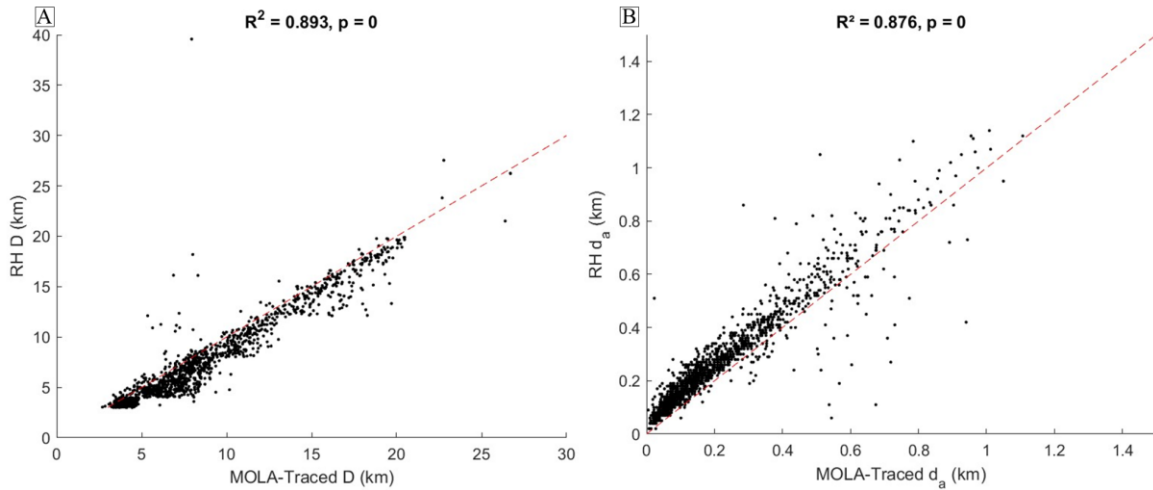


Figure 5.7. Comparison of crater diameter and apparent depth derived here and from Robbins and Hynek (2012). (A) D measured using the MOLA-Traced method versus D from Robbins and Hynek (2012). (B) d_a derived from MOLA-Traced versus d_a from Robbins and Hynek (2012). In both panels, the red dashed line indicates the 1:1 relationship. d_a in each case is calculated as $d_a = Z_{rim} - Z_{floor}$. Linear regression is used to quantify agreement between the two datasets.

For CCFs in craters with $D < 7$ km, the mean deviation between d_a derived here—using Z_{rim} from the MOLA-Traced workflow and Z_{floor} derived from mean CCF surface elevation (MOLA)—and d_a derived from the Robbins and Hynek (2012) Z_{rim} and Z_{floor} parameters is 19.1%, whereas craters with $D > 7$ km show a smaller mean deviation of 9.47% (Figure 5.7B). In both cases, apparent depth is defined as $d_a = Z_{rim} - Z_{floor}$. Overall, 58.8% of CCFs (815 craters in the refined subset) fall within $\pm 10\%$ of the 1:1 relationship, and 72.2% (1,001 craters) fall within $\pm 15\%$, indicating generally good agreement between the two estimates despite methodological differences. This agreement is further supported by a high coefficient of determination ($R^2 = 0.88$).

Consistency of d_a - D behaviour

Before analysing d_a vs D (Section 5.3.2), it is necessary to confirm that the d_a vs D behaviour is consistent between the two datasets. Figures 5.8A and 5.8B show that both approaches yield the same first-order behaviour: d_a increases with D , with increasing scatter and widening bounds at larger D . When plotted on log-log axes and binned by D (Figures 5.8C–D), both datasets define comparable upper and lower envelopes of d_a , confirming that the observed scaling is not an artefact of the measurement method.

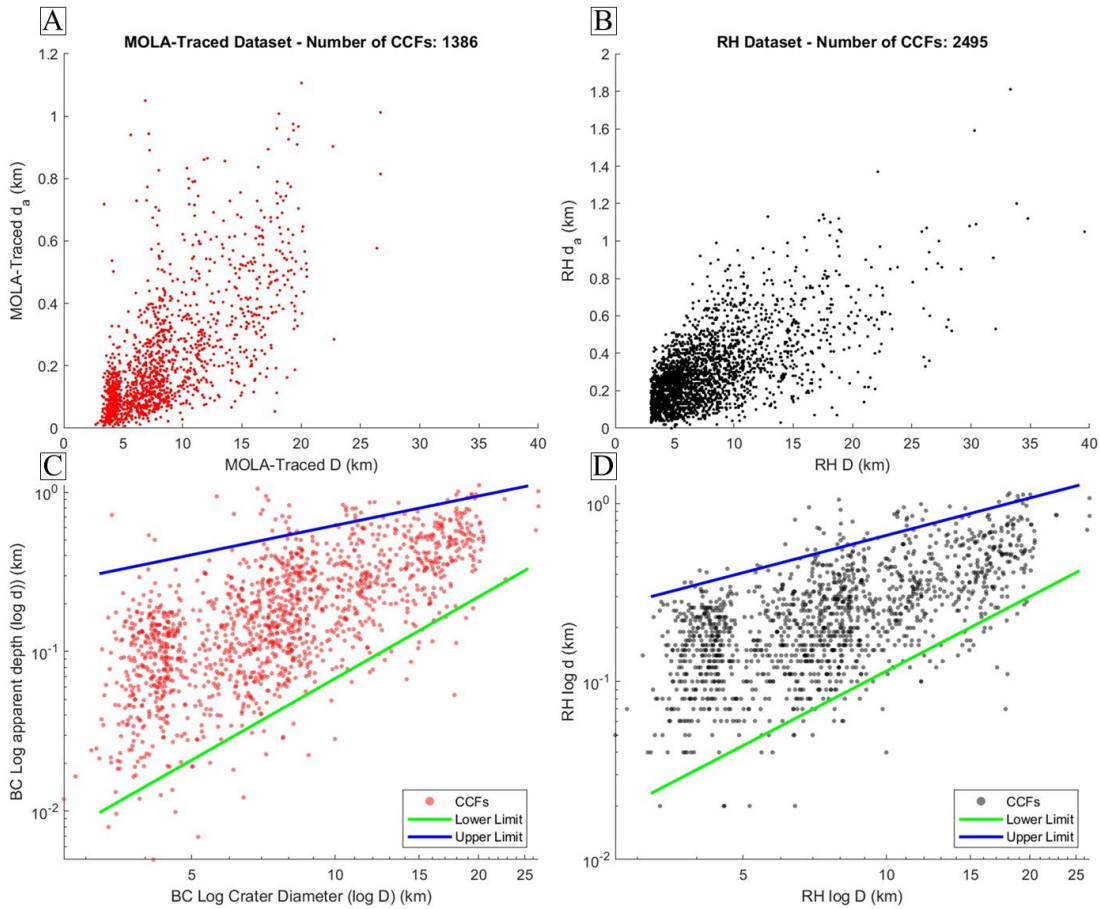


Figure 5.8. Comparison of apparent depth vs crater diameter derived here and from Robbins and Hynek (2012). (A) d_a vs D using values from Robbins and Hynek (2012). (B) d_a vs D using values derived from the MOLA-Traced measurements in this study. (C) Same as (A) plotted on log-log axes, with upper (blue) and lower (green) 95% envelope curves derived from 16 D bins of equal population size. (D) Same as (B) with corresponding envelope curves. Both datasets exhibit the same first-order behaviour, with the lower envelope of d_a increasing rapidly with D and the upper envelope rising more gradually.

Implications for subsequent analysis

This validation demonstrates that the Robbins and Hynek (2012) crater inventory provides internally consistent and reliable host-crater parameters for the purposes of this chapter. Consequently, for the

remainder of this chapter and the Thesis, I obtain D and Z_{rim} values from the Robbins and Hynek (2012) crater inventory to maximise CCF hosting crater sample size and ensure consistency with the non-CCF crater population. CCF surface elevations are taken from the MOLA-Traced measurements presented here. Together, this combination allows d_a to be evaluated robustly while retaining a coherent reference frame for comparison between CCF-hosting and non-CCF-hosting craters.

5.2.3. Calculating CCF host crater depth

Calculating d_i for each CCF-hosting crater allows d_a to be interpreted as a position within the crater cavity rather than simply as a depth below the rim. At any given D , d_i defines the maximum permissible d_a , such that $d_a \leq d_i$, and therefore provides the geometric reference needed to assess whether CCFs consistently maintain a finite vertical offset above the theoretical crater floor, or whether this offset varies systematically with D . To derive d_i , I follow the upper-envelope depth–diameter scaling approach introduced by Boyce and Garbeil (2007) and subsequently applied by Robbins and Hynek (2012b). The Robbins and Hynek (2012) crater database was first filtered to retain only craters classified as pristine according to their morphological classification scheme, excluding craters identified as having rim degradation, interior infill, or resurfacing. Pristine craters were then divided into simple (2–7 km) and complex (7–40 km) D ranges and then subdivided into 1 km D bins. Within each D bin, the mean depth of the deepest n craters ($n = 10$ –25, depending on data availability) was taken to represent the pre-modification crater depth for that bin. Power-law ratios were then fitted separately for simple and complex crater populations to these bin-wise mean depths, yielding distinct simple- and complex-crater $d_i:D$ scaling ratios. This procedure was applied both across the full mid-latitude CCF domain in each hemisphere and within 10° δ bins, enabling hemispheric estimates of d_i while resolving systematic latitudinal variation in crater depth.

5.2.4. Uncertainty

Uncertainty in D for both CCF-hosting and non-CCF craters is taken directly from the Robbins and Hynek (2012) database. In their inventory, D is derived by fitting a circle to manually digitised rim points, and the associated D uncertainty is reported as the formal standard deviation of the circle fit. These uncertainties scale weakly with crater size, amounting to ~ 0.07 km for craters of $D \approx 3$ km ($\approx 2.3\%$) and ~ 0.6 km for craters of $D \approx 40$ km ($\approx 0.4\%$), consistent with the few-percent deviations identified through independent checks using higher-resolution images (Robbins and Hynek, 2012a). Given their small fractional magnitude, D uncertainties are considered negligible relative to the 1 km binning intervals used in the subsequent analysis and consequently excluded.

A separate source of uncertainty arises from the use of gridded MOLA topography. The $1/128^\circ$ MOLA DEM is derived from irregularly spaced laser-altimetry points, and the number of DEM pixels within a crater does not necessarily correspond to the number of original MOLA shot points that constrain the crater topography. The pixel-based filtering applied to small craters therefore assesses whether a feature

is adequately represented at the raster resolution of the DEM, but it does not provide a direct measure of underlying MOLA point density. MOLA ground-track spacing is generally greater at lower latitudes and closer at higher latitudes, so interpolation effects may vary spatially. However, the present analysis is restricted to mid-latitude bands rather than an equator-to-pole transect, and the MOLA grid resolution is based on equatorial point spacing; consequently, the studied midlatitudes are generally better constrained than the equatorial regions for which the grid spacing was designed. Nevertheless, local interpolation errors may affect crater rim, crater floor, and CCF surface-elevation measurements, particularly for smaller craters.

Uncertainty in d_a is obtained by propagating errors from Z_{rim} and $Z_{Surface}$ measurements. For Z_{rim} , I use the standard deviation of rim elevation reported by Robbins and Hynek (2012), which reflects variability around the crater rim sampled from MOLA topography. An additional ± 50 m uncertainty is applied to account for combined uncertainty in CCF $Z_{Surface}$ and Z_{rim} derived from MOLA data. This value represents the maximum observed discrepancy between gridded and point-based MOLA elevations reported by Robbins and Hynek (2012a), corresponding to ± 25 m uncertainty in each elevation measurement. These components were combined to define the total uncertainty on d_a for CCF-hosting craters used throughout this chapter.

5.3. Results and Analysis

This section presents the results for d_a . The analysis begins by examining the influence of D on d_a (Section 5.3.2), recognising that crater geometry constrains the vertical position CCF surfaces can occupy. The results then evaluate how d_a varies with longitude (Section 5.3.3), δ (Section 5.3.4), and elevation (Section 5.3.5), in order to assess the role of geographic and environmental controls on CCF surface elevation. The final section integrates these variables using multilinear regression to quantify their relative importance (Section 5.3.6). Throughout, Northern and Southern hemisphere CCF populations are analysed separately to isolate potential hemispheric differences.

5.3.1. Mid-Latitude host crater theoretical depth

Whilst the influence of D on d_a is of leading interest for this chapter (Section 5.3.2), its interpretation is inherently constrained by host-crater depth, which itself increases with D (Section 5.1.2). Because CCF surfaces cannot lie below the true crater floor, d_a is bounded by the pre-CCF d_t , such that $d_a \leq d_t$. Establishing how d_t scales with D is therefore a necessary prerequisite for interpreting variations in d_a , as it defines the theoretical maximum depth available for CCF emplacement within a given host crater. Deriving an appropriate mid-latitude $d_r:D$ ratio allows d_a to be evaluated not simply as a depth below the rim ($d_a = 0$), but as a position within the crater cavity relative to its theoretical floor. Establishing an appropriate $d_r:D$ ratio for mid-latitude craters is thus necessary to investigate whether CCFs maintain a consistent vertical offset above the theoretical crater bottom, potentially revealing a required threshold of fill for CCF formation or preservation. This baseline defines the geometric upper bound on

permissible d_a values, against which observed CCF surface positions can be assessed in Sections 5.3.2–5.3.5, allowing genuine variability in CCF surface position to be distinguished from limits imposed solely by host-crater geometry.

As discussed in 5.1.1, global $d_r:D$ ratios for Martian craters are well established, but their direct application to CCF-hosting craters is inappropriate because d_t varies systematically with latitude and is found to vary in either hemisphere (Robbins and Hynek, 2012b). To capture this variability, I derive $d_r:D$ ratios directly from the Robbins and Hynek (2012) crater inventory using craters classified as pristine within the mid-latitude bands occupied by CCFs. Craters are binned into 10° δ bands and analysed separately for each hemisphere to quantify latitudinal changes in d_r and to establish representative mid-latitude baselines for CCF host craters. These fitted $d_r:D$ ratios are shown in Figure 5.9 and summarised in Table 5.2.

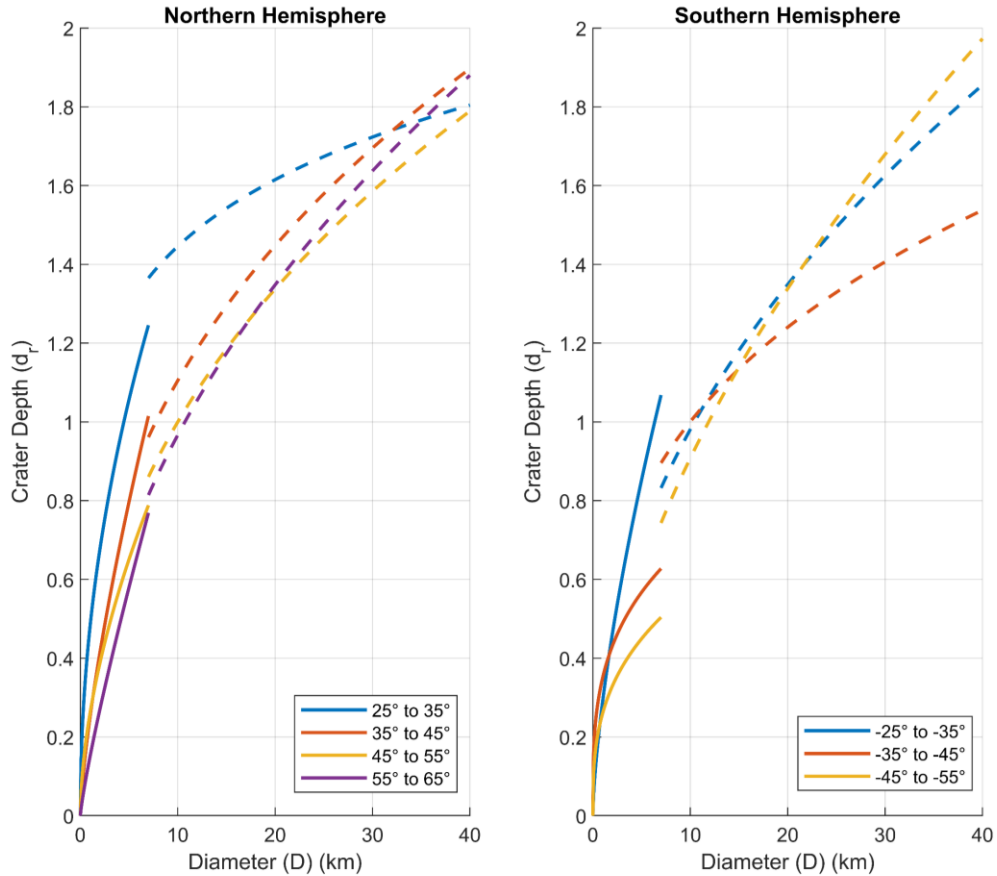


Figure 5.9. Mid latitude pristine crater $d_r:D$ ratios derived from fitted crater data of Robbins and Hynek (2012), separated in the (A) North hemisphere and (B) South hemisphere, in 10° latitudinal bins, & for simple (solid curve) and complex (dashed) craters.

Table 5.2. Crater $d_r:D$ Ratios across latitude bins

	Simple Crater Ratio	n	Complex Crater Ratio	n
Northern Hemisphere				
Mid Latitude Average	$d_r = 0.4D^{0.59}$	673	$d_r = 0.72D^{0.26}$	238 ^a
25° to 35°	$d_r = 0.48D^{0.49}$	462	$d_r = 1D^{0.16}$	136 ^a
35° to 45°	$d_r = 0.25D^{0.72}$	161	$d_r = 0.45D^{0.39}$	74
45° to 55°	$d_r = 0.26D^{0.57}$	24	$d_r = 0.38D^{0.42}$	16
55° to 65°	$d_r = 0.15D^{0.84}$	26	$d_r = 0.32D^{0.48}$	12
Southern Hemisphere				
Mid Latitude Average	$d_r = 0.48D^{0.29}$	431	$d_r = 0.53D^{0.31}$	88
-25° to -35°	$d_r = 0.29D^{0.67}$	321	$d_r = 0.34D^{0.46}$	49
-35° to -45°	$d_r = 0.35D^{0.30}$	84	$d_r = 0.49D^{0.31}$	30
-45° to -55°	$d_r = 0.26D^{0.34}$	26	$d_r = 0.25D^{0.56}$	9
^a 9 deepest craters removed				

The fitted $d_r:D$ curves (Figure 5.9) demonstrate that crater depth varies substantially across the mid-latitudes, even within the population classified as pristine. Differences between δ bands are pronounced, indicating that crater cavities of comparable D retain significantly different preserved d_r depending on δ . This variability stems not only from intrinsic $d_r:D$ differences but also from progressive latitudinal modification that persists despite a 'pristine' designation. Inspection of the deepest pristine-classified craters shows that even these examples commonly exhibit interior modification at mid- to high latitudes. CTX images reveals mantling and partial infill affecting both simple and complex craters poleward of $\sim 35^\circ$, with particularly clear examples in the 55° – 65° latitude bands (Figure 5.10). These observations confirm that craters classified as pristine at higher δ do not necessarily preserve unmodified excavation cavities, consistent with the findings of widespread mid-latitude resurfacing (Tornabene et al., 2018). As a consequence, high latitude $d_r:D$ ratios derived from pristine-classified craters are expected to underestimate true excavation depths, with the bias increasing toward higher latitudes.

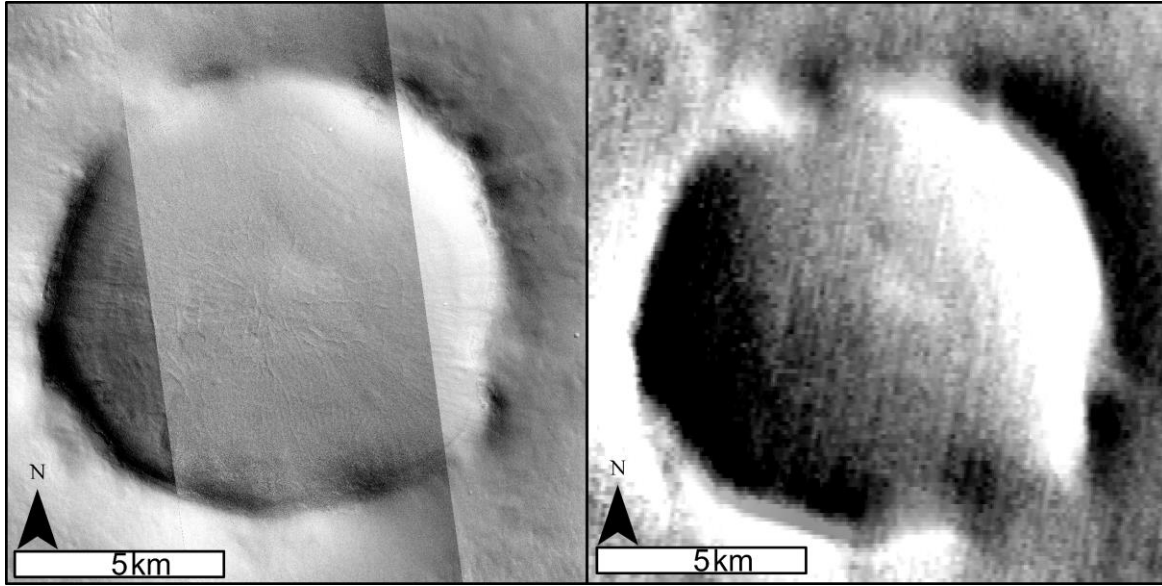


Figure 5.10. Infilled crater at high northern latitudes classified as pristine in the Robbins and Hynek (2012) global crater inventory. (A) HiRISE image (ESP_016925_2435) and CTX image (B19_017136_2433) showing extensive infilling of the crater interior. (B) The same crater in THEMIS images, where the interior appears morphologically simple and lacks clear expression of infill at this resolution. Although classified as pristine with a pristine interior in the Robbins and Hynek dataset, higher-resolution CTX and HiRISE images reveal substantial interior fill and the absence of an exposed, bedrock crater floor. Crater located at 63.32°N, 127.20W°.

Given these limitations, d_t for CCF-hosting craters is estimated using hemispheric mid-latitude $d_r:D$ ratios (bold in Table 5.2; Figure 5.11) rather than latitude-segmented fits. These are derived from the deepest craters within D bins across the full mid-latitude range in each hemisphere, following established upper-envelope approaches for approximating minimally modified crater geometries. This strategy reduces sensitivity to latitudinally variable infill while retaining hemispheric differences in crater morphology, providing a consistent geometric reference against which CCF apparent depths can be evaluated. In subsequent analyses, d_t is shown where appropriate as a reference level to illustrate the vertical offset of CCF surfaces from the inferred crater floor and is later used in Chapter 6 to quantify CCF volumes.

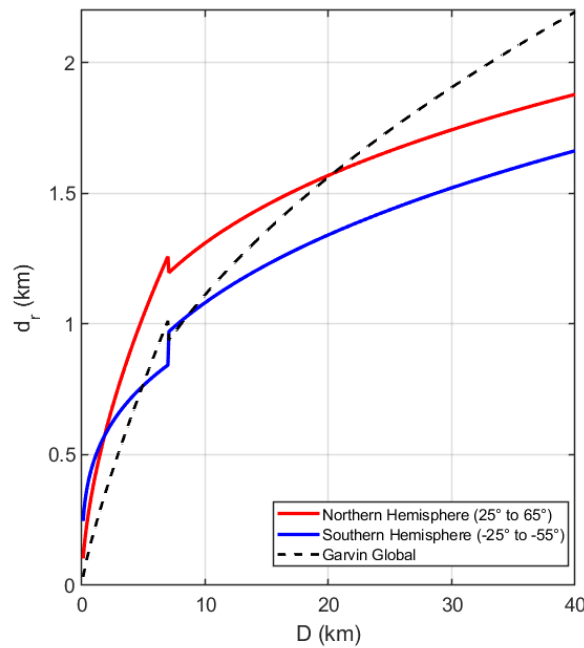


Figure 5.11 Global and mid-latitude pristine crater depth: diameter ratios. The global crater relationship from Garvin et al. (2003) is shown in black, with the corresponding equation listed in Table 5.1. The mid latitude $d_r:D$ ratios are presented in Table 5.2, with mid-latitude averages for the northern hemisphere in red and the southern hemisphere in blue. For all equations, the simple to complex crater morphology transition occurs at a diameter of 7 km.

5.3.2. Apparent depth vs host crater diameter

The distribution of CCF d_a vs D is shown in Figure 5.12A. While d_a generally increases with D , this increase is accompanied by substantial scatter, demonstrating that D alone does not uniquely determine a CCFs d_a . At a given D , CCFs span a wide vertical range, from near-rim surfaces ($d_a \approx 0$) to surfaces positioned far below the rim—but possess some separation from d_i perhaps suggesting a minimum CCF thickness. The distribution also narrows progressively toward larger D , with CCFs with low d_a absent at high D , reflecting the progressively larger ice volumes required to achieve near-rim-full states in larger craters (CCF volumes are addressed in Chapter 6). As established in Section 5.1, D exerts a dual control on d_a : it imposes an explicit cut-off for CCF occurrence—such as the absence of CCFs in craters with D exceeding ~ 40 km—and through the increasing range of permissible d_a values as d_i increases with D .

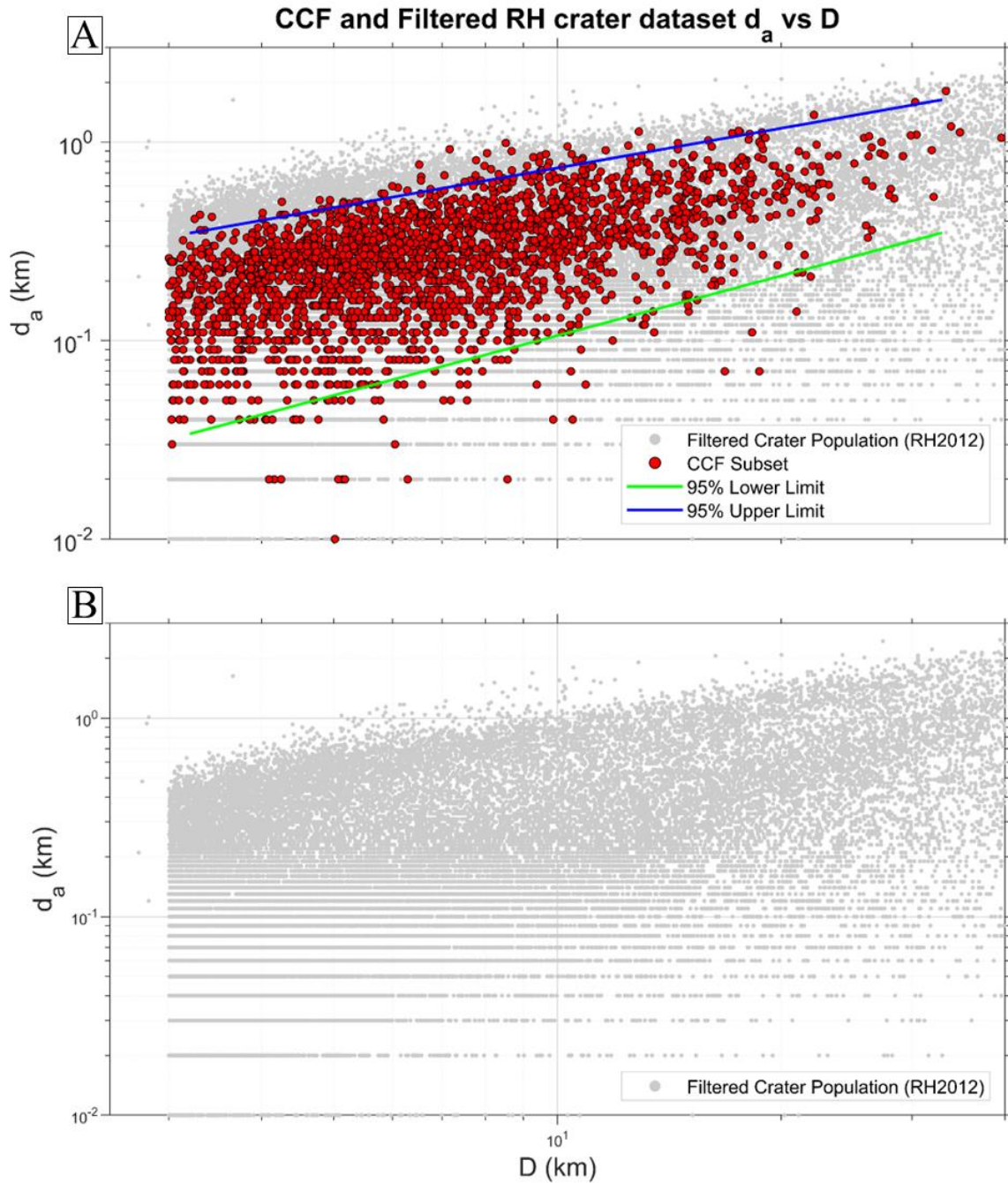


Figure 5.12. Scatter plots of d_a vs D in log scale for craters within the mid latitudes (filtered $\pm 30^\circ$ - 60° and D : 2.5–40 km) with 44484 non-CCF hosting craters from (Robbins and Hynek, 2012) (dark grey circles) and 2781 CCF hosting craters (with measurements obtained from this study, edited from Levy et al., 2014, and Robbins and Hynek, 2012) (red circles). (A) CCF-hosting craters d_a vs D overlaid on the filtered non-CCF hosting craters. Confidence intervals (green and blue lines) were derived by binning CCF apparent depths into 20 logarithmically spaced D bins. For each bin, the 2.5th and 97.5th percentiles were calculated, and a power-law regression was fitted to these values to define the 95% population boundaries. (B) 44484 non-CCF hosting craters d_a vs D .

To interpret these CCF distributions, it is necessary to place them in the context of the broader mid-latitude crater population. Figure 5.12A therefore overlays the CCF d_a vs D distribution onto the filtered population of non-CCF-hosting craters (shown in dark grey; Chapter 4). The CCF population occupies

only a restricted subset of this broader cloud, concentrated toward its central portion rather than spanning the full range of crater depths available at each D . Upper and lower bounds of this distribution were quantified using logarithmic diameter binning, extracting quantiles within each bin and fitting regression lines through bin midpoints. Figure 5.12B shows the d_a vs D distribution for the filtered mid-latitude non-CCF-hosting craters derived from the Robbins and Hynes (2012) inventory. This distribution reflects the full range of rim-to-floor depths produced by impact cratering and subsequent modification, expressed through depth-to-diameter scaling. When comparing the filtered crater population in Figure 5.12B and the CCF population in Figure 5.12A, it is clear that CCFs are confined to a restricted horizontal band within this broader envelope, remaining below d_t and often not completely rim full.

To examine this behaviour more systematically, CCFs were subdivided into sixteen D bins for the remainder of this section, normalising for the dominant influence of the host crater—represented by D . Within each D bin, d_a values are broadly distributed rather than concentrated around a single preferred depth, as confirmed by Kolmogorov–Smirnov tests (Table 5.3). For the combined dataset, the d_a distribution departs strongly from normality ($KS = 0.11$; $p = <0.001$), and this behaviour persists across all D bins. Each D class therefore permits a characteristic range of stable d_a values rather than a unique equilibrium depth.

Figure 5.13 presents box plots of d_a by D bin for the Northern and Southern hemispheres, with the theoretical crater depth (d_t)—calculated using the upper envelope of the mid-latitude depth–diameter scaling—shown for reference. In both hemispheres, median and maximum d_a increase with crater diameter, consistent with the expanding cavity available for ice emplacement. In the northern hemisphere, minimum d_a values remain close to zero for craters smaller than ~ 20 km, indicating that near-rim CCF surfaces can persist across a wide diameter range. In the southern hemisphere, minimum d_a values lie systematically closer to d_t , reflecting the shallower cavity depths of southern host craters and the reduced vertical separation available between CCF surfaces and crater floors. Across most bins, substantial separation remains between maximum observed d_a and d_t , indicating that CCFs do not typically occupy the full available cavity depth. Only in the largest D bin (30–40 km) does this separation narrow substantially, particularly in the south, where maximum d_a approaches d_t , representing an end-member rather than typical behaviour.

Table 5.3. Descriptive statistics and Kolmogorov–Smirnov tests of d_a (km) against D (km) for the high-confidence CCF population subdivided into sixteen D bins.

D range (km)	n	Minimum d_a (km)	Maximum d_a (km)	Median d_a (km)	Std Da (km)	Skewness	KS-test	KS-test P-value
3-4	373	0.0099	0.37	0.11	0.121	0.068	0.495	< 0.001
4-5	401	-0.012	0.43	0.15	0.162	0.089	0.36	< 0.001
5-6	354	0.0104	0.48	0.18	0.189	0.1	0.25	< 0.001
6-7	277	0.0098	0.6	0.21	0.219	0.11	0.17	< 0.001
7-8	216	0.031	0.83	0.2	0.25	0.17	0.11	< 0.001
8-9	182	0.0058	0.93	0.26	0.28	0.16	0.12	0.011
9-10	142	0.018	0.85	0.31	0.33	0.17	0.19	< 0.001
10-11	127	0.024	0.82	0.31	0.34	0.18	0.21	< 0.001
11-12	67	0.072	0.9	0.31	0.35	0.16	0.267	< 0.001
12-13	56	0.064	1.091	0.4	0.43	0.19	0.39	< 0.001
13-14	45	0.021	0.81	0.32	0.36	0.2	0.29	< 0.001
14-15	52	0.099	0.801	0.38	0.39	0.17	0.35	< 0.001
15-20	138	0.026	1.085	0.48	0.48	0.23	0.44	< 0.001
20-25	39	0.11	1.2	0.52	0.52	0.21	0.58	< 0.001
25-30	18	0.23	0.99	0.63	0.63	0.22	0.7	< 0.001
30-40	8	0.49	1.51	0.88	0.97	0.33	0.91	< 0.001

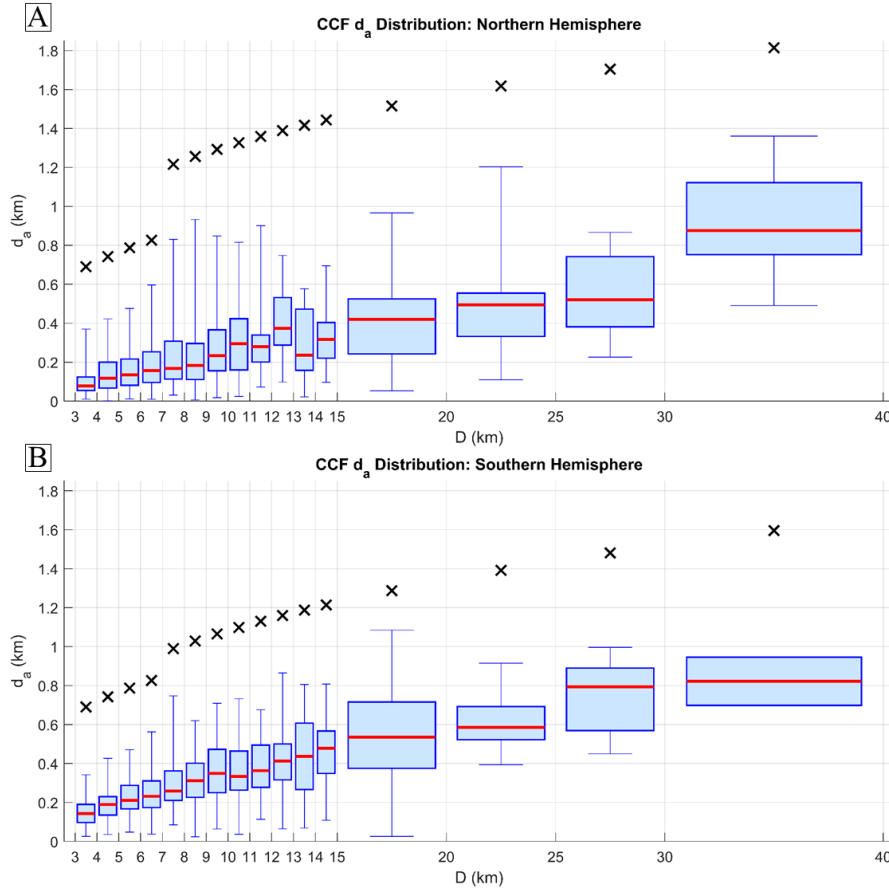


Figure 5.13. Hemispheric box plots of CCF d_a binned by D . (A) Northern Hemisphere CCF d_a by host crater D . (B) Southern Hemisphere equivalent. Boxes represent the interquartile range (IQR), where the red horizontal line indicates the median value for each bin. Whiskers extend to the minimum and maximum observed values within each diameter range. Bin widths are scaled to actual diameter ranges. Black crosses indicate d_r estimated from mi-latitude crater d_r/D ratios from Table 5.2.

These observations demonstrate that host-crater D defines plausible d_a range but does not solely resolve it. The consistent confinement of CCFs to a restricted portion of the available depth range suggests that additional spatial or environmental factors influence where CCF surfaces stabilise within their host craters. These factors—including longitude, δ , and elevation—are examined in the following sections using D binned analyses, ordinary linear regression (LR), and weighted linear regression (WLR). In the WLR, each observation is weighted by the inverse square of its estimated apparent-depth uncertainty, so observations with larger y-axis error bars exert less influence on the fitted regression

5.3.3. Variations of apparent depth with longitude

This section evaluates whether d_a varies with longitude and whether such variation, if present, offers insight into the mass-balance history of CCFs. Longitudinal effects are plausible given the presence of zonally structured processes in the Martian atmosphere. Longitudinal atmospheric tides modulate upper-atmospheric density by approximately 8–16% (Liu et al., 2017), reflecting variations in temperature, composition, and circulation that could influence spatial differences in ice accumulation

or sublimation. If such effects influence CCF mass balance over glacial timescales, CCFs located at similar longitudes might be expected to exhibit comparable d_a values. The analysis therefore proceeds in two steps. First, d_a is examined directly as a function of longitude for each hemisphere. Second, d_a is evaluated within host-crater diameter bins to determine whether any longitudinal dependence persists once the dominant influence of D is accounted for.

Figure 5.14 shows d_a plotted directly as a function of longitude for the Northern and Southern hemispheres. In both cases, the distributions are dominated by the strong dependence of d_a on D identified in Section 5.3.2, producing broad, vertically structured clouds of points with no visually coherent longitudinal organisation. At any given longitude, d_a spans nearly the full range permitted by crater geometry, from near-rim-full surfaces to deeply recessed CCFs. This confirms that any potential longitudinal signal, if present, is subordinate to the first-order control exerted by D and cannot be isolated in the unstratified data. Consequently, the analysis is repeated after separating the population into diameter bins to test whether a longitudinal dependence emerges once the geometric control imposed by crater size is accounted for.

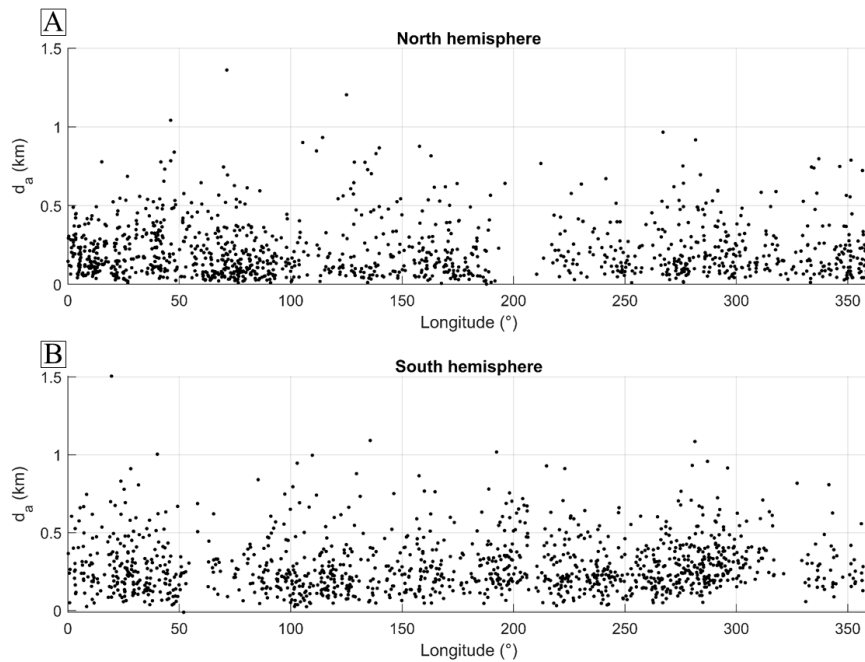


Figure 5.14. Scatter plot of d_a vs longitude. (A) Northern Hemisphere CCF d_a vs longitude. (B) Southern Hemisphere equivalent.

Figures 5.15 and 5.16 present stacked d_a -longitude plots for the Northern and southern hemispheres, separated by diameter bin. Across all bins in both hemispheres, no systematic longitudinal structure is observed. Values of d_a are distributed broadly and continuously with longitude, with no recurring maxima, minima, or preferred longitude bands that persist across multiple D ranges. This indicates that longitude does not impose a first-order control on CCF d_a , nor does it modulate the increase of d_a with D established in Section 5.3.2.



Figure 5.15. Stacked scatter plots of d_a vs. longitude, binned by D , for the Northern Hemisphere.

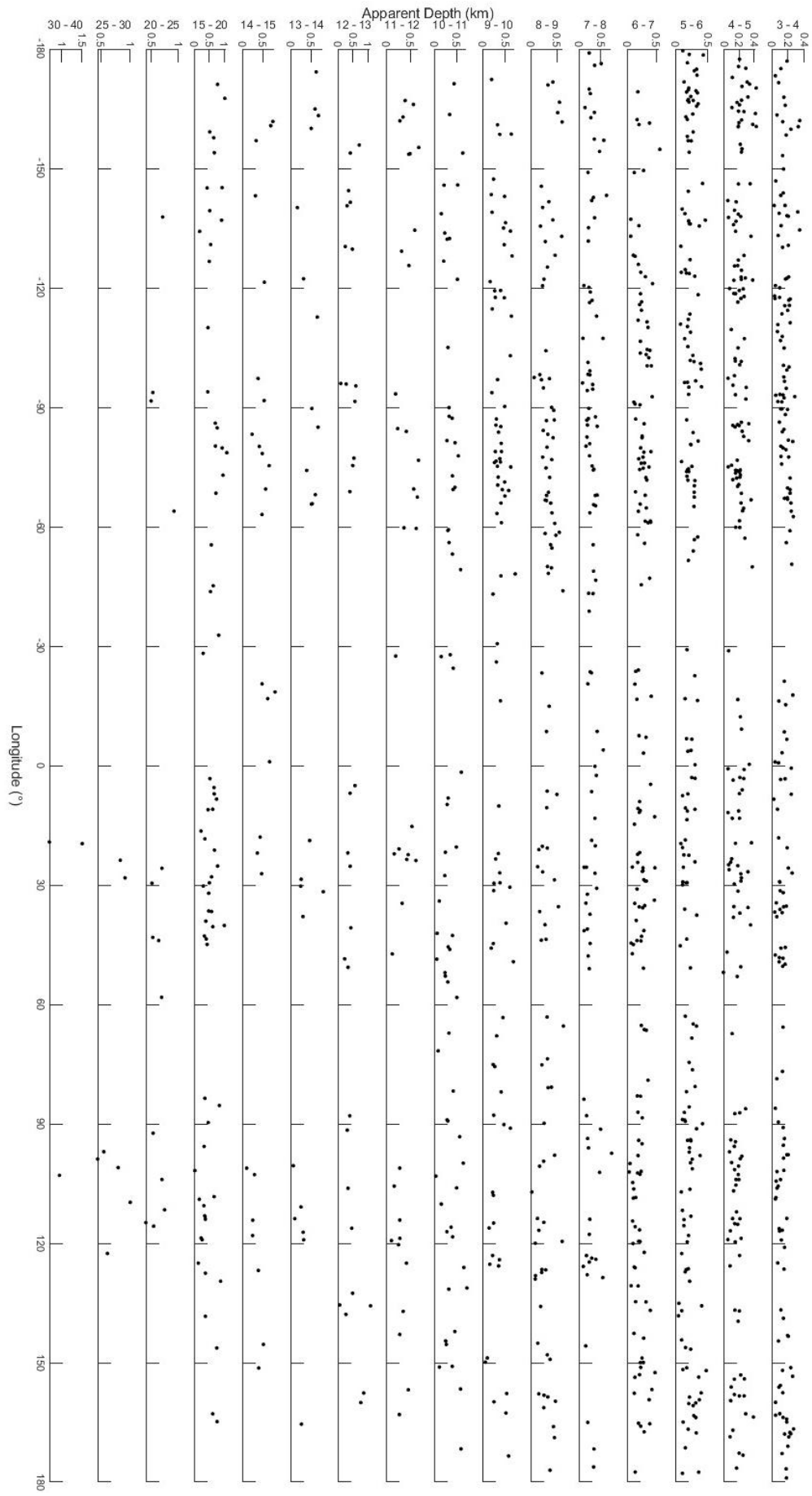


Figure 5.16. Stacked scatter plots of d_a vs. longitude, binned by D , for the Southern Hemisphere.

5.3.4. Variations of apparent depth with latitude

Figure 5.17 shows d_a plotted against absolute δ for CCF centres in the Northern (Figure 5.17A) and Southern (Figure 5.17B) hemispheres. In both hemispheres, the data form broad, vertically dispersed clouds, indicating substantial scatter in d_a at all latitudes. Despite this scatter, both populations show a weak but statistically significant decrease in d_a toward higher absolute latitudes (North: $R^2 = 0.013$, $p < 0.001$; South: $R^2 = 0.026$, $p < 0.001$), indicating that δ alone explains only a small fraction of the variance in d_a when the full population is considered—a result expected given the primary dependence of d_a on D identified previously. The presence of a statistically significant but weak latitudinal signal therefore motivates further analysis in which CCFs are binned by D , allowing the influence of latitude on d_a to be examined independently of the dominant host-crater control.

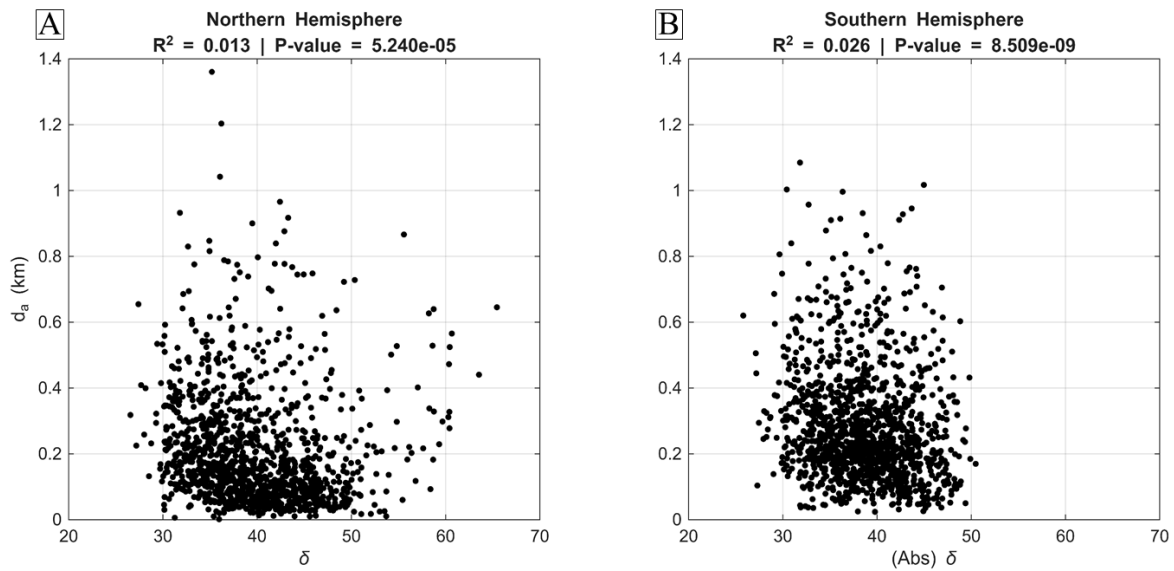


Figure 5.17. Scatterplots of d_a versus absolute latitude. (A) Northern hemisphere CCFs. (B) Southern hemisphere CCFs. In both hemispheres, d_a shows a weak but statistically significant decrease toward higher absolute latitudes.

When binned by D , the latitudinal signal in d_a becomes clearer as the dominant influence of D is partially constrained. Figures 5.18 and 5.19 show d_a plotted against latitude for the Northern and Southern hemispheres, respectively, with CCFs separated into discrete D bins. Within most D bins in either hemisphere, the data clouds show a consistent poleward gradient: d_a values decrease and the vertical spread narrows as δ increases.

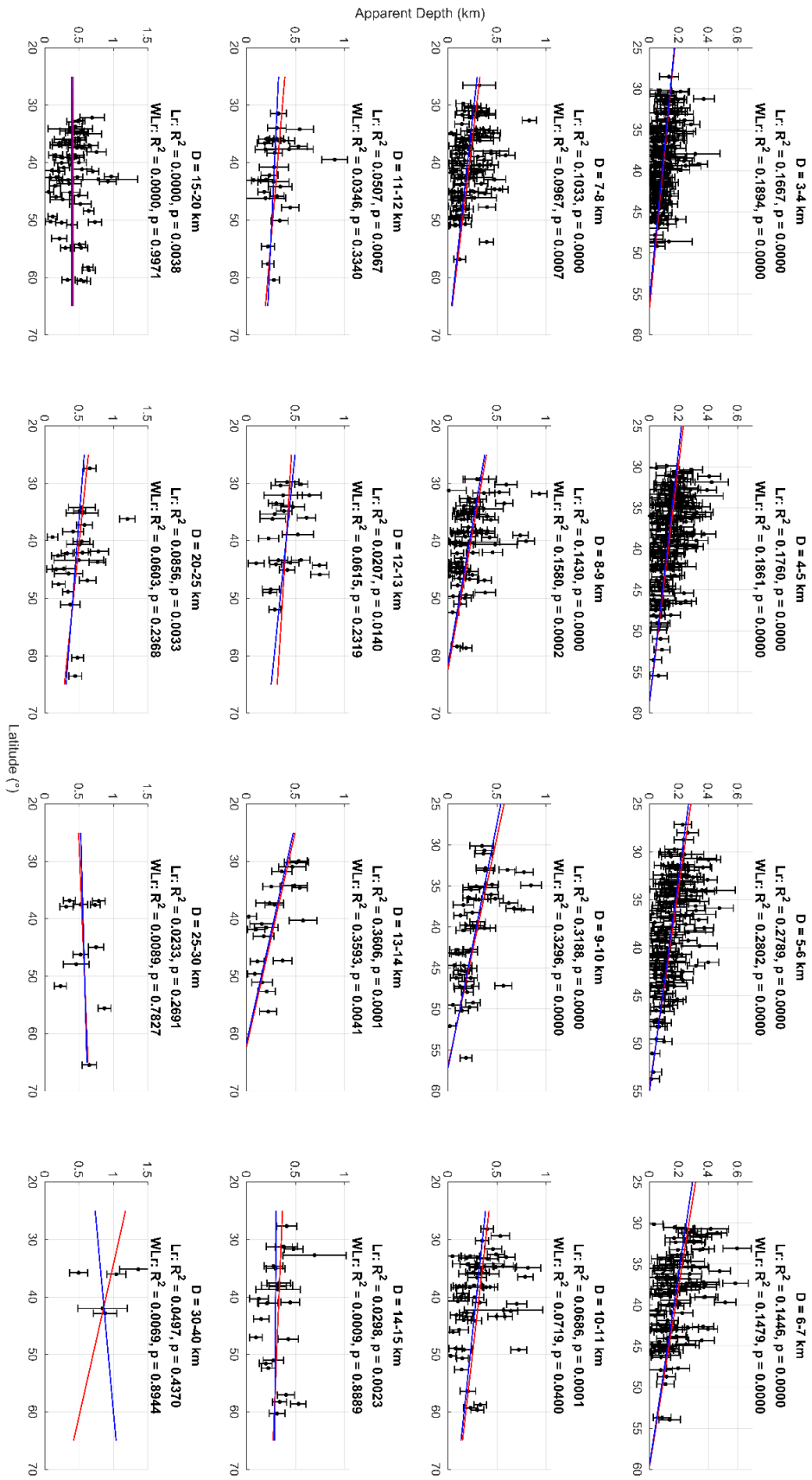


Figure 5.18. Scatter plot of d_f against latitude for CCFs in the northern hemisphere across crater-diameter bins. Each scatter plot corresponds to distinct diameter ranges outlined in Table 5.4. Red lines denote LR, while blue represents WLR. LR and WLR R^2 values, along with corresponding p -values, are labelled for each plot.

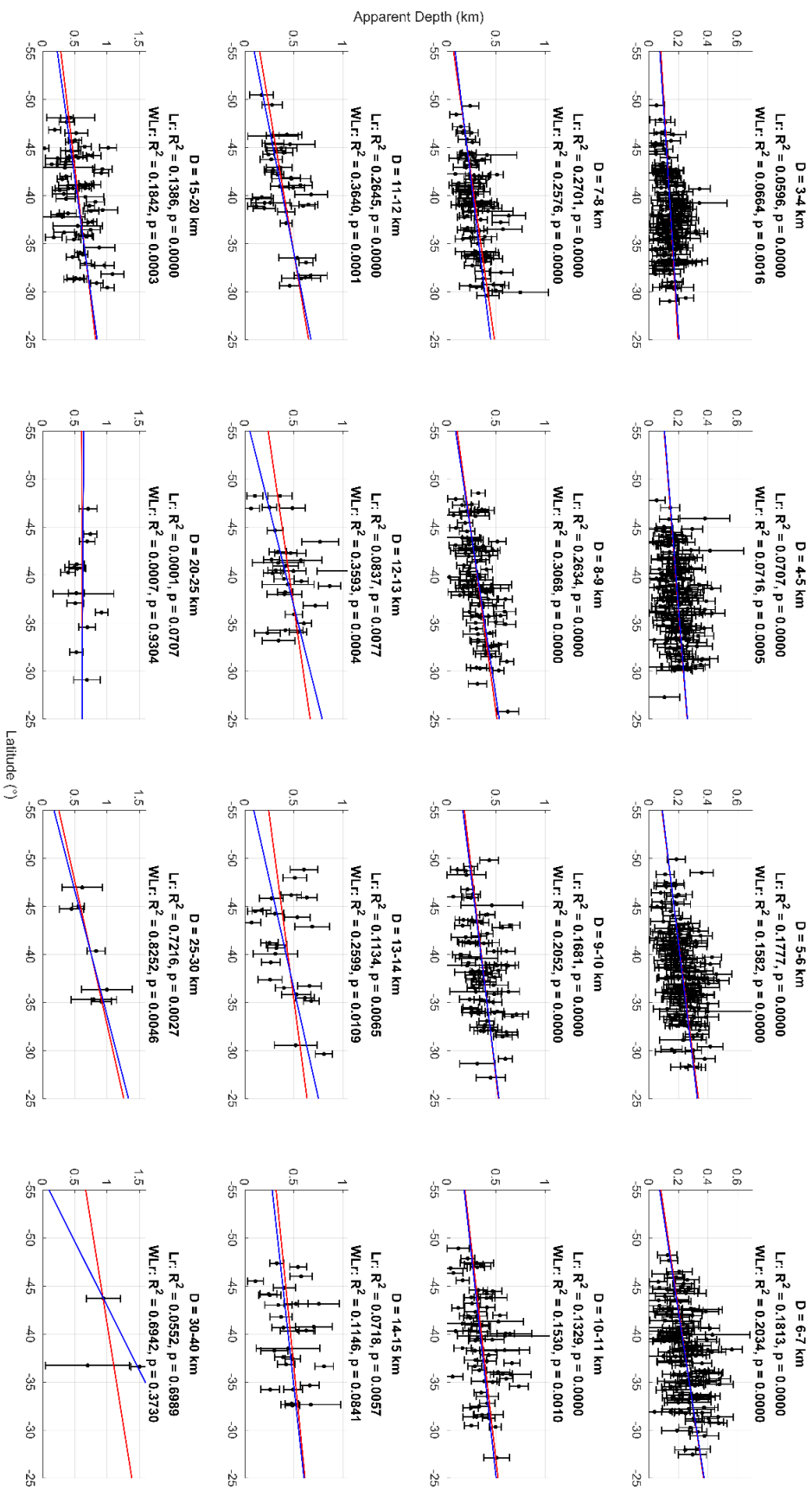


Figure 5.19. Scatter plot of d_f against latitude for CCFs in the southern hemisphere across crater-diameter bins. Each scatter plot corresponds to distinct diameter ranges outlined in Table 5.5. Red lines denote LR, while blue represents WLR. LR and WLR R^2 values, along with corresponding p -values, are labelled for each plot.

In the northern hemisphere (Figure 5.18; Table 5.4), this behaviour is most clearly expressed in bins with sufficient CCF sample sizes ($n > 25$). For smaller host craters ($D < 10$ km), the data show a pronounced decline in d_a with increasing δ , with both the upper envelope and the central distribution migrating toward lower d_a values. As D increases beyond ~ 10 km, the latitudinal gradient weakens, but the reduction in the upper range of d_a values with δ remains apparent. Spearman rank statistics support these observations, with moderate negative correlations in D bins below 10 km and weaker correlations at larger D (Table 5.4). The southern hemisphere (Figure 5.19; Table 5.5) exhibits a comparable structure when plotted against absolute δ . Within well-populated bins, d_a generally decreases toward higher δ , and the spread in d_a narrows poleward. Although individual D bins display more variability than in the north—partly reflecting smaller sample sizes—the overall structure mirrors that observed in the northern hemisphere: CCFs at higher latitudes occupy a restricted range of low d_a values, and high- d_a CCFs become increasingly scarce. Spearman rank statistics are also variable, with weak to moderate positive correlations ($r_s = 0.22$ to 0.51) that are highly significant ($p < 0.01$) in several bins, weakening as D increases, becoming statistically insignificant ($p > 0.05$).

Table 5.4. Latitude dependence of d_a in the northern hemisphere (by D bin).

D range (km)	n	Lr R^2	Lr p -value	95% Bootstrap Confidence Interval for Slope	WLr R^2	WLr p -value	Spearman's Rank	Spearman's Rank p -value
3-4	193	0.17	< 0.01	-0.0069, -0.0039	0.189	< 0.01	-0.456	< 0.01
4-5	214	0.18	< 0.01	-0.0086, -0.0051	0.186	< 0.01	-0.433	< 0.01
5-6	174	0.28	< 0.01	-0.012, -0.0077	0.28	< 0.01	-0.584	< 0.01
6-7	121	0.15	< 0.01	-0.0126, -0.0057	0.148	< 0.01	-0.404	< 0.01
7-8	115	0.1	< 0.01	-0.011, -0.0036	0.097	< 0.01	-0.39	< 0.01
8-9	82	0.14	< 0.01	-0.016, -0.0057	0.158	< 0.01	-0.465	< 0.01
9-10	56	0.32	< 0.01	-0.026, -0.012	0.33	< 0.01	-0.664	< 0.01
10-11	59	0.069	< 0.01	-0.013, -0.0016	0.072	0.04	-0.341	< 0.01
11-12	29	0.051	< 0.01	-0.011, -0.00078	0.035	0.33	-0.271	0.155
12-13	25	0.021	0.014	-0.012, 0.0067	0.061	0.23	-0.238	0.251
13-14	21	0.36	< 0.01	-0.021, -0.0073	0.36	< 0.01	-0.623	< 0.01
14-15	25	0.029	< 0.01	-0.0095, 0.0027	< 0.01	0.89	-0.258	0.211
15-20	70	< 0.01	< 0.01	-0.0055, 0.0047	< 0.01	0.99	-0.0622	0.608
20-25	25	0.087	< 0.01	-0.025, -0.0001	0.06	0.24	-0.432	0.032
25-30	11	0.023	0.269	-0.018, 0.018	< 0.01	0.78	0.0364	0.924
30-40	5	0.045	0.437	-0.67, 0.056	< 0.01	0.89	-0.3	0.683

Linear regression and Spearman's rank analysis results for apparent depth (d_a) (km) vs. latitude, binned by diameter (D) bins in the northern hemisphere.

Table 5.5. Latitude dependence of d_a in the southern hemisphere (by D bin).

D range (km)	n	Lr R^2	Lr p -value	95% Bootstrap Confidence Interval for Slope	WLr R^2	WLr p -value	Spearman's Rank	Spearman's Rank p -value
3-4	180	0.059	< 0.01	0.002, 0.0058	0.066	< 0.01	0.22	< 0.01
4-5	187	0.056	< 0.01	0.0018, 0.0076	0.072	< 0.01	0.28	< 0.01
5-6	180	0.18	< 0.01	0.0059, 0.011	0.16	< 0.01	0.45	< 0.01
6-7	156	0.18	< 0.01	0.0067, 0.013	0.2	< 0.01	0.45	< 0.01
7-8	101	0.27	< 0.01	0.0091, 0.019	0.26	< 0.01	0.51	< 0.01
8-9	100	0.26	< 0.01	0.0089, 0.018	0.31	< 0.01	0.51	< 0.01
9-10	86	0.17	< 0.01	0.0065, 0.017	0.21	< 0.01	0.41	< 0.01
10-11	68	0.13	< 0.01	0.0049, 0.018	0.15	< 0.01	0.38	< 0.01
11-12	38	0.27	< 0.01	0.0097, 0.022	0.36	< 0.01	0.39	0.015
12-13	31	0.084	< 0.01	-0.0019, 0.03	0.36	< 0.01	0.32	0.07
13-14	24	0.11	< 0.01	-0.0022, 0.029	0.26	< 0.01	0.29	0.16
14-15	27	0.072	< 0.01	-0.0029, 0.024	0.12	< 0.01	0.24	0.22
15-20	68	0.14	< 0.01	0.0073, 0.028	0.18	< 0.01	0.34	< 0.01
20-25	14	< 0.01	0.071	-0.014, 0.018	< 0.01	0.93	-0.13	0.65
25-30	7	0.72	< 0.01	0.017, 0.053	0.83	< 0.01	0.68	0.11
30-40	3	0.055	0.7	-0.054, 5.7	0.69	0.37	0.5	1
Linear regression and Spearman's rank analysis results for apparent depth (d_a) (km) vs. latitude, binned by diameter (D) bins in the southern hemisphere.								

Across both hemispheres, the latitudinal signal is most clearly expressed through constraints on the range of d_a values within each D bin. The upper envelope of d_a declines as δ increases poleward, with high d_a values becoming progressively absent and the distribution narrowing. Lower bounds in d_a vary with D and sample size; however, in well-populated bins, d_a commonly approaches zero at higher δ , consistent with near rim-full configurations. These observations show that δ limits the maximum attainable d_a while promoting convergence toward shallow d_a once the influence of the host crater is accounted.

5.3.5. Control of CCF surface elevation on apparent depth

This section evaluates whether d_a varies with elevation, or whether elevation acts as a secondary contextual constraint relative to D and δ . As established in Chapter 4, CCF spatial distribution spans a broad elevation range, with elevation-dependent restrictions emerging only toward the equatorward and poleward margins of the mid-latitude belts. Consequently, elevation is expected to exert a relatively

weak or modulating influence on d_a , in contrast to the primary controls of D (Section 5.3.2) and δ (Section 5.3.4).

Two alternative elevation measures can be defined for CCFs: (i) the mean elevation of the CCF surface itself (Z_{CCF}), and (ii) the surrounding crater-rim or ejecta elevation, intended to approximate the pre-impact surface. The latter requires assumptions about the scale over which “pre-impact” topography should be defined and is sensitive to regional relief, degradation state, and post-impact modification (Robbins and Hynes, 2012a). Given these uncertainties, Z_{CCF} is adopted here as the representative elevation metric.

Direct comparison of d_a with Z_{CCF} reveals no consistent linear relationship in either hemisphere (Figures 5.20–5.21). Spearman’s rank correlations identify weak to moderate positive associations in a small number of D bins, but these are not spatially or morphologically persistent and occur primarily where sample sizes are limited. In the northern hemisphere (Figures 5.20), only intermediate- D craters (notably 15–20 km) show statistically significant rank correlations, while larger D bins exhibit a constrained elevation range rather than a monotonic change in d_a . In the southern hemisphere, similar weak correlations appear in isolated D bins, but none persist across both hemispheres and remain robust under weighted regression.

A consistent observation in either hemisphere is the absence of very shallow d_a values at the highest surface elevations, suggesting that high-elevation settings may limit the development or long-term preservation of near rim-full CCFs. However, this effect is subtle and does not produce a continuous scaling relationship comparable to those observed for D or δ . Overall, elevation does not emerge as a first-order control on CCF d_a when the populations of each hemisphere are observed separately. Across either hemisphere, d_a spans a wide range at most elevations, and no persistent or monotonic relationship between d_a and Z_{CCF} is observed. Weak correlations identified within isolated D bins are not spatially coherent and do not persist across hemispheres, indicating that they do not reflect a general elevation-dependent control on CCF surface position. Instead, the results suggest that elevation constrains the *range* of d_a values attainable in certain geographic settings—most notably through the absence of very shallow d_a values at the highest elevations—without imposing a scaling on d_a . Elevation therefore appears to act as a secondary modulating factor on CCF geometry, rather than a primary determinant of d_a .

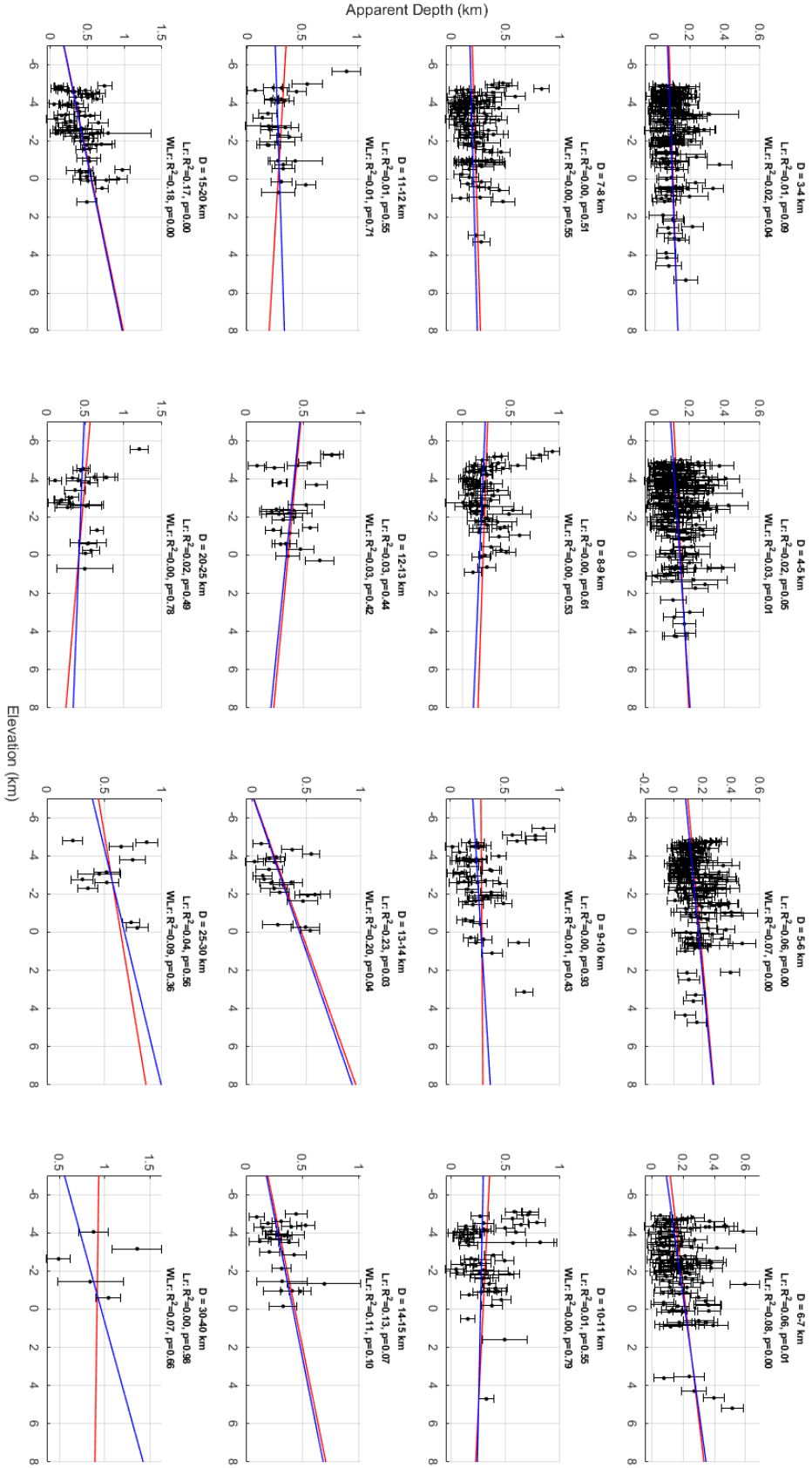


Figure 5.20 Scatter plot of d_a against elevation for CCFs in the Northern hemisphere across crater-diameter bins. Each scatter plot corresponds to distinct diameter ranges outlined in Table 5.6. Red lines denote LR, while blue represents WLR. LR and WLR R^2 values, along with corresponding p -values, are labelled for each plot.

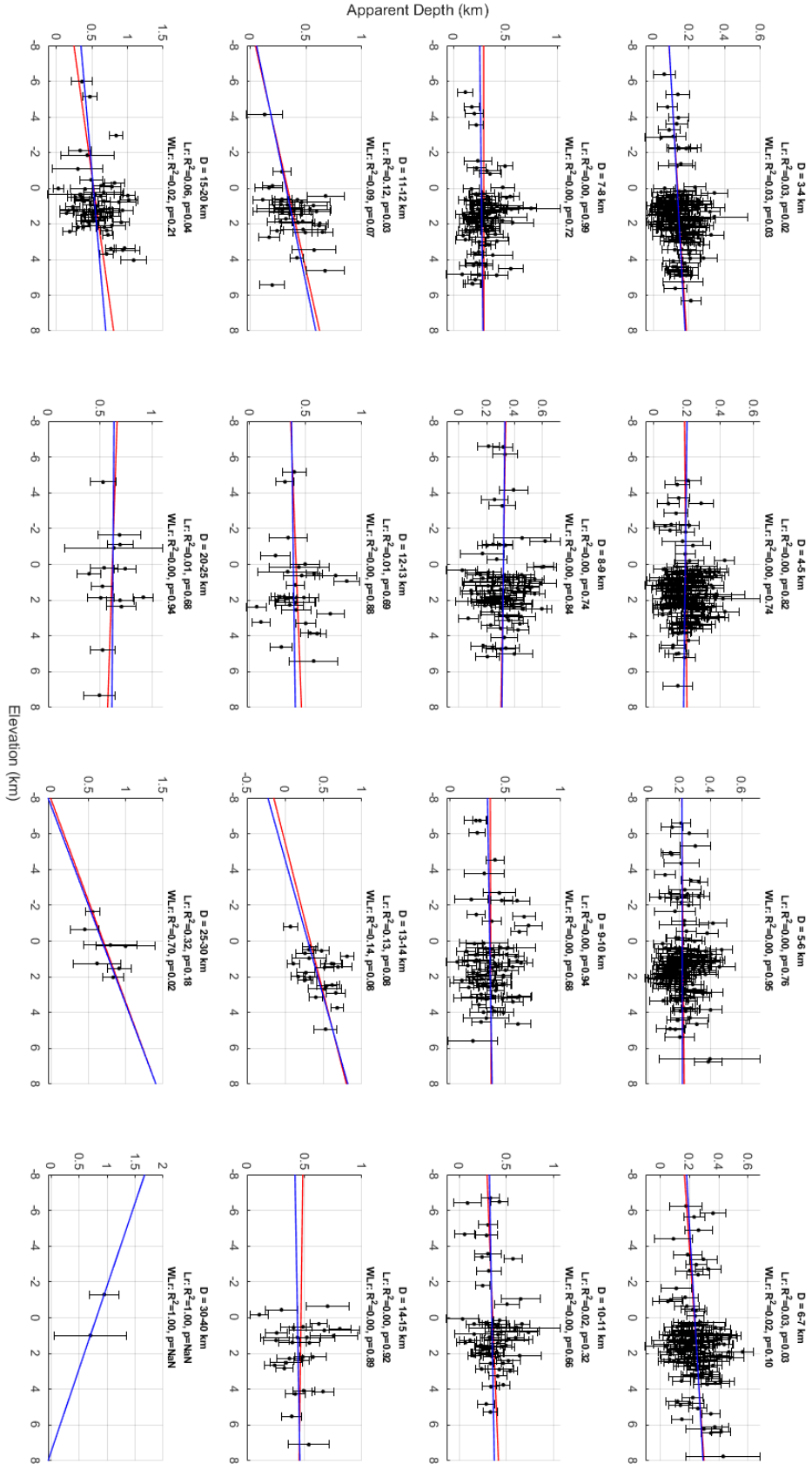


Figure 5.21. Scatter plot of d_f against elevation for CCFs in the Southern hemisphere across crater-diameter bins. Each scatter plot corresponds to distinct diameter ranges outlined in Table 5.7. Red lines denote LR, while blue represents WLR. LR and WLR R^2 values, along with corresponding p -values, are labelled for each plot.

Table 5.6. Elevation dependence of d_a in the northern hemisphere (by D bin)

D range (km)	n	Lr R^2	Lr p -value	95% Bootstrap Confidence Interval for Slope	WLr R^2	WLr p -value	Spearman's Rank	Spearman's Rank p -value
3-4	193	0.015	< 0.01	-0.000039, 0.0073	0.022	0.041	0.1	0.16
4-5	214	0.018	< 0.01	0.00051, 0.011	0.032	< 0.01	0.18	< 0.01
5-6	174	0.057	< 0.01	0.0054, 0.021	0.068	< 0.01	0.26	< 0.01
6-7	121	0.058	< 0.01	0.0019, 0.026	0.084	< 0.01	0.22	< 0.01
7-8	115	< 0.01	< 0.01	-0.012, 0.012	< 0.01	0.55	0.098	0.3
8-9	82	< 0.01	< 0.01	-0.04, 0.023	< 0.01	0.53	0.06	0.59
9-10	56	< 0.01	< 0.01	-0.044, 0.036	0.011	0.43	0.041	0.76
10-11	59	< 0.01	< 0.01	-0.041, 0.018	< 0.01	0.79	-0.019	0.89
11-12	29	0.013	< 0.01	-0.057, 0.032	< 0.01	0.71	0.028	0.89
12-13	25	0.026	0.014	-0.062, 0.041	0.03	0.42	-0.089	0.68
13-14	21	0.23	< 0.01	0.011, 0.11	0.2	0.043	0.49	0.08
14-15	25	0.13	< 0.01	-0.00027, 0.073	0.11	0.1	0.3	0.14
15-20	70	0.17	< 0.01	0.022, 0.084	0.19	< 0.01	0.39	< 0.01
20-25	25	0.021	< 0.01	-0.097, 0.042	< 0.01	0.78	-0.073	0.73
25-30	11	0.04	< 0.01	-0.14, 0.11	0.092	0.36	0.11	0.76
30-40	5	< 0.01	0.083	-0.31, 0.28	0.073	0.66	-0.1	0.1

linear regression and Spearman's rank analysis results for apparent depth (d_a) (km) vs. elevation (Z_{CCF}) (km), binned by diameter (D) bins in the northern hemisphere.

Table 5.7. Elevation dependence of d_a in the southern hemisphere (by D bin).

D range (km)	n	Lr R^2	Lr p -value	95% Bootstrap Confidence Interval for Slope	WLr R^2	WLr p -value	Spearman's Rank	Spearman's Rank p -value
3-4	180	0.029	< 0.01	0.002, 0.0058	0.026	0.031	0.18	0.014
4-5	187	< 0.01	< 0.01	0.0018, 0.0076	< 0.01	0.78	0.014	0.85
5-6	180	< 0.01	< 0.01	0.0059, 0.011	< 0.01	0.953	-0.0778	0.3
6-7	156	0.031	< 0.01	0.0067, 0.013	0.018	0.099	0.19	0.018
7-8	101	< 0.01	< 0.01	0.0091, 0.019	< 0.01	0.72	-0.078	0.44
8-9	100	< 0.01	< 0.01	0.0089, 0.018	< 0.01	0.84	-0.037	0.72
9-10	86	< 0.01	< 0.01	0.0065, 0.017	< 0.01	0.68	-0.0037	0.97
10-11	68	0.15	< 0.01	0.0049, 0.018	< 0.01	0.66	0.024	0.85
11-12	38	0.12	< 0.01	0.0097, 0.022	0.089	0.068	0.35	0.031
12-13	31	< 0.01	< 0.01	-0.0019, 0.03	< 0.01	0.89	0.06	0.74
13-14	24	0.13	< 0.01	-0.0022, 0.029	0.14	0.077	0.34	0.11
14-15	27	< 0.01	< 0.01	-0.0029, 0.024	< 0.01	0.89	-0.12	0.55
15-20	68	0.064	< 0.01	0.0073, 0.028	0.024	0.21	0.25	0.041
20-25	14	0.014	< 0.01	-0.014, 0.018	< 0.01	0.94	-0.16	0.59
25-30	7	0.32	< 0.01	0.017, 0.053	0.69	0.019	0.64	0.14
30-40	3	0.02	0.2	-0.054, 5.7	0.65	0.41	-0.5	1

Linear regression and Spearman's rank analysis results for apparent depth (d_a) (km) vs. elevation (Z_{CCF}) (km), binned by diameter (D) bins in the southern hemisphere.

5.3.6. Multilinear regression

Multi-linear regression (MLR) analysis (Table 8) was conducted, on the global CCF population to explore the combined effects of D , Z_{CCF} , and unsigned δ on d_a . Unsigned δ was used to capture the magnitude of latitudinal environmental gradients independent of hemisphere, reflecting the expectation that the primary controls on CCF mass balance vary with distance from the equator rather than with hemispheric sign.

The regression model explains approximately 50% of the variability in d_a ($p < 0.01$). D accounts for 40.8% of the variance, with δ explaining an additional 7.1% and elevation an additional 3.6%. While elevation does not significantly control d_a within a hemisphere, it exerts a detectable (small) influence across the large elevation range of both hemispheres (Z_{CCF} ranges between -6.73 to 7.75 km). Statistically, the model indicates that for every 10 km increase in D , d_a increases by ~ 300 m, a 10° increase in δ poleward reduces d_a by ~ 90 m, and an elevation increase of 5 km increases d_a by ~ 40 m. The remaining 50% is likely due to a mix of localised parameters, different crater depths or regional

trends that were not included in the model; some of it also stems from the measurement uncertainty in d_a , Z_{CCF} and D .

Table 5.8. Multilinear regression results for CCF d_a (km)

Variable	MLR slope	Std.error	T.Stat	p -value
D (km)	0.0297	0.0006	45.41	< 0.001
(abs) δ (°)	-0.0089	0.0005	-17.02	< 0.001
Z_{CCF} (km)	0.0079	0.002	3.89	< 0.001
Intercept:	= 0.427			
R^2 :	= 0.502			
RMSE:	= 0.138			
F-statistic:	= 627 p -value < 0.001			
n :	= 2494			

To assess the performance of the MLR model, residuals (calculated as observed d_a minus predicted d_a) were analysed for both unweighted and weighted MLR models. The Lilliefors test for normality yielded p -values of <0.001 for both models, leading to the rejection of the null hypothesis of normality. Additionally, residuals plotted against predicted values (Figure 5.22) exhibited a cone-shaped pattern, indicating the presence of non-constant variance, or heteroscedasticity. These findings indicate that the residuals significantly deviate from a normal distribution, violating the normality assumption fundamental to multi-linear regression analysis. To address this issue, several approaches were considered, including variance-stabilising transformations (e.g., Box-Cox, logarithmic, and square root transformations), robust standard error estimation methods, and the use of a Generalised Linear Model (see Williams et al., 2013 for a detailed discussion). Among the methods explored, only the Box-Cox transformation resulted in a noticeable improvement in the normality of the residuals, increasing the Lilliefors test p -value to 0.008. However, despite this transformation, the residuals still did not follow a normal distribution, as evidenced by the significant p -value.

Consequently, to further assess the reliability of the MLR model and account for potential non-normality in the residuals, bootstrapping was performed, with the results presented in Table 5.9 alongside the weighted MLR results for comparison. The slope coefficients for both the MLR and weighted MLR are consistent with the 95% confidence intervals obtained from bootstrapping, suggesting that the model coefficients are robust despite residual non-normality. Furthermore, the minimal discrepancy between the MLR and weighted MLR coefficients reinforces the reliability of the identified relationships;

however, the broader confidence intervals derived from bootstrapping provide a more accurate representation of the uncertainty in these estimates.

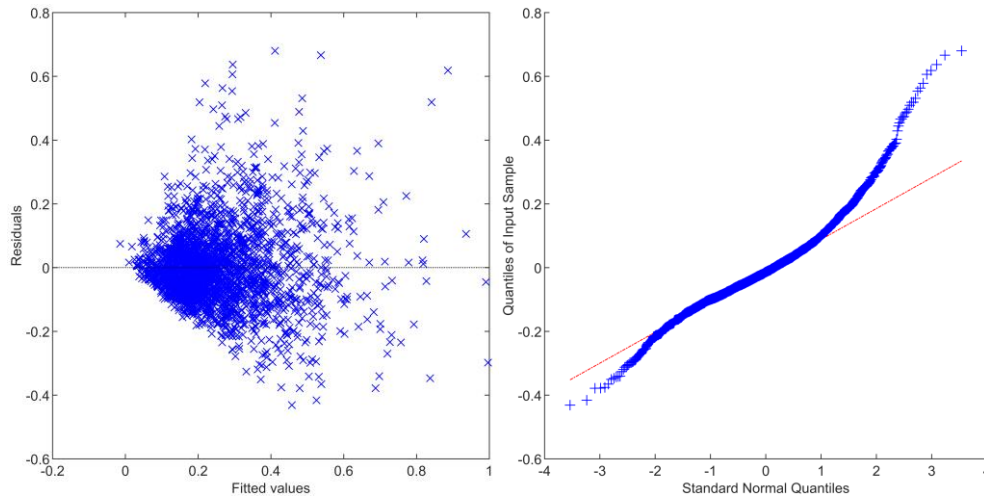


Figure 5.22. Analysis of Residuals for Multi-Linear Regression Model. (A) Residuals vs. Fitted Values, highlighting heteroscedasticity with a cone-shaped pattern in the distribution. (B) Quantile-Quantile Plot of Residuals, revealing deviations at the distribution tails, indicating potential departures from normality at extreme values.

Table 5.9. Comparison of coefficients for CCF d_a (km) using Multi-Linear Regression (MLR), Weighted MLR, and Bootstrapped MLR ($n = 1000$ iterations).

Variable	MLR		Weighted MLR		Bootstrapping (MLR)	
	slope	Std.error	slope	Std.error	95% Confidence Intervals	Std.error
D (km)	0.029	0.0006	0.0294	0.0007	0.028, 0.032	0.0009
(abs) δ ($^{\circ}$)	-0.0089	0.0005	-0.009	0.0004	-0.01, -0.0079	0.0006
Z_{CCF} (km)	0.0079	0.002	0.0095	0.0018	0.0029, 0.013	0.0027

While both the MLR and weighted MLR slope coefficients were consistent with the 95% confidence intervals obtained from bootstrapping, it is notable that the interval for elevation is relatively substantial, ranging from 0.0029 to 0.0135. So, whilst elevation control is statistically significant across the inter-hemisphere scale, there exists a higher level of uncertainty in the estimation of the elevation coefficient compared to other controls (D and δ) on d_a . This uncertainty is consistent with earlier results, indicating

that in each hemisphere, the effect of elevation on d_a is generally more subtle than that of D and δ but remains statistically significant.

Host crater d_t (incorporating the equations from Section 5.3.1), were not included as a variable, as whilst the depth of the host crater will influence d_a , crater depth variations are inherently captured by D , through the $d_t:D$ ratio, making explicit depth inclusion redundant. Additional variables, including categorical distinctions between the Northern and southern hemispheres (found to reflect hemispheric elevation differences), Geologic age of the host crater and small versus complex crater morphology, were tested but excluded from the final model, as they did not significantly improve R^2 . Their inclusion risked overfitting, reducing model interpretability without adding substantial explanatory power. Additionally, they did not improve residual normality, further supporting their exclusion.

5.4. Discussion

5.4.1. Summary of findings

This chapter sets out to identify which (if any) environmental and topographical factors govern CCF d_a once the dominant effect of host-crater is accounted for. Three first-order controls on d_a emerge from the global analysis: D exerts the dominant influence, δ imposes a consistent secondary control, and elevation contributes a weaker effect, detectable on the global scale. Together, these variables explain ~50% of the observed variance in d_a (MLR, $R^2=0.50$, $p<0.01$).

The influence of the host crater (represented here by D) is the primary predictor of d_a . The multilinear regression indicates that a 10 km increase in D corresponds to an increase in d_a of ~300 m (Figure 5.23A), with D alone accounting for 40.8% of the total variance. This reflects the primary geometric control imposed by crater size on both dt and the crater volumetric capacity, which together bound the range of permissible CCF surface elevations. The magnitude of this D signal exceeds that of all other tested variables and defines the baseline structure of the d_a distribution and must be isolated before weaker environmental influences become apparent.

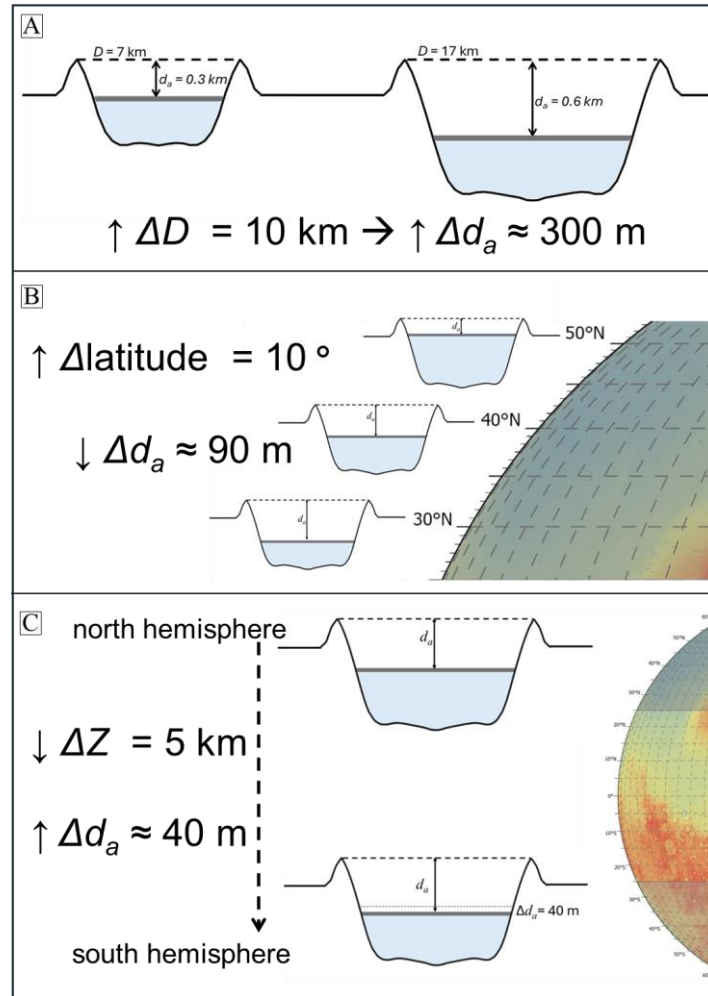


Figure 5.23. Schematic summary of the principal constraints on CCF apparent depth. (A) Geometric scaling with crater diameter, illustrating the $\sim 300 \text{ m}$ increase in d_a per 10 km increase in D . (B) Latitudinal sensitivity in the Northern Hemisphere, where a 10° poleward shift corresponds to an approximate 90 m decrease in d_a . (C) Hemispheric offset, illustrating the $\sim 40 \text{ m}$ difference in d_a associated with a $\sim 5 \text{ km}$ difference in elevation (Z) between hemispheres.

After isolating the dominant host-crater signal, δ and elevation emerge as secondary environmental controls on d_a . δ imposes a clear secondary control; with the MLR results showing a 10° poleward increase in δ corresponds to a reduction in d_a of $\sim 90 \text{ m}$ (Figure 5.23B). In both hemispheres, d_a decreases poleward and its spread narrows, implying that higher-latitude CCF-hosting craters are both more filled and more uniformly filled by CCFs. This latitudinal control is expressed similarly in both hemispheres and is independent of D . Elevation exerts a weaker but statistically significant influence on d_a at global scale (Figure 5.23C). The regression indicates that a 5 km increase in elevation corresponds to an increase in d_a of $\sim 40 \text{ m}$, accounting for an additional 3.6% of the variance. Within individual hemispheres, elevation does not emerge as a strong control on d_a ; its influence becomes detectable only when the full inter-hemispheric elevation range is considered ($Z_{CCF} = -6.73 \text{ to } 7.75 \text{ km}$). This global elevation dependence plausibly accounts for the observed north–south offset, with southern hemisphere CCFs exhibiting d_a values $\sim 50 \text{ m}$ greater than northern hemisphere CCFs at comparable D .

Approximately half of the variance in d_a remains unexplained by the global regression. These include further host-crater parameters, such as the specific modification state prior to palaeo-CCF formation and the degree of rim degradation, as well as localised microclimatic factors. The latter may include topographic shadowing from external features, the localised availability of dust or debris for overburden, and variations in the initial accumulation supply—both in terms of total ice volume and the duration. These influences operate at spatial scales smaller than those resolved by global gradients, and likely account for the significant fraction of residual variance observed within the population.

The following discussion now returns to the resolved controls identified above. Section 5.4.2 examines the physical role of the host-crater in shaping d_a . Section 5.4.3 evaluates latitudinal controls in the context of mid-latitude glaciation. Section 5.4.4 examines the influence of surface elevation; Section 5.4.5 then integrates these results with the asymmetric north–south CCF distribution established in Chapter 4. Sections 5.4.3–5.4.5 frame interpretation in terms of whether observed d_a variations reflects signals inherited from initial palaeo-CCF accumulation or signals generated by subsequent post-glacial ablation.

5.4.2. How does the host crater impact the fill of CCFs?

In this chapter, I have shown that host crater morphology and morphometry are primary controls on CCF d_a , with D explaining 40.8% of the variance in the MLR model. This dominance arises because D encapsulates the first-order geometric scaling of both d_t and cavity volume. As D increases, d_t increases and the volumetric accommodation space expands, such that higher d_a values may arise even if the absolute thickness of the CCF remains unchanged. Part of the observed increase in d_a with D may therefore reflect a geometric consequence of volume scaling rather than a change in glacial behaviour per se. Whether CCFs of comparable thickness occupy similar absolute ice volumes across crater sizes, or whether larger craters systematically host thinner or thicker ice bodies, is examined explicitly in Chapter 6.

Beyond this geometric constraint, large host craters impose additional challenges that may encode genuine glacial signals. Achieving low d_a values in large craters requires the accumulation and long-term retention of substantial ice volumes. Craters that remain far from rim-full configurations at large D may therefore do so because the volumetric demands of infilling become increasingly difficult to satisfy. This raises the question of whether such ice volumes could realistically be emplaced and preserved over geologically meaningful timescales, even under favourable climatic conditions. A volumetric framework is required to distinguish between geometric scaling and true limits on accumulation or preservation, and this is developed in the following chapter.

Preservation of large ice volumes further depends on the formation and maintenance of an insulating surface overburden. The composition of CCF overburden remains uncertain and may include atmospherically deposited dust, debris supplied from crater walls, or sublimation lag derived directly

from ice loss. If overburden formation is dominated by sublimation lag, it may be insufficient to stabilise large ice bodies before significant mass loss occurs. If instead stabilisation depends on debris or dust supply, the availability, continuity, and spatial distribution of that material may limit the ability of large CCFs to reach or maintain near rim-full states during post-glacial intervals.

An alternative explanation is that large CCFs were accumulation-limited rather than preservation-limited. Ice supply may have been insufficient in magnitude, duration, or accumulation rate to fill large craters to low d_a values, even if preservation conditions were otherwise favourable. In this scenario, elevated d_a values at large D would reflect incomplete emplacement rather than subsequent degradation. Discriminating between accumulation limits and preservation limits requires explicit consideration of ice volume, accumulation rates, and timescales, which is addressed quantitatively in Chapter 6.

Although D captures the dominant control on d_a , approximately 50% of the variance remains unexplained by the regression model. This residual variance includes uncertainty and may also reflect localised microclimate influences, additional environmental gradients not tested here, and additional host-crater influences not resolved by D alone. In the following subsections, I therefore consider two further crater-scale factors that may contribute to variability in d_a : pre-CCF crater depth modification through degradation or infilling, and the broader spatial context of the host crater.

5.4.1.1. Host Crater Degradation and variable crater depths

Many CCF-hosting craters formed during the Noachian and Hesperian and may have undergone substantial resurfacing (reducing d_i) and rim erosion (reducing Z_{rim}) prior to the onset of Amazonian mid-latitude glaciation. Estimates of Noachian crater modification suggest that typical crater depths may have been reduced by several hundred metres during the Noachian (~350 m; Boyce and Garbeil, 2007), well before the period of CCF formation. After this interval, erosion rates declined markedly (Craddock and Howard, 2002), limiting the magnitude of subsequent large-scale modification during the Amazonian.

Quantifying the degree of pre-CCF crater degradation or infilling for individual craters is not currently feasible. Morphological classifications alone do not uniquely diagnose cumulative erosion, and the presence of both Noachian and Hesperian craters with fresh or near-pristine morphologies indicates that some ancient craters can retain deep cavities over geological timescales (Carr and Head, 2010). Nevertheless, it is unlikely that all host craters entered the CCF-forming interval with identical effective depths. Localised erosion, sediment infilling, mantling, and earlier ice-related deposition would therefore have modified crater cavities to differing degrees across both hemispheres, introducing additional scatter into the relationship between present-day d_a and D . This pre-conditioning of crater geometry provides a physically plausible, but unquantifiable, contribution to the residual variance in d_a that remains after accounting for D alone.

5.4.1.2 Background Topography and Spatial Context of the Host Crater

As discussed in Section 5.2.2, host craters are not always isolated topographic features but are frequently embedded within complex terrain, including plateau margins, valley walls, degraded crater rims, and regional slopes. Some CCF-hosting craters also occur within the rims or ejecta blankets of larger impact structures. These settings introduce local variations in slope, shadowing, wind exposure, and sediment supply that may influence both the accumulation and long-term preservation of crater-confined ice.

Such local topographic context can modify surface energy balance, sediment redistribution, and debris availability at the scale of individual craters, potentially shifting the equilibrium surface position of a CCF within its host cavity. These effects are inherently local and cannot be isolated or parameterised in a global analysis such as this chapter. Nevertheless, spatial context provides a physically plausible additional source of crater-to-crater variability in d_a beyond that imposed by D alone.

5.4.3. Mid latitude glaciation and latitudinal controls

5.4.3.1 Overall trend: d_a decreases as δ increases

In both hemispheres, CCF d_a decreases poleward. A leading hypothesis for the cause of this is decreasing insolation rates towards the pole, which results in reduced ablation rates and favouring the preservation of thicker ice deposits (smaller d_a) at higher latitudes.

Here, I examine this idea, drawing on theoretical insolation curves for different planetary obliquity. CCFs are believed to have formed during the Amazonian glaciation period at intermediate ($\sim 35^\circ$) obliquity periods (Laskar et al., 2004; Head et al., 2006b; Madeline et al., 2009). Ward (1974) (Figure 5.24A) calculated the δ distribution of solar insolation under different obliquity. The average yearly insolation (I) varies with δ on a flat surface as:

$$I = \frac{S}{2\pi^2} (1 - \varepsilon^2)^{-\frac{1}{2}} \int_0^{2\pi} \{1 - (\sin\delta \cos\theta - \cos\delta \sin\theta \sin\varphi)^2\}^{\frac{1}{2}} d\varphi \quad 5.1$$

where φ is the diurnal variation, θ is the obliquity, ε is eccentricity and S is Mars' solar constant at 1.5 AU.

At each δ , I weakly depend on ε , which drives hemispheric asymmetries (Ward, 1974). However, to isolate the primary controls of δ and θ on ice stability, I assume a circular orbit ($\varepsilon = 0$) for the calculations in Figure 5.24. This removes the confounding influence of precession, resulting in idealised, hemispherically symmetric insolation curves. Figure 5.24A presents the resulting insolation curves plotted across 0° to 90° δ for a range of obliquity values. Figure 5.24B plots average annual insolation against obliquity for the specific latitudes occupied by CCFs ($\sim 25^\circ$ to 65°). In both instances, I is normalised to S .

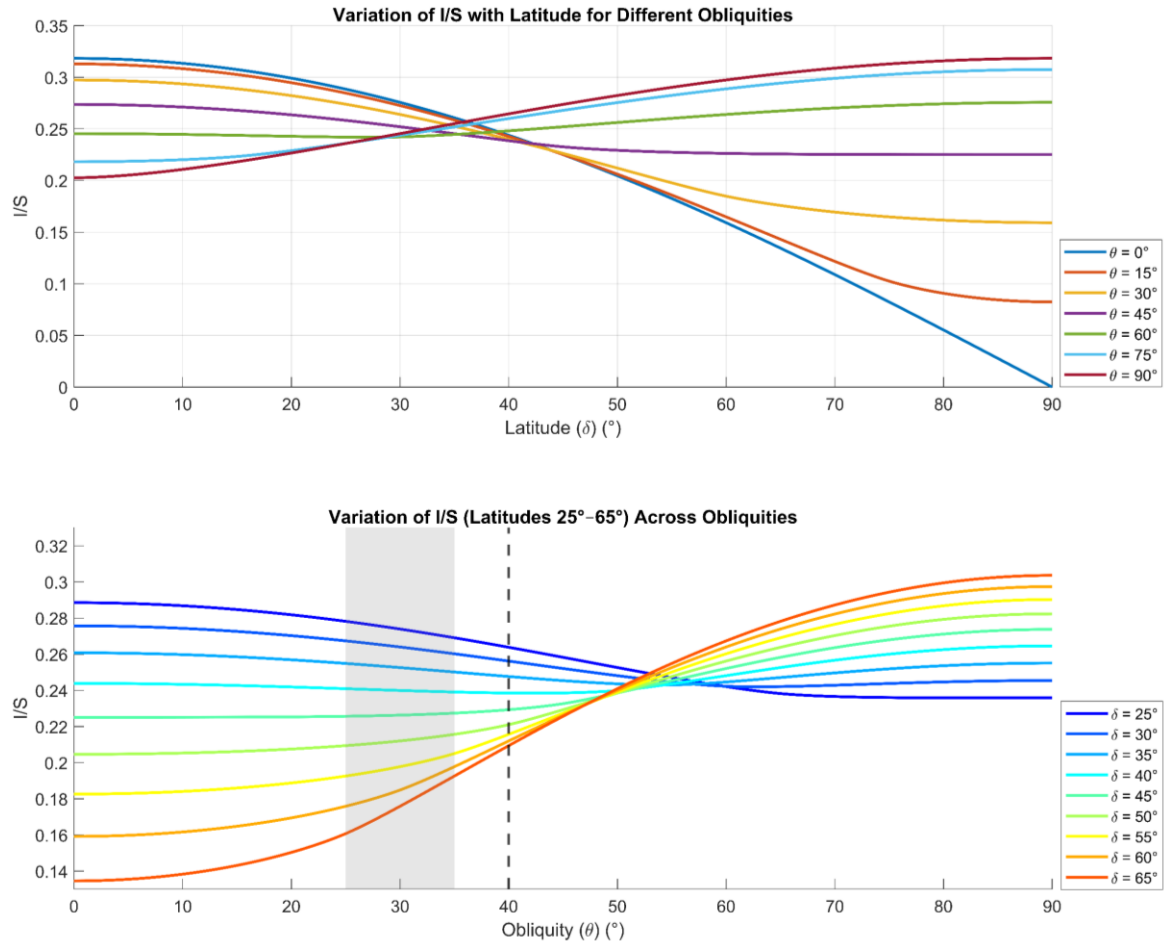


Figure 5.24. Annual variation of insolation (I/S) for different obliquities. (A) Variation of I/S with latitude across different obliquities. (B) Annual variation of I/S within the CCF latitude range for varying obliquities. The shaded box highlights intermediate obliquities ($\sim 25^\circ$ to 35°), as proposed by Madeleine et al. (2009) for the mid-latitude Amazonian glaciation. The black dashed line indicates the most likely obliquity of 40° , as suggested by Laskar et al. (2004). For both graphs, insolation is adjusted by the Solar constant ($S = 1.52$ AU), and eccentricity is set to 0 to eliminate hemispheric differences. Calculated using Equation 5.1, derived from Ward (1974).

Within the latitude range occupied by CCFs, mean annual insolation decreases systematically with increasing absolute latitude for all obliquities thought to be relevant to Amazonian mid-latitude glaciation (i.e., ~ 25 – 35° , and the most likely obliquity of 40° through Mars history; Madeleine et al., 2009; Laskar et al., 2004), and this behaviour persists up to obliquities of at least ~ 50 – 55° (Figure 5.24B). This monotonic poleward reduction in available solar energy provides a physically consistent explanation for the observed decrease in d_a toward higher latitudes.

Lower mean insolation at higher latitudes implies reduced net ablation rates and/or more favourable long-term mass balance, allowing CCF surfaces to remain closer to the crater rim and thus to preserve smaller d_a values. This effect is consistent with the multilinear regression results, which indicate that a 10° poleward shift in latitude corresponds to a reduction in d_a of approximately 90 m. Crucially, this latitudinal control does not depend on whether present-day d_a primarily reflects initial accumulation

geometry or subsequent post-glacial modification: under either interpretation, reduced insolation at higher latitudes favours the preservation of more rim-proximal CCF surfaces. The consistency of this relationship across both hemispheres further supports latitude as a first-order environmental control acting through surface energy balance rather than through hemispheric asymmetries or longitudinal forcing.

5.4.3.2. Increase in uniformity of d_a as latitude increases poleward in both hemispheres.

As δ increases, the dispersion of d_a values decreases in either hemisphere (Figure 5.17, 5.18, 5.19). In d_a –latitude space, the CCF population defines a distinctly wedge-shaped envelope, in which low-latitude craters exhibit a wide range of fill states, while high-latitude craters cluster into a much narrower band of d_a values (Figure 5.17). Two related features characterise this convergence: first, the range of d_a values decrease poleward; second, high-latitude CCFs lacking substantial infill (i.e., with large d_a) become increasingly rare, except at the highest latitudes discussed in Section 5.4.3.3. The following subsections examine this behaviour under two complementary interpretations: first, as a signal of latitudinal differences in palaeo-CCF accumulation during glacial phases; and second, as a consequence of latitude-dependent post-glacial ablation and preservation.

One plausible explanation for the poleward convergence of d_a is that it reflects the increasing dominance of latitude over regional- and crater-scale controls on CCF mass balance, regardless of whether d_a primarily encodes spatial variability in palaeo-accumulation or differential post-glacial ablation and preservation. In either case, the same first-order control applies: mean insolation decreases monotonically with latitude over the obliquity range relevant to Amazonian mid-latitude glaciation (Section 5.4.3.1), progressively reducing the absolute energy available for ablation and thereby compressing the range of viable mass-balance outcomes.

Under this framework, higher insolation at lower latitudes may either (i) inhibit some palaeo-CCFs from ever attaining near-rim-full conditions during accumulation, or (ii) enhance post-glacial ablation such that equilibrium surfaces stabilise at larger d_a values. Conversely, at higher latitudes, reduced insolation both favours greater net preservation during accumulation phases and suppresses post-glacial mass loss, forcing CCFs toward a narrower range of long-term fill states regardless of which process dominates.

This latitude-dominated energy regime may also progressively weaken the influence of secondary controls such as elevation, rim geometry, wall shadowing, and mesoscale setting. At low latitudes, these localised factors may exert a strong leverage on CCF mass balance. For example, a crater with a sharp or elevated rim may preserve a near-rim-full CCF surface through enhanced shadowing, whereas a neighbouring crater lacking such protection may experience substantially greater ice loss and stabilise at much larger d_a . At higher latitudes, by contrast, the background radiative forcing is uniformly low, so

even craters with less favourable local geometry experience limited ablation. The net effect is a convergence of long-term mass balance and therefore a convergence of d_a toward a narrower interval.

A contrasting low-latitude regime follows directly from the same logic. There, higher mean insolation increases the sensitivity of both accumulation and ablation rates to small differences in local conditions. As a result, CCFs span a much wider range of evolutionary pathways: some systems may never have achieved near-rim-full states because accumulation was insufficient, while others may have done so but subsequently lost substantial ice. This dual sensitivity naturally produces the broad dispersion of d_a values observed at lower latitudes, in contrast to the tightly clustered population at higher latitudes.

An alternative explanation is that the poleward convergence of d_a reflects latitudinal signals of net accumulation regimes during mid-latitude glaciation and paleo-CCF formation. This interpretation is evaluated here using the northern hemisphere only, because in the southern hemisphere there is a pronounced mismatch between CCF distribution, observed d_a values, and the net accumulation fields predicted by the GCM simulations of Madeleine et al. (2009).

In the northern hemisphere, the GCM simulations of Madeleine et al. (2009), for moderate obliquity (35°) and elevated eccentricity (0.1), predict an accumulation regime that becomes increasingly regionally constrained at higher latitudes (Figure 5.25). While there is significant spatial variability in net accumulation rate at all latitudes, the climatic envelope within which CCFs can persist narrows poleward. Under this climate state, high-latitude CCFs ($>45^\circ$ N) are confined to moderate-to-high net accumulation zones of $\sim 4\text{--}12\text{ mm yr}^{-1}$, whereas lower-latitude CCFs are distributed across the full modelled accumulation spectrum, from near-zero to $\sim 16\text{ mm yr}^{-1}$ (Figure 5.25A).

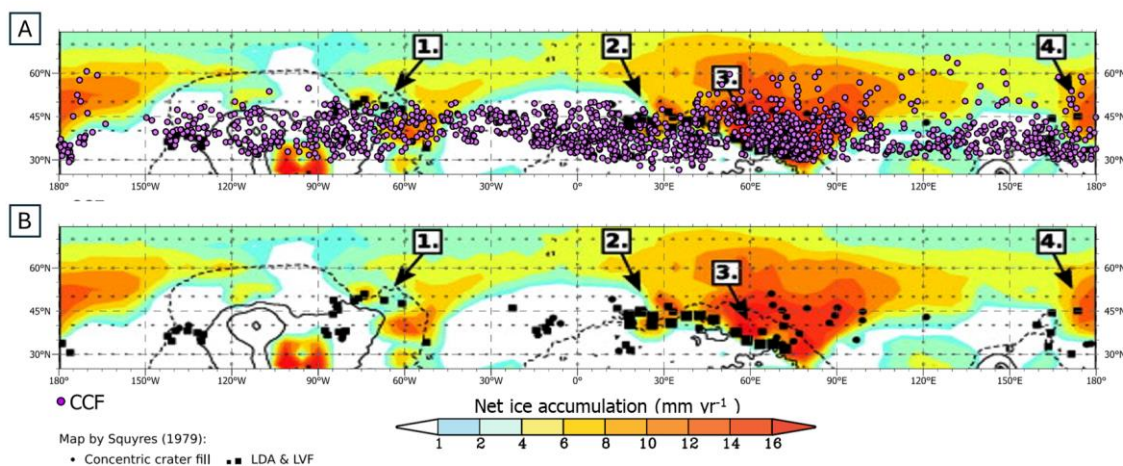


Figure 5.25. North hemisphere CCF distribution superposed on modelled net ice accumulation under an intermediate-obliquity climate state. (A) Distribution of CCFs (purple) superposed on net ice accumulation (mm yr^{-1}) predicted by the GCM of Madeleine et al. (2009). (B) The same modelled net ice accumulation field shown without the CCF overlay. The climate simulation uses an obliquity of 35° , eccentricity of 0.1, perihelion at $L_s = 270^\circ$, and a dust opacity of 2.5. Modified and redrawn from Madeleine et al. (2009).

Although this represents only a single snapshot within the range of orbital states likely experienced during the Amazonian (Section 4.4.3), it implies a fundamental difference in the climatic envelopes occupied by CCFs at different latitudes. High-latitude CCFs in the north are restricted to a relatively narrow and consistently favourable accumulation regime, whereas lower-latitude CCFs persist across a much broader range of net accumulation conditions. This climatic filtering provides a natural explanation for the observed reduction in d_a dispersion poleward. In this framework, many low-latitude CCFs are likely to include systems that never attained or sustained near-rim-full conditions because accumulation rates were frequently too low to offset ablation over long timescales. By contrast, the poleward population of CCFs is preferentially drawn from craters that experienced persistently favourable mass-balance conditions during their formation, leading to a more uniform degree of infill and hence a narrower range of d_a values.

A third control does not explain the convergence itself but helps explain why high-latitude CCFs do not converge toward perfectly rim-full states. Global analyses of near-surface relative humidity identify persistent oversaturated equilibrium regions across much of the northern mid-latitudes between $\sim 25^\circ$ and 55° (Pál et al., 2019; Figure 5.26), including Amazonis, Alba Patera, Tharsis, Elysium, and Arabia Terra. No comparable stable high-humidity regions occur poleward of $\sim 55^\circ$ (Figure 5.26), implying systematically drier conditions for CCFs in the $\sim 55^\circ$ – 65° band. Lower humidity increases the atmospheric capacity for water vapour, enhancing vapour gradients and promoting sublimation. Mesoscale modelling further indicates that regolith–atmosphere interactions within craters can reduce vapour exchange between crater floors and the overlying atmosphere (Steele et al., 2017), acting as a partial buffer against sublimation. This provides a plausible mechanism for maintaining non 0 d_a values in a subset of high-latitude craters, despite the otherwise strongly favourable radiative conditions for CCF preservation.

Surface relative humidity

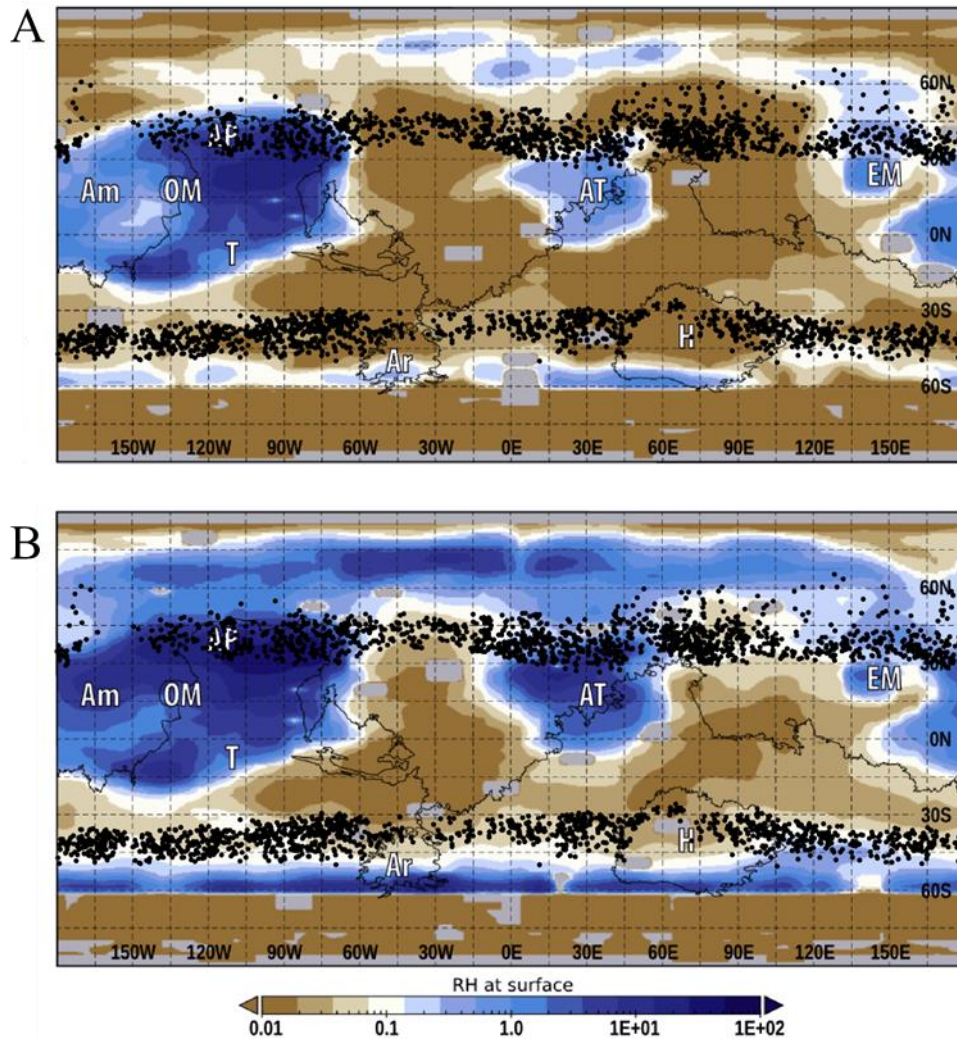


Figure 5.26. Global Regions of Elevated Near-Surface Relative Humidity and CCF Distribution. A) Relative humidity calculated with 2 am vapor data coupled with 2 am TES surface temperature and pressure values. B) Nighttime TES measurements with 2 pm GCM vapor data with the nighttime TES measurements (see Pál et al., 2019 for model parameters). Regardless of using daytime or nighttime water vapor data, the global distribution of elevated RH regions remains consistent in the model, as does the distribution of north hemisphere high latitude CCFs, which are predominantly situated in areas of higher relative surface Pressure. In comparison, the majority of South hemisphere CCF population consistently remains within regions of low relative humidity. Edited from Pál et al., 2019.

I interpret the poleward narrowing of d_a as a transition from a low-latitude regime dominated by local controls and variable accumulation histories to a high-latitude regime dominated by climatic admissibility and energy-balance constraints. Obliquity cycling and low atmospheric humidity then regulate the upper bound of preservation, preventing universal convergence to rim-full states. This framework explains both the wedge-shaped envelope of the CCF population and the persistence of a small number of anomalously inset high-latitude examples, discussed briefly in the next section.

5.4.3.3 Hypothesising causes for the anomalous increase in d_a at high northern latitudes

In the highest-latitude band ($>55^\circ$), CCFs exhibit slightly larger d_a values than those between 50° and 55°N , despite lying in a zone that should favour both palaeo-CCF formation and long-term preservation under either accumulation- or ablation-based interpretations (Figure 5.17A). This anomalous behaviour involves only a small subset of the population (~ 25 CCFs), which occupy a narrow and systematically low elevation range (-4.77 to -3.66 km), whereas CCFs between 50° and 55°N span a much broader elevation interval (-4.86 to -1.55 km). Given the regression results in Section 5.3.7, this combination of high latitude and low elevation should, in principle, favour the smallest d_a values in the dataset. The fact that these CCFs instead retain moderately inset surfaces therefore indicates that additional processes have acted to limit their long-term degree of infill.

A first possibility is modulation by extreme obliquity states. Within the latitude range occupied by CCFs, mean annual insolation decreases monotonically with increasing absolute latitude for all obliquities thought to be relevant to Amazonian mid-latitude glaciation ($\sim 25\text{--}35^\circ$, and the long-term mean of $\sim 35\text{--}40^\circ$), and this behaviour persists up to obliquities of $\sim 50\text{--}55^\circ$ (Figure 5.24B; Ward, 1974; Laskar et al., 2004; Madeleine et al., 2009). However, at still higher obliquities ($>55^\circ$), this monotonic relationship breaks down, and the highest latitudes experience strongly enhanced seasonal and annual insolation. Although such extreme obliquities are not predicted for the last ~ 20 Myr, orbital solutions extended to ~ 250 Myr indicate that Mars has undergone large-amplitude chaotic obliquity variations up to a maximum $\sim 65^\circ$ (Laskar et al., 2004), with episodic excursions of $>60^\circ$ statistically likely within the past ~ 1 Gyr (Ward, 1973; Touma and Wisdom, 1993; Laskar et al., 2004). If such states occurred after CCF emplacement, the $>55^\circ\text{N}$ population would have experienced disproportionately high insolation relative to slightly lower latitudes (e.g., 50°), potentially driving enhanced ablation or inhibiting complete infill and leaving a residual signal in the form of systematically larger d_a values.

A second, complementary factor may be atmospheric humidity. Global analyses of near-surface relative humidity indicate that persistent oversaturated equilibrium regions are largely confined to $\sim 25\text{--}55^\circ\text{N}$ and do not extend into the $>55^\circ\text{N}$ latitude band (Pál et al., 2019; Figure 5.26), implying systematically drier conditions at the highest latitudes. Lower humidity increases the atmosphere's capacity to accept water vapour, steepening vapour gradients and enhancing sublimation rates. Mesoscale modelling further suggests that regolith–atmosphere interactions within craters can buffer vapour exchange (Steele et al., 2017), but this buffering is likely to be less effective under low-humidity conditions, further increasing susceptibility to net ice loss.

In combination, these effects offer a physically consistent explanation for why a small subset of the highest-latitude northern CCFs retain anomalously large d_a values despite otherwise favourable radiative and topographic conditions. Rather than contradicting the general poleward convergence of

d_a , these outliers may record the imprint of episodic obliquity-driven instability and enhanced sublimation efficiency under an unusually dry climatic regime.

5.4.4. The global elevation gradient in apparent depth: physical forcing or hemispheric signal?

On Earth, glacier survival generally increases with altitude, as net surface accumulation increases (driven by an increase in snow and ice from direct precipitation), and ablation decreases (due to the decrease in mean temperature with air density) (Benn and Evans, 2010). Curiously, CCFs on Mars exhibit the opposite behaviour. d_a increases with elevation, such that higher-elevation craters are less well-filled by CCFs. The global regression indicates that a 5 km increase in elevation corresponds to an increase in d_a of ~ 40 m. This effect is weak relative to D and δ , but it is statistically significant when the full inter-hemispheric elevation range is considered. Given a mean elevation difference of ~ 6 km between the hemispheres, the elevation term accounts for ~ 40 m of the observed north–south offset in d_a (Section 5.3.5), indicating that southern hemisphere CCFs are systematically less well filled than their northern hemisphere counterparts at comparable D . This inverse elevation dependence contrasts with terrestrial glacial systems and indicates that, for CCFs, higher elevations are associated with conditions that favour reduced ice preservation rather than initial accumulation. However, because elevation covaries with the hemispheric dichotomy, this relationship may alternatively reflect hemispheric differences in glacial history rather than a direct elevation-dependent control.

To date, and to the author’s knowledge, only one study has explicitly examined the relationship between CCF elevation and infilling. Weitz et al. (2018) analysed 22 CCFs in Utopia Planitia, comparing inferred fill depth percentage $((d_r - d_a)/d_i) \times 100$ with crater bed elevation, defined as $Z_{\text{rim}} - d_i$. They reported that craters at lower elevation are less filled, finding that craters smaller than ~ 15 km are typically 26–94 % filled, whereas larger craters are commonly filled by < 60 %. On this basis, they inferred that CCFs at lower elevations experienced reduced net accumulation.

When an equivalent comparison is performed here—relating fill fraction to crater-floor elevation inferred from d_r —the association is negligible when the hemispheres (Figure 5.27A–B) are considered separately (North: $R^2 = 0.015$, $p = 1.9 \times 10^{-5}$; South: $R^2 = 0.006$, $p = 5.3 \times 10^{-3}$). However, when both hemispheres are combined (Figure 5.27C), a substantially stronger relationship emerges ($R^2 = 0.153$, $p = 7.7 \times 10^{-92}$), with fill fraction decreasing with increasing elevation. This behaviour mirrors the elevation dependence identified in this chapter MLR results (Section 5.3.6), demonstrating that elevation exerts a detectable influence on CCF infilling only when the full inter-hemispheric elevation range is sampled. The apparent discrepancy with Weitz et al. (2018) therefore most likely reflects differences in spatial scope and in the quantities being compared. Their results capture local behaviour within the low-elevation Utopia Planitia basin, whereas the signal identified in this chapter emerges only at global scale.

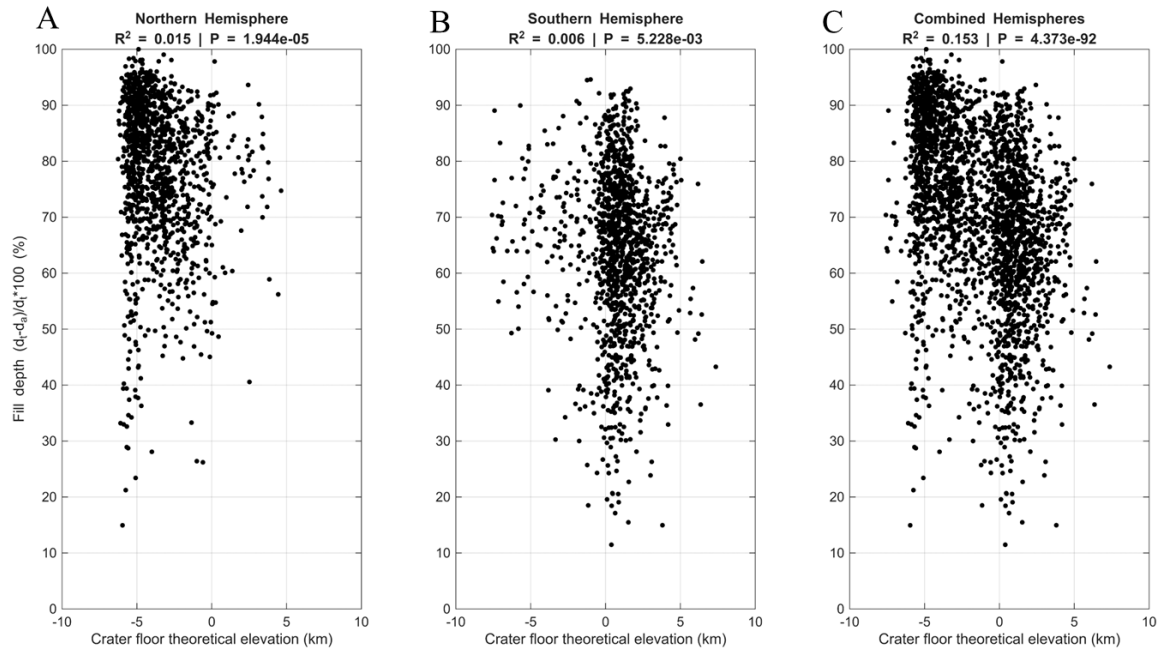


Figure 5.27. Correlation between CCF fill fraction and inferred crater-floor elevation. Scatter plots show estimated fill fraction, expressed as $(d_i - d_a)/d_i * 100$, plotted against theoretical crater-floor elevation (defined as $Z_{rim} - d_i$). Panels show results for (A) the Northern Hemisphere, (B) the Southern Hemisphere, and (C) the combined dataset. For each panel, the coefficient of determination (R^2) and associated p -value from linear regression are reported on the plot. When the hemispheres are considered separately, the association between fill fraction and crater-floor elevation is negligible (North: $R^2 = 0.015$, $p = 1.9 \times 10^{-5}$; South: $R^2 = 0.006$, $p = 5.3 \times 10^{-3}$), whereas combining both hemispheres produces an apparent correlation

Having established elevation as a weak but detectable control on d_a , I now consider the physical origin of this inverse trend. One possibility is that the signal reflects hemispheric differences in glacial history rather than a direct elevation-dependent process, given the pronounced north–south elevation contrast imposed by the crustal dichotomy. In this view, differences in d_a could arise from contrasts in accumulation or ablation histories between hemispheres—for example, more recent or more sustained ice accumulation in the north or reduced accumulation and/or enhanced post-glacial ablation in the south—without elevation itself being the causal driver. Under this interpretation, elevation functions as a proxy for hemispheric context rather than as a controlling mechanism, and the observed correlation should not be taken to imply causation. The fact that the elevation signal emerges only when both hemispheres are analysed together, and not within individual hemispheres, is consistent with this interpretation. If, however, the elevation dependence is physically real, it must operate at global scale rather than regionally, in line with its absence in intra-hemispheric analyses.

As it is currently not possible to distinguish unambiguously between these alternatives, I next evaluate a process-based explanation in which elevation influences CCF preservation at the global scale through sublimation efficiency. In this framework, higher elevations are expected to experience higher

sublimation rates, while lower elevations experience reduced sublimation due to higher surface pressure and a denser CO₂ atmosphere. Ingersoll (1970) formalised this behaviour through parameterisations for buoyancy-driven (E_b) and turbulence-driven (E_t) sublimation, later reformulated by Haberle and Jakosky (1990), which I apply here to assess whether elevation-dependent sublimation provides a physically plausible explanation for the observed increase in d_a with elevation.

Ingersoll (1970) derived the following expressions for E_b and E_t sublimation, as presented by Haberle and Jakosky (1990):

$$E_b = (0.17)\rho_w D \left[\frac{\left(\frac{\Delta\rho}{\rho}\right)g}{v^2} \right]^{\frac{1}{3}}, \quad (5.2)$$

$$E_t = 0.002\rho_w U(1 - r). \quad (5.3)$$

Here ρ_w is the saturation density of water vapor; D is the diffusion coefficient of water vapor in carbon dioxide; g is the acceleration of gravity on Mars; v is the kinematic viscosity of carbon dioxide; $\left(\frac{\Delta\rho}{\rho}\right)$ is the difference between the density of the ambient gas (mostly CO₂) and that of the gas at the surface, divided by the density of the gas at the surface; U is the wind velocity 1 m above the surface and r is the near-surface relative humidity (Ingersoll, 1970; Haberle and Jakosky, 1990). $\left(\frac{\Delta\rho}{\rho}\right)$ is calculated as

$$\left(\frac{\Delta\rho}{\rho}\right) = \frac{(m_e - m_w)e}{[m_e P_0 - (m_e - m_w)e]}, \quad (5.4)$$

with e being the saturation vapor pressure of ice at temperature T_0 ; P_0 is Martian surface pressure; m_w and m_e are the molecular weight of H₂O and CO₂.

The sublimation flux formulations express E_b and E_t as functions of atmospheric density contrasts, diffusion, wind speed, and temperature. In particular, the buoyancy term depends on the fractional density contrast which scales inversely with surface pressure (P_0). Equation (5.4) shows that lower elevations, characterised by higher P_0 , yield smaller $\frac{\Delta\rho}{\rho}$ values and directly suppressing E_b sublimation (Eq. 5.2). The denser atmosphere at low elevations also inhibits upward molecular diffusion of H₂O vapour, further limiting sublimation (Ingersoll, 1970). In contrast, higher elevations experience lower P_0 , larger $\frac{\Delta\rho}{\rho}$, and more efficient vapour transport, resulting in higher sublimation rates. Ingersoll's framework therefore predicts a monotonic increase in sublimation efficiency with elevation, all else being equal.

Temperature modulates these fluxes through its control on ρ_w and e , such that warmer surfaces enhance sublimation. To evaluate whether temperature offsets the pressure effect, Figure 5.28 (adapted from

Piqueux et al., 2024) shows median surface temperatures across the mid-latitudes derived from seven years of Mars Climate Sounder observations. The dominant control on temperature is seemingly latitudinal rather than elevational, although extreme topographic highs (e.g., Olympus Mons, Elysium Mons, Thaumasia Highlands) exhibit substantially cooler temperatures capable of offsetting the pressure-driven increase in sublimation. Outside such extremes, present mid-latitude surface temperatures are primarily governed by surface thermophysical properties and illumination geometry rather than elevation alone (Piqueux et al., 2024). Consequently, when considered at the global scale—where elevation differences are large and persistent—the pressure dependence dominates, yielding higher net sublimation at higher elevations. This behaviour is consistent with the regression results and provides a physically coherent explanation for why CCFs at higher elevations are, on average, less well filled and exhibit larger d_a values.

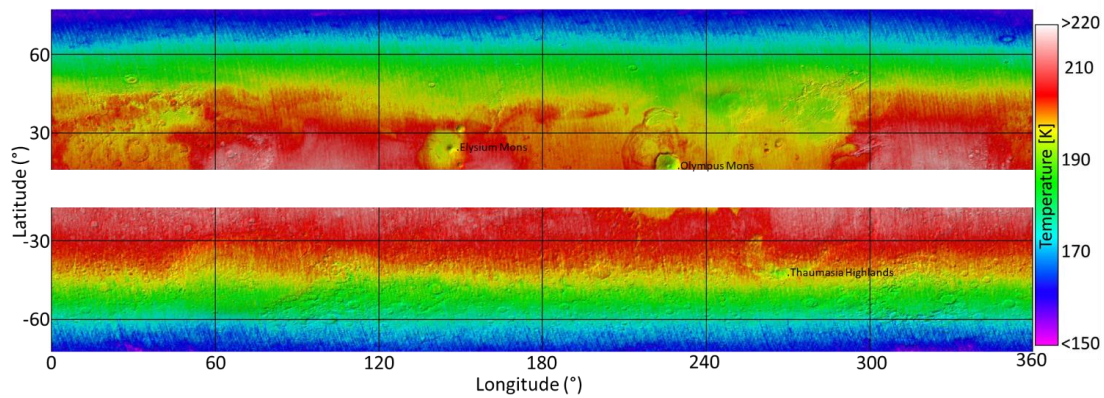


Figure 5.28. Average Martian surface Temperature over the last 16 years. Piqueux et al. (2024)

calculated median yearly AM and PM temperature maps from MY 29 to 36 of Mars Climate Sounder observations. At each location, the AM and PM median maps were then averaged together to form this figure. Note that this final map is not equivalent to a true annual average temperature map due to diurnal and seasonal biases. The map has been modified to show just the CCF hemispheric distribution. The background is a shaded relief from MOLA. Modified from Piqueux et al. (2024).

5.4.5. Can the exploration into apparent depth explain the north–south differences in latitudinal distribution of CCFs?

One of the overarching research questions this Thesis seeks to explore is, why are CCFs present in some craters and not others? Chapter 4 established that while there is no statistically significant difference between the latitudinal distributions of CCF-hosting craters in the two hemispheres, notable regional variations exist. In particular, 56 northern hemisphere CCFs (25 of which were discussed in Section 5.4.2.2) are concentrated within Arcadia Planitia and Utopia Planitia, extending the northern CCF latitudinal range by 15°. Can the absence of a similar latitudinal range in the south, where CCFs are not found beyond ~50°S, be explained by this chapter’s interpretations and findings?

As shown in Table 5.8, a 10° poleward increase in δ corresponds to a 90 m decrease in d_a and so the poleward southern mid-latitudes should theoretically be capable of preserving CCFs. 2450 craters exist within the 50° to 65°S δ range and 3–40 km D range have the potential to host CCFs. However, the northern hemisphere distribution of CCFs between 50° to 65°N are restricted to elevations between -1.5 km and -5 km, potentially indicating the only zone where CCFs could form or remain preserved at the higher mid latitudes (see Section 5.4.2.2). If the same elevation constraint applies to the southern hemisphere, then only two regions, the southern halves of Argyre and Hellas Basins, meet this criterion containing just 121 and 124 craters within the CCF-appropriate diameter range, respectively.

Notably, CCFs are found in the northern half of Hellas Basin, which has fragmented LDM surface cover (Voelker et al., 2017). This patchy LDM coverage may have allowed for their identification as CCF due to visible and identifiable tracelines. In contrast, the southeastern side of Hellas, despite its high crater density (Tanaka and Leonard, 1995), is heavily mantled by a smoother, younger LDM (Voelker et al., 2017), which would obscure CCF identification. These deposits superimpose tracelines preventing classification as CCFs, and so some infilled craters may still contain ice but lack the necessary surface expressions for identification.

5.5. Conclusion

This chapter presents the first global analysis of the variability of CCF surface depths, providing a spatially extensive assessment of the factors controlling the degree of crater infilling. Measured from the host-crater rim to the CCF surface, d_a serves as a measurable proxy for the total volume of ice and underfilling of a crater accumulated or preserved within a crater. By quantifying these values across a global inventory, this study identifies large-scale spatial trends, allowing for a systematic evaluation of whether CCF surface positions are governed by initial formation conditions or subsequent modification.

I draw the following conclusions regarding the controls on CCF apparent depth:

1. Host crater diameter exerts the primary influence on CCF apparent depth

D exerts the primary influence on d_a , explaining 40.8% of the variance in the MLR model and corresponds to an increase in d_a of ~ 30 m per 1 km increase in D (Table 5.8). This reflects the fundamental role of D in defining host crater morphology, including d_i and the crater volumetric capacity (explored in Chapter 6), which together establishes the geometric accommodation space available for CCF infill. The magnitude of this host-crater signal is sufficiently large that, if not explicitly accounted for, it obscures weaker environmental gradients; only after isolating the D -signal through diameter binning do secondary latitudinal and elevational influences become detectable.

D alone, however, cannot capture host crater-scale heterogeneity, and such factors may further contribute to the remaining $\sim 50\%$ of unexplained d_a variance. Localised modification

of the host crater prior to palaeo-CCF formation—including variable degrees of infilling, degradation, or resurfacing—will alter both effective d_i and volumetric capacity independently of D . Rim degradation may also be important, as more eroded rims reduce topographically driven shading of the CCF surface, increasing insolation and promoting enhanced ablation relative to sharper, better-preserved rims. Rim condition may also influence the availability of debris from crater walls to form an insulative overburden, although the relative contributions of dust and debris to CCF mantling remain poorly constrained. When combined with broader spatial context—such as shadowing from surrounding topography—these crater-specific microclimatic factors plausibly exert additional control on d_a that is not captured by D , δ , or elevation alone, and likely account for a significant fraction of the residual variance.

2. Latitude exerts a secondary, systematic control on CCF apparent depth

δ exerts a statistically robust but secondary control on CCF apparent depth once the dominant effect of D is removed. Multivariate regression (Table 5.8) and D -binned analyses (Figure 5.18, 5.19; Table 5.4, 5.5) show that d_a decreases poleward in both hemispheres, with a 10° increase in δ corresponding to a reduction in d_a of approximately 90 m. Although this effect is weaker than the control exerted by D , it is consistent across hemispheres and persists after accounting for elevation, indicating that δ represents a genuine environmental gradient on CCF surface evolution.

I attribute this poleward decrease in d_a primarily to the monotonic decrease in mean annual insolation with latitude over low-to-intermediate obliquity states, a relationship that persists up to obliquities of $\sim 50\text{--}55^\circ$, beyond which the latitudinal insolation gradient reverses (Ward, 1974; Laskar et al., 2004; Madeleine et al., 2009; Figure 5.24). Reduced insolation at higher latitudes both favours greater net preservation during post-glacial periods by suppressing ablation and, potentially, favours thicker initial ice accumulation during glacial phases by shifting mass balance toward more positive values. Under an accumulation-dominated interpretation, higher-latitude CCFs should therefore tend to approach more rim-proximal surfaces. Alternatively, under an ablation-dominated interpretation, higher-latitude CCFs should equilibrate at smaller d_a values because long-term sublimation losses are reduced. The persistence of the observed latitudinal gradient under either framework indicates that latitude acts as a control on CCF surface position through its influence on surface energy balance rather than through hemispheric asymmetry or longitudinal forcing.

This latitudinal dependence is physically consistent with the monotonic decrease in mean annual insolation toward the poles for all obliquities relevant to Amazonian mid-latitude glaciation. Reduced insolation at higher latitudes implies reduced net ablation and/or more favourable long-term mass balance, allowing CCF surfaces to remain closer to the crater rim.

Crucially, this interpretation does not depend on whether present-day d_a primarily records initial accumulation geometry or subsequent post-glacial modification. Under either scenario, reduced insolation toward the poles favours the preservation of more rim-proximal CCF surfaces, establishing latitude as a first-order environmental control acting through surface energy balance.

3. Poleward latitude drives a convergence in apparent depth toward rim-full stability

In addition to shifting the mean value of d_a , latitude also exerts a strong control on its dispersion. As δ increases, the spread of d_a values narrows in both hemispheres, producing a wedge-shaped envelope in d_a – δ space in which low-latitude CCFs exhibit a wide range of fill states, while high-latitude CCFs cluster into a much narrower band. This convergence is expressed both as a progressive reduction in the range of d_a values poleward and as the near-disappearance of poorly infilled (large- d_a) CCFs at high latitudes (Figure 5.18, 5.19), aside from a small anomalous subset discussed separately.

This behaviour indicates a transition from a low-latitude regime in which CCF evolution is strongly modulated by local and regional controls to a high-latitude regime dominated by climatic admissibility and energy-balance constraints. In the north, GCM simulations by Madeleine et al. (2009) (Figure 5.25) indicate that high-latitude CCFs are restricted to a narrow and consistently favourable accumulation envelope, whereas lower-latitude CCFs persist across a much broader range of mass-balance conditions, naturally producing greater dispersion in d_a . Obliquity cycling and systematically lower atmospheric humidity at the highest latitudes likely prevent universal convergence to perfectly rim-full states, explaining the persistence of a small number of anomalously inset examples. Together, these effects account for both the wedge-shaped structure of the CCF population and the poleward convergence toward a restricted apparent-depth state.

At the highest northern latitudes ($>55^\circ$), a small subset of CCFs departs from this general behaviour by exhibiting slightly larger d_a values than expected despite occupying the most favourable preservation zone. The existence of this anomalous group indicates that additional processes, including obliquity-driven redistribution of ice and systematically lower atmospheric humidity, can locally counteract the otherwise strong tendency toward rim-full configurations at high latitude.

4. Elevation exerts an inverse control on apparent depth, attributed to pressure-dependent sublimation efficiency

The observed inter-hemispheric control of elevation on CCF d_a , where a 5 km increase in elevation corresponds to a ~ 40 m increase in d_a , suggests an inverse relationship compared

to terrestrial glacial systems. I tentatively hypothesise that the weak but statistically significant increase in d_a with elevation at a global scale is caused by elevation-dependent sublimation rates, where lower elevations experience higher surface pressures, reducing sublimation efficiency through lower $(\Delta p/p)$ values. Higher elevations, particularly in the southern hemisphere, sustain greater sublimation rates due to reduced surface pressure and enhanced wind-driven ablation, leading to less infilling and higher d_a values. This process contributes significantly to the hemispheric asymmetry in CCF preservation, accounting for ~40 m of the observed North-South offset.

Whilst 50% of the variance in apparent depth remains unaccounted for, this is likely influenced by factors beyond the scope of this chapter, such as localised microclimatic effects on mass balance, resurfacing happening before CCF occupation, uneven rim degradation leading to asymmetric shadowing within the crater, host crater resurfacing or local topographic barriers which may influence the amount of overburden available. Nonetheless, the environmental gradients, once separated from the effects of the host crater, provide quantifiable support for existing models, providing further insights into the histories of CCFs and climatic and geomorphological processes shaping Martian mid-latitude landscapes.

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Chapter 6: Estimating the Volume of CCFs

6.1. Introduction

This chapter estimates the potential volumes of the CCFs, both to establish their upper volume limits and to explore how such volume estimates can enhance our understanding of CCF glacial histories. Building upon the work in Chapter 5, this analysis combines the apparent depth (d_a) with host crater diameter (D), established host crater depth to diameter ($d_i:D$) ratios (e.g., Garvin et al., 2003; Chapter 5) and crater cavity scaling (Garvin et al., 2003) to estimate the theoretical maximum volume of each CCF, V_{CCF} . By finding d_t (Chapter 5) and assuming a rotationally-symmetric crater floor geometry, V_{CCF} is calculated as the volume between the theoretical crater floor (O) and the CCF surface elevation (Z_{floor}) (Figure 6.1). Each volume estimate is an upper bound on V_{CCF} , because in these calculations I take/assume hemispheric $d_i:D$ ratios derived for pristine craters, which do not account for any crater resurfacing or degradation/infill (which would reduce d_t) prior to CCF occupation.

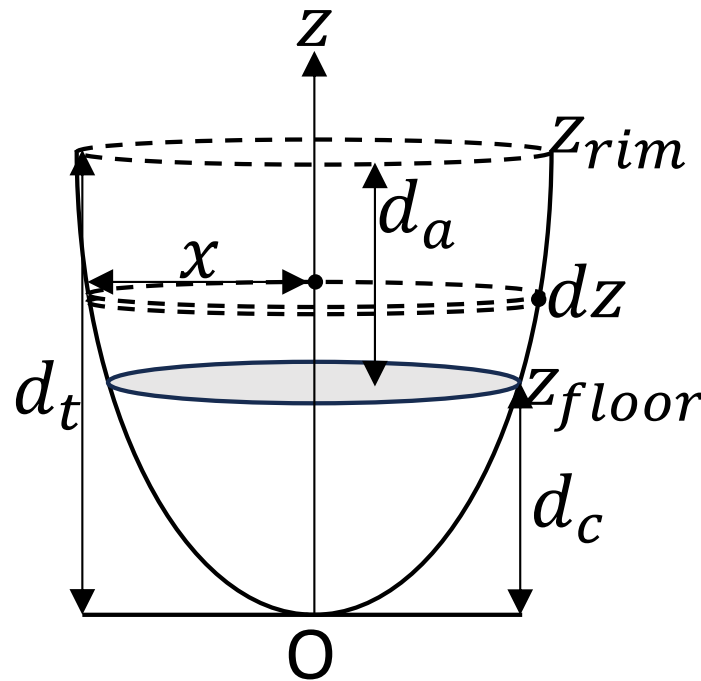


Figure 6.1. Schematic of idealised CCF hosting crater geometry, showing radial and depth parameters used in the disc-based volume integration. Symbols denote the theoretical depth (d_t), apparent depth (d_a), CCF depth d_c , rim and floor elevations (Z_{rim} and Z_{floor} respectively), and CCF surface profile. x is radial distance; dz is disc thickness; z is depth coordinate, and O is the depth origin.

The computed volumes not only constrain the potential ice volumes stored within CCFs but also provide a basis for interrogating their glacial histories. Chapter 5 showed that d_a increases systematically with D , raising the question how V_{CCF} varies with the host crater size across multiple CCFs (recall that a higher d_a means a CCF surface lying farther below the rim). One scenario is that V_{CCF} increases with D ,

perhaps linearly or following a power law. Another possibility, an end-member scenario, is that CCFs in different-sized craters have similar volumes, such that increases in D (and d_a) do not yield higher V_{CCF} . This scenario is consistent with a glacial history where the CCFs formed by an accumulation mechanism that is supply-limited, or where CCFs have a maximum-volume threshold controlled by mass-balance equilibrium under interglacial conditions, such that excess ice above the threshold would ablate regardless of host-crater capacity.

Accordingly, besides finding the total V_{CCF} of the CCF population, this chapter explores how individual V_{CCF} values scale with D and vary spatially. If V_{CCF} increases with D across hemispheric, regional, and local scales, then accumulation was not strongly supply-limited at those scales, implying that a sufficient moisture/precipitation source during the mid-latitude glacial phases to permit more infill in larger craters. Conversely, if CCFs of different D have similar maximum volumes, this would indicate some kind of “volumetric capping”—either due to limited accumulation or evolution toward equilibrium volumes. In this context, although the scaling between V_{CCF} and D might not uniquely resolve CCF origin scenarios (discussed in Section 2.4.1), it helps to constrain their plausibility.

The scenarios above link with broader concepts about the style of mid-latitude glaciation. The widespread distribution of CCFs (Levy et al., 2014; Chapter 4) argues against formation beneath a single hemispheric ice sheet, which would require Amazonian-era ice volumes far exceeding available Amazonian GEL estimates (Fastook and Head, 2014; Section 2.1.2 & 2.4.1.). However, it remains plausible that many of the ancestral ice masses of the CCFs were ice caps with vertical and horizontal extent far exceeding the craters—more than filling their cavities, with surfaces lying above their rims—before receding/shrinking under interglacial ablation to the current configuration. Accordingly, quantifying V_{CCF} and its scaling with D might (also) help us discern whether the CCF morphometries today reflect a heritage from past regional ice caps or past ice accumulation confined to craters.

To date, only the study of Levy et al. (2014) has produced a global volume estimate for CCFs: $\sim 8.75 \times 10^4 \text{ km}^3$, translating to a GEL of $\sim 0.6 \text{ m}$. However, their inventory includes both ‘low-definition’ CCFs (Levy et al., 2010; Chapter 3) and features now classified as pole-facing flow features (e.g., Dickson et al., 2012; Sinha and Vijayan, 2017), thus likely inflating the total volume. In contrast, in this chapter I examine the subset of high-confidence CCFs that meet the strict morphological criteria defined in Chapter 3—those displaying clear crater-filling geometry and viscous flow lineations indicative of ice-rich deformation. My figure for the total volume pertains only to these CCFs; hence it is a *minimum* estimate for the total amount of ice in CCFs globally.

Comparable volume estimates exist for other VFF populations, notably for 281 LVF features (totally $6.51 \times 10^4 \text{ km}^3$; Levy et al., 2014), 1428 LDA features (totalling $1.55 \times 10^5 \text{ km}^3$; Karlsson et al., 2015), and 1243 GLFs (with volumes ranging between $523 \pm 132 \text{ km}^3$ and $1570 \pm 397 \text{ km}^3$, based on assumed 30% and 90% ice concentrations; Brough et al., 2019). Each estimate carries uncertainties in ice content,

purity, and surface mantling. My findings in this chapter will reveal how the high-confidence CCFs compare volumetrically against these other VFF types, as a contribution towards the mid-latitude ice inventory. I will also use the V_{CCF} estimates of these CCFs as a controlled dataset for investigating whether CCF volumes encode broader climatic or glacial gradients across Mars.

Several questions guide the work in this chapter:

1. What are the total volume and GEL of CCFs?
2. What proportion of the host crater capacity is (typically) filled by a CCF?
3. How do CCF volumes vary by crater diameter, latitude, elevation, and across different regions?
4. Do CCFs within a given region have similar ice volumes? On isovolumetric frameworks (Sec. 6.5), do CCF volumes show systematic patterns that indicate a threshold for preservation?
5. To what extent can volumetric thresholds be used to explore and reconstruct the accumulation and preservation history of CCFs?

In Section 6.2, my method of estimating the volume of a CCF is first described. The exposition then unfolds along three lines, to report the findings on the V_{CCF} – D scaling and “total V_{CCF} ” (Section 6.3); to examine the spatial variation of V_{CCF} , e.g., how it varies between hemispheres and with latitude and elevation, and whether any trends mirror the ones found for d_a (Section 6.4); and, finally, the formulation of an “isovolumetric framework” to assess CCF volume variability, focusing on the minimum volumes necessary for preservation and on regional volumetric divergence from constant-volume trends (Section 6.5).

6.2. Datasets and methods

For a CCF and host crater with known surface geometries, at what elevation is the CCF’s base (i.e. the obscured crater floor)? We need the surface-to-base elevation difference to find V_{CCF} ; the basal contact is a 3D surface, typically concave downwards. Because radar data (e.g. MARSIS/SHARAD) struggle to resolve steep crater walls, profiles of the base is scarce/non-existent for CCFs, and so for the majority population of CCFs the essential problem is one of *inferring* their bed elevation.

Methods for estimating the volume of terrestrial ice masses fall into two groups. One group, which is physically-based, retrieves the ice thickness from observable surface variables (notably surface slope and/or surface velocities) by incorporating processes such as mass conservation and ice-flow dynamics (e.g., Fastook et al., 1995; Farinotti et al., 2009; Farinotti et al., 2013; Huss and Farinotti, 2014), or by using a plastic yield stress to parameterise the glacier force balance (Li et al. 2012). These methods are not applicable to CCFs, because (i) CCFs flow undetectably slowly (if at all) so no surface velocity observations presently exist, (ii) CCF surface slopes are exceedingly shallow (Chapter 7) and no study

has tried to use their slopes and thickness to ascertain a yield stress (though this has been done for LDAs; Karlsson et al., 2015), and (iii) we have no mass-balance records for CCFs. Another group of methods uses empirically-based scaling relations between area and volume (e.g., Bahr et al., 1997; Bahr et al., 2015) to estimate volume from surface area, bypassing the step of finding the bed elevation or ice thickness. However, these relations are known only for terrestrial ice masses and not usable for CCFs. (In fact, my results in this chapter can be used to estimate a CCF area-volume scaling for Mars for the first time.)

Here, I estimate CCF volume from D and the d_a by using a geometric approach that idealises the host crater's shape and that draws on published analyses of impact crater morphometry on Mars. Focussing on pristine, unmodified craters, some studies used DEMs (e.g., from MOLA) together with visual images to reconstruct the pre-impact surfaces by Delaunay-triangulation based interpolation across the crater cavities, thus estimating the cavity volumes and uplift and ejecta volumes (Stewart and Valiant, 2006; Black and Stewart, 2008). The corresponding volume– D relationships do not aid finding V_{CCF} , because most CCF surfaces lie below rim (Chapter 5). Fortunately, other studies have used MOLA DEM data to derive D -dependent shape functions to represent the crater cavity surface of pristine craters (e.g. Garvin et al., 2003; Section 6.2.1). I use these functions in my method of finding V_{CCF} from D and d_a . I use them also to find the host crater (cavity) volume V_{host} from D .

The resulting V_{CCF} and V_{host} will be upper-bound estimates for the true (ice and cavity) volumes because the use of shape functions for pristine craters neglects (i) resurfacing or infilling of the host crater before CCF formation/occupation and (ii) the potential presence of a central peak (in large/complex craters). V_{CCF} is an upper bound also because (iii) surficial debris may form part of the computed volume and (iv) the CCF may not have a pure ice composition. Deviation from idealised crater geometries is also possible. The associated volumetric uncertainties will be assessed (Sec. 6.2.2) after I detail the method.

6.2.1. Formulation of the method

Figure 6.1 shows the geometry and variables of interest. The core idea is to take the volume of a known crater/cavity shape function and calculate the volume up to the CCF surface (shaded grey) by integration. The height between the lowest point and the CCF surface—the CCF depth, d_c —is not known from measurement (not observable without radar imaging), but I calculate it via

$$d_c = d_t - d_a, \quad (6.1)$$

where d_t is the host crater depth obtained in Chapter 5 (considered further below).

In their study, Garvin et al. (2003) parameterised pristine craters by using the cavity shape function (Figure 6.1)

$$z = kx^n, \quad (6.2)$$

where x represents radial distance from the crater centre, z is height above the lowest point, k is the cavity shape coefficient, and n is the cavity shape exponent. A perfectly conical crater has $n = 1$, a perfectly paraboloid crater has $n = 2$, and $n > 2$ indicates more U-shaped craters with steeper walls and flatter floors. Simple craters tend to approximate conical or parabolic geometries and have $1 \lesssim n \lesssim 2$ (Grieve and Garvin, 1984), whereas complex craters often have $n > 2$ (Melosh and Ivanov, 1999; Sturm et al., 2016).

Garvin et al. (2003) derived different sets of k and n values for simple ($D < 7$ km) and complex ($D \geq 7$ km) craters. In fact, these authors presented two versions of these morphological-class based parameter sets—one version assuming fixed values of k and n in each crater class, the other version with k and n both varying with crater diameter D in each crater class. I refer to these as ‘fixed’ and ‘ D -dependent’ shape parameterisations, respectively. Accordingly, when calculating V_{CCF} and V_{host} with my method, besides using a cavity shape function with parameters corresponding to the morphological class of the crater in question, one can adopt either type of parameterisation. As explained later, I choose the D -dependent parameterisation throughout the work of this chapter. More recently, Watters et al. (2015) also assessed pristine crater geometry on Mars, but their shape parameters differ minimally from those of Garvin et al. (2003). I therefore proceed with using the Garvin et al. (2003) parameters.

Turning to the integration, the volume between heights a and b is given by

$$V = \int_a^b \pi [f(z)]^2 dz, \quad (6.3)$$

where the inverse function $x(z) = (z/k)^{1/n}$ gives the disc radius at height z . Consequently, the volume estimate for the CCF is

$$V_{CCF} = \pi \left[\frac{n}{(n+2)} \right] k^{\frac{-2}{n}} d_c^{\frac{(n+2)}{n}}. \quad (6.4)$$

Similarly, the host crater depth can be found by using

$$V_{host} = \pi \left[\frac{n}{(n+2)} \right] k^{\frac{-2}{n}} d_t^{\frac{(n+2)}{n}}. \quad (6.5)$$

From these volumes, the fraction fill of the host crater occupied by the CCF is calculated as $V_{CCF}/V_{host} = (d_c/d_t)^{(n+2)/n}$, which provides a normalised measure of infilling.

The input data for the method are readily available from Chapters 4 and 5, where, for each CCF with known geographical position (geometric centroid coordinates), the host crater diameter D and rim elevation Z_{rim} were obtained from the Robbins and Hynek (2012) global crater database, and the CCF’s surface elevation Z_{CCF} evaluated as the mean MOLA elevation within its digitised outline in ArcGIS Pro. As in Chapter 5, we have $d_a = Z_{rim} - Z_{CCF}$ (Fig. 6.1).

The host crater depth d_t is needed in Eqns. (6.1) and (6.5). I calculate it from D by using my $d_t:D$ ratios listed in Section 5.3.1 (Table 5.2), which distinguish simple and complex craters with the same D threshold (7 km) used by Garvin et al. (2003) for k and n . These ratios were computed specifically for this Thesis from pristine mid-latitude craters in the Robbins and Hynes (2012) catalogue and were derived separately for the Northern and Southern hemispheres to reflect known hemispheric variation in pristine crater depths (Chapter 5). The alternative approach—using the Garvin et al. (2003) cavity-shape function in Eqn. (6.2), i.e., $d_t = k(D/2)^n$ —is not adopted here because the Garvin coefficients are obtained from global fits and therefore impose a single $d_t:D$ ratio that cannot represent hemispheric structure. Thus, my hemispheric $d_t:D$ ratios developed in Chapter 5 therefore provide the appropriate basis for obtaining d_t .

The specific crater parameters of Garvin et al. (2003) are as follows: in the fixed parameterisation, $k = 0.204$ and $n = 1.76$ for simple craters, and $k = 0.008$ and $n = 2.65$ for complex craters; and in the D -dependent parametrisation, $k = 0.76D^{-1.17}$ and $n = 1.04D^{0.366}$ for simple craters ($D < 7$ km), and $k = 5.62D^{-2.51}$ and $n = 1.62D^{0.14}$ for complex craters ($D > 7$ km). Further experimentation shows that the cumulative V_{CCF} difference between these parameterisation schemes was 3533 km³ across my CCF sample, corresponding to only 6.3% of the total V_{CCF} . I choose the D -dependent scheme for all subsequent volume calculations, whilst no regression fits (R^2) are stated by Garvin et al. (2003), this approach better captures the transition toward a more U-shaped cavity as d_t scales with D , as discussed in Chapter 5.

Previous global CCF volume estimates produced by Levy et al. (2014) employed the same Garvin et al. (2003) geometric formulation and volume integration detailed above, but differed in how the required inputs were obtained. In the approach of Levy et al. (2014), d_t was calculated directly from the Garvin et al. (2003) $d_t:D$ ratios, Z_{rim} was estimated as the maximum MOLA elevation within a fixed 2 km buffer around the mapped CCF outline, and Z_{CCF} was taken as the mean MOLA elevation within their CCF outline. Thus, they would obtain different values of d_a , d_t and d_c . Their study also adopted the fixed Garvin et al. (2003) parameterisation for k and n , rather than the D -dependent form used here. Consequently, differences between their V_{CCF} estimates and those presented in this chapter are expected, despite the shared mathematical framework.

6.2.2. Assumptions and uncertainty

As explained before, V_{CCF} and V_{host} found by the method are upper bounds. Separately, the use of parameterisations to find d_t and for the cavity shape do not capture variations of real crater geometry from the best-fit. Assessing the uncertainty on the volume results is therefore necessary.

Uncertainties arise from both instrument and measurement error in crater geometry and systematic differences between cavity shape coefficients (n and k). Because V_{CCF} is derived from the reconstructed host crater, uncertainty in D propagates through both V_{host} and V_{CCF} . Since the cavity shape coefficients

k , n and d_t are D -dependent, any uncertainty in D influences the reconstructed cavity shape and therefore the integrated volume of both the host and the fill.

As discussed in Section 5.2.4, the number of DEM pixels within a crater reflects the raster resolution of the gridded product rather than the density of original MOLA shot points used to constrain the surface. The pixel-based filtering applied here therefore ensures that each crater is represented by sufficient gridded DEM cells for volume reconstruction, but it does not directly measure the underlying MOLA point density. Local interpolation between MOLA tracks may influence rim, floor, and CCF surface-elevation estimates, and therefore propagate into d_a , d_c , V_{CCF} and V_{host} , particularly for smaller craters. This limitation is acknowledged qualitatively here but is not independently propagated through the volume uncertainty calculations below.

The uncertainties in D reported by Robbins and Hynek (2012) are approximately ± 0.034 km (1.3%) for craters of 3 km and ± 0.12 km (0.3%) for craters of 40 km (Chapter 5). To quantify their impact, volumes were recalculated using the upper and lower bounds of $D \pm \delta D$, and the propagated volume uncertainty was taken as half the difference,

$$\delta V_{horz} = \frac{V_{(D+\delta D)} - V_{(D-\delta D)}}{2} . \quad (6.6)$$

Besides this error source, I estimate the potential uncertainty associated with the choice of crater-shape parameterisation (i.e. a model-driven systematic error), by calculating the difference in V_{host} between the fixed and D -dependent parameterisation schemes:

$$\delta V_{coeff} = V_{fixed} - V_{diameter-dependent} . \quad (6.7)$$

The combined uncertainty in V_{host} , assuming the errors to be independent, is

$$\delta V_{host} = \sqrt{\delta V_{horz}^2 + \delta V_{coeff}^2} . \quad (6.8)$$

Uncertainty in V_{CCF} arises from three main sources: horizontal and vertical variables in the geometry shown in Figure 6.1, and systematic components. The horizontal and systematic components propagate from uncertainty in crater D (as described above), and the vertical component reflects uncertainties in the measurement of d_a derived from Z_{rim} and Z_{CCF} . Instrumental uncertainty in MOLA data is estimated at ± 25 m, giving a combined vertical measurement error of ± 50 m (Robbins and Hynek, 2012; Chapter 5). In addition, the standard deviation of the Z_{rim} and Z_{CCF} measurements is included to account for variability in surface topography. The total uncertainty in d_a (in metres) is estimated as

$$\delta d_a = \sqrt{0.05^2 + \sigma_{rim}^2 + \sigma_{floor}^2} . \quad (6.9)$$

This uncertainty propagates through to d_c and ultimately to V_{CCF} via the integral in Eqn. (6.4). The resulting vertical component of uncertainty in V_{CCF} is estimated to be

$$\Delta V_{vert} = \frac{V(da+\delta d_a) - V(da-\delta d_a)}{2} \quad (6.10)$$

The total estimated uncertainty in V_{CCF} is then the root-sum-square of all three components,

$$\delta V_{CCF} = \sqrt{\Delta V_{vert}^2 + \Delta V_{horz}^2 + \Delta V_{coeff}^2}. \quad (6.11)$$

6.3. Total CCF volumes

This section first quantifies the total V_{CCF} and equivalent GEL across this Thesis' global CCF sample (Chapter 3) and situates these results within broader volatile inventory (Section 6.3.1). The analysis then evaluates how V_{CCF} scales with D (Section 6.3.2), beginning with normalisation as fill fraction to assess the robustness of preservation patterns under alternative parameterisations (Section 6.3.2.1) before considering absolute volumes to examine hemispheric contrasts in scaling with D (Section 6.3.2.2).

This two-step approach provides a critical test of whether V_{CCF} is simply a function of D , or whether additional climatic and preservation controls must be invoked to explain the observed patterns, which are explored further in the subsequent analysis (Sections 6.4 and 6.5).

The method was applied to every CCF in my CCF sample (1225 in the North; 1268 in the South). This section presents the cumulative V_{CCF} and its corresponding GEL, situating these results in the mid-latitude water ice inventory (Section 6.3.1). It also analyses how V_{CCF} scales with D (Section 6.3.2) and uses the normalised fill fraction to assess consistency of crater infilling across different crater sizes under alternative parameterisations (Section 6.3.2.1). This ensures that any identified variation of V_{CCF} with D reflects real physical trends rather than artifacts of the specific crater-shape parameterisation. Then, I examine hemispheric contrasts in scaling of V_{CCF} with D (Section 6.3.2.2). These studies ascertain whether V_{CCF} is solely a function of D , or whether climatic and preservation factors must be invoked to explain its variations. The findings pave the way for the work in Sections 6.4 and 6.5.

6.3.1. CCF total volume

The totalled V_{CCF} across the global dataset amounts to $5.6 \pm 0.47 \times 10^4 \text{ km}^3$, corresponding to $0.12 \pm 0.01 \text{ m GEL}$ and $0.35 \pm 0.03 \text{ m GEL}$, assuming ice purities of 30% and 90% respectively. Of this total V_{CCF} , ~64% is contained within the northern hemisphere, highlighting a pronounced hemispheric asymmetry. This disparity reflects both the systematically shallower crater morphometries (lower d_i) and d_a of southern CCFs (Chapter 5), the latter likely indicating more extensive post-glacial ablation. Consequently, the reduced southern volumes likely result from a combination of geometric capacity differences and enhanced ablation or limited accumulation during glacial phases. These total V_{CCF}

estimates are lower than the only previous global CCF volume estimate, in which 9533 CCFs have a total volume of $8.75 \times 10^4 \text{ km}^3$ (Levy et al., 2014). Notably, the present value represents $\sim 64\%$ of that estimate, despite being derived from a sample amounting to only 26.1% of the previously mapped inventory.

The total V_{CCF} represents a small fraction of the present-day H_2O budget on Mars, whose exchangeable near-surface reservoir is estimated at $\sim 20\text{--}30 \text{ m GEL}$ (Carr and Head, 2015; 2019; Jakosky and Hallis, 2024). Table 6.1 shows total V_{CCF} alongside the published volume estimates for GLFs (Brough et al., 2019), LDAs (Karlsson et al., 2015), and LVFs (Levy et al., 2014). The total V_{CCF} is modest relative to other VFF classes, indicating that CCFs form a significant but non-dominant component of the mid-latitude ice budget, with volumes comparable to LVFs but well below those of LDAs. Combining the total V_{CCF} with those estimates (shown in bold in Table 6.1), the total mid-latitude VFF ice storage comes to $0.61\text{--}1.82 \text{ m GEL}$ at $30\text{--}90\%$ ice purity. This range is tighter than the ~ 0.9 to 2.6 m GEL inferred by Levy et al. (2014) for CCFs, LDAs and LVFs because, since their syntheses, LDA volumes have been down revised by Karlsson et al. (2015), and now, the high-confidence CCF volumes revised by me. This is despite the inclusion of GLF volumes not accounted for in the Levy et al. (2014) inventory.

Table 6.1. Volume and GEL estimates for mid-latitude VFFs assuming ice purities of 30% and 90%. Values are compiled from previous studies and this chapter.

Landform and count	Estimated Ice Volume (km ³)	GEL (m) ^a	Ice purity (%)	Source	
CCF	2493	5.6 ±0.47	0.39±0.033	100	This chapter
	2493	1.68 ±0.14 x 10 ⁴	0.12 ±0.01	30	This chapter
	2493	5.04 ±0.43 x 10 ⁴	0.35 ±0.03	90	This chapter
	9533	2.62× 10 ⁴	0.18	30	Levy et al., 2014 ^b
	9533	7.87× 10 ⁴	0.55	90	Levy et al., 2014 ^b
GLF	1243	5.23 ± .13 x 10 ³	0.04±0.009	30	Brough et al., 2019
	1243	1.57 ± .4 x 10 ⁴	0.11	90	Brough et al., 2019
LVF	281	1.95 × 10 ⁴	0.14	30	Levy et al., 2014 ^b
	281	5.86 × 10 ⁴	0.41	90	Levy et al., 2014 ^b
LDA	1428	7.89 × 10 ⁴	0.55	30	Levy et al., 2014 ^b
	1428	2.37 × 10 ⁵	1.6	90	Levy et al., 2014 ^b
	1428	0.45× 10 ⁵	0.31	30	Karlsson et al., 2015
	1428	1.4× 10 ⁵	0.95	90	Karlsson et al., 2015

Total GEL of Mid latitude VFFs: 0.61 to 1.82 m

^a 1 m GEL is equivalent to 100 g-cm^{-2} (Jakosky, 2021)

^b ice purities calculated from total volume given in paper

Importantly, being restricted to my CCF sample, my total V_{CCF} figure overlooks a range of other crater-confined, ice-rich deposits (Table 6.2), whose volume remains unquantified at the global scale. These include deposits previously grouped within the broader CCF population, such as low-definition CCFs and pole-facing flow features, as well as crater-fill units that fall outside my CCF classification (Chapter 3) whose ice content remains possible but unknown. Among the most likely glacial candidates are equatorial craters containing “CCF-like” infill, interpreted as ice-bearing material preserved beneath thick debris mantles (Shean et al., 2010). The volumetric approaches applied here, and by Levy et al. (2014), can be extended to such examples, provided appropriate adjustments are made for lower ice purity. Pole-facing flow features and crater-hosted LDAs (e.g., Dickson et al., 2012; Hartmann et al., 2014; Sinha and Vijayan, 2017) also represent plausible ice reservoirs, though their high surface inclination means that my method (assuming filling of cavity up to a fixed level) cannot be used to estimate their volume. More uncertain are heavily mantled or otherwise featureless crater fills at mid-latitudes, whose ice content remains unknown. Although not tackled here, a global survey can be made to assess whether supplementary VFF morphologies occur on their surfaces.

Table 6.2. Candidate populations of potentially ice-bearing crater-confined deposits not included in current global inventories.

Candidate Population	Description
Low Definition CCF	Craters containing deposits which possess morphologies reminiscent of CCFs but lacking diagnostic surface textures such as tracelines or brain terrain. These features, often interpreted as highly degraded concentric fills or debris-covered lobate floor deposits (Head et al., 2008; Levy et al., 2010), may still preserve ice beneath mantling or debris cover.
Low latitude /equatorial CCFs?	Low-latitude craters host material morphologically similar to CCFs, generally interpreted as debris-covered ice occupying crater floors (Shean et al., 2010). Owing to their equatorial setting, however, any surviving ice is likely to occur at lower purity, requiring burial beneath a relatively thick, insulating debris layer (Shean, 2010).
Mid latitude infilled craters	Thousands of buried and partially infilled craters occur across the mid-latitudes, but lack the morphologies required for classification as CCFs (Chapter 3) or as low-definition CCFs (Levy et al., 2010). Nevertheless, such craters may still retain ice beneath mantling deposits or debris cover, rendering any CCF-like material presently undetectable.
Flow Features (pole facing) and Crater Confined LDA	Lobate shaped flow features similar to small-scale LDAs inside craters, which show flow characteristic thought to be different from CCF (e.g., Dickson et al., 2012; Sinha and Vijayan, 2017)

6.3.1.1 Volumetric Comparison with Levy et al. (2014)

Here I compare the V_{CCF} estimates derived in this Thesis with those reported by Levy et al. (2014) for the same 2491 high-confidence CCFs. As outlined in Section 6.2.1, both studies employ the same formulation based on the Garvin et al. (2003) cavity shape functions and volume integration, but differ in how the required geometric inputs are obtained. Differences in the choice of k and n parameterisation are expected to have only a minor effect: adopting fixed versus D -dependent parameters changes the cumulative V_{CCF} by only $\sim 6.3\%$. By contrast, systematic differences in d_t and in the determination of Z_{rim} are expected to exert a much stronger control on d_c and therefore on V_{CCF} .

The hemispheric $d_t:D$ ratios used in this Thesis are systematically larger than the global scaling of Garvin et al. (2003), in the northern hemisphere d_t is ~ 200 m deeper across most D , and in the southern hemisphere d_t is ~ 100 m deeper up to $D \approx 25$ km (Figure 5.11). This increase in d_t is expected to increase d_c and therefore increase V_{CCF} relative to the values obtained by Levy et al. (2014). A second, and potentially larger, source of divergence arises from the estimation of Z_{rim} . Whereas I obtained the Z_{rim} values from the Robbins and Hynes (2012) crater database (values validated and discussed in Section 5.2.2), Levy et al. (2014) determined Z_{rim} as the maximum elevation within a 2 km buffer around each mapped CCF outline. In large craters, the buffer can fall short of the crater rim (Figure 6.2A), leading to underestimation of Z_{rim} and d_a and overestimation of d_c and V_{CCF} . In small craters, the buffer commonly extends beyond the rim and may intersect adjacent high-standing terrain or neighbouring crater rims (Figure 6.2B), increasing Z_{rim} and d_a , which in turn reduces d_c and V_{CCF} . Therefore, it is expected that Levy et al. (2014)'s volumes will be higher in large craters and lower in small craters relative to my volume estimates, these expectations can be tested directly by comparing the two volume inventories on a crater-by-crater basis.

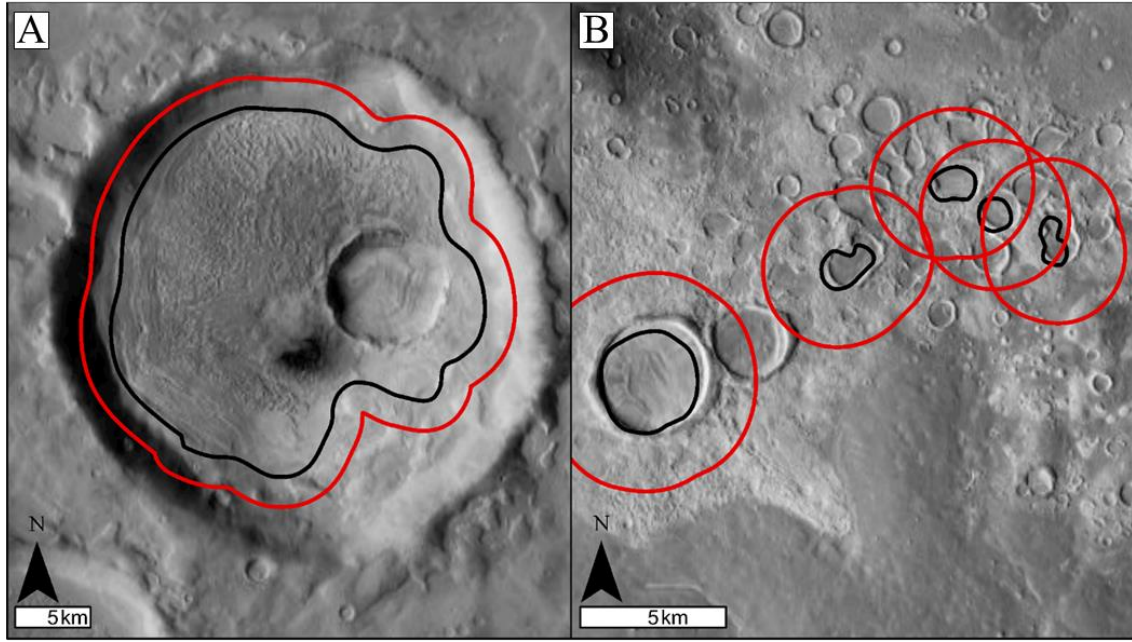


Figure 6.2. Effects of fixed-radius buffer placement on crater-rim elevation estimates. Levy CCF outlines are shown in black and the 2 km buffer outlines used to estimate rim elevation are shown in red. (A) Example of a large crater in which the 2 km buffer fails to reach the true crater rim (35.54° N, 71.48° E; CTX images: B19_017063_2155 and B19_016852_2156). (B) Example of smaller craters in which the buffer extends beyond the crater rim and overlaps adjacent high-standing terrain (75.96° N 14.82° E; CTX image B18_016799_2256).

Figure 6.3A plots the V_{CCF} estimates of Levy et al. (2014) (hereafter V_{Levy}) against my own volumes estimates (hereafter V_{BC}) for the 2491 shared high confidence CCFs revealing significant disagreement at the individual crater scale. To isolate the source of this discrepancy, Figure 6.3B plots the residuals ($V_{Levy} - V_{BC}$) against D . If the buffer-based rim method were the primary driver of error, positive residuals would be concentrated in large craters and negative residuals in small ones. Contrary to this expectation, the residuals show the opposite trend: for craters with $D < 7$ km, nearly 30% (402/1404) have positive residuals, whereas only 13% (140/1087) of craters with $D \geq 7$ km exceed my estimates.

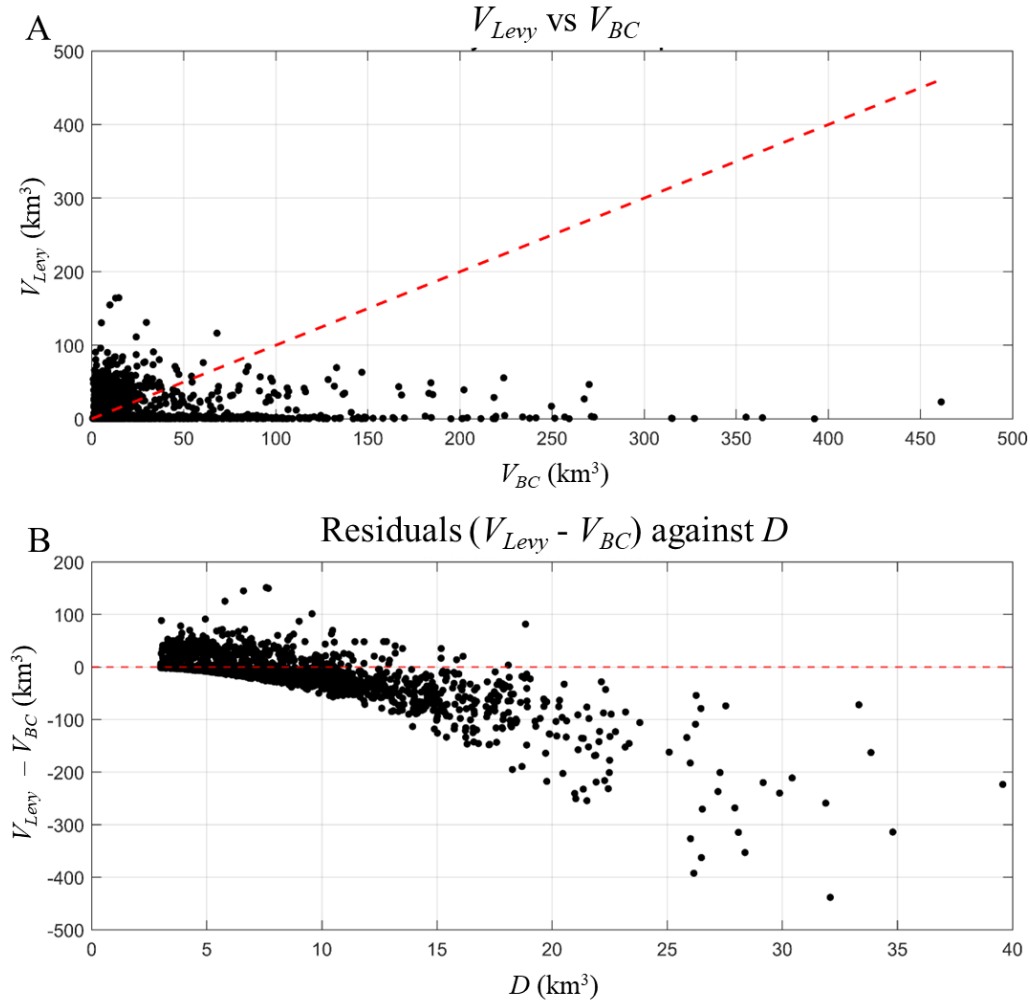


Figure 6.3. Comparison between Levy et al. (2014) and this chapter's CCF volume estimates.

(A) Scatter plot comparing CCF volumes reported by Levy et al. (2014) (V_{Levy}) with the volumes derived in this study (V_{BC}) for the 2491 high-confidence CCFs common to both datasets. The red dashed line indicates the 1:1 correspondence. (B) Residuals between the two estimates ($V_{Levy} - V_{BC}$) plotted as a function of crater diameter D . The red dashed line marks zero residual.

In practice, the comparison reveals two distinct trends. For the majority of the sample, Levy et al. (2014) report lower volumes than this study (71% of small craters and 87% of large craters). In large craters, this may indicate that the deeper $d_i:D$ ratios established in Chapter 5 exert a stronger influence on V_{CCF} than the rim-estimation bias. However, a significant contradiction appears in the remaining population: despite a methodological bias that should suppress V_{CCF} in small craters, nearly 30% of their small-crater estimates ($D < 7$ km) actually exceed the values derived here. Crucially, as shown by the vertical scatter in Figure 6.3A, many of these exceedances involve anomalously large volume values that appear physically implausible for craters of such small D . The magnitude of these residuals—often exceeding 50 km³—points to a fundamental internal inconsistency in the Levy et al. (2014) methodology that overrides geometric expectations.

The same issue is even more clearly revealed when considering the full Levy et al. (2014) CCF inventory. Figure 6.4 plot the volume estimates of Levy et al. (2014) for CCFs in simple and complex craters (points) and the theoretical volume of pristine, unfilled craters calculated using the Garvin et al. (2003) d/D power laws (i.e., $d_a = d_i$) (curves); the latter constrains the maximum possible CCF volume at each crater size D . For 2913 of the 7475 CCFs in Levy et al. (2014)'s inventory hosted by simple craters, their volume estimates exceed the theoretical maximum, some by several orders of magnitude (Figure 6.4A), running counter to the directional biases anticipated from their rim-elevation method. Since the curves portraying the Garvin-based power laws themselves enter Levy et al. (2014)'s method of estimating CCF volume, these exceedances indicate internal inconsistencies between d_a and d_i in their calculations. The associated overestimation of V_{CCF} from small craters by Levy et al. (2014) contributes towards why their total CCF volume in Table 6.1 exceeds my total V_{CCF} , even though a more obvious reason is the small number of CCFs in my sample. To showcase such overestimation with an example: for one CCF in a crater with $D = 6.6$ km (Appendix A), I estimated $V_{CCF} = 9.9$ km³ with my method, whereas Levy et al. (2014) report 154 km³; their volume seems implausible because a circular cylinder of fill at diameter D would need to be 4.5 km deep to explain it.

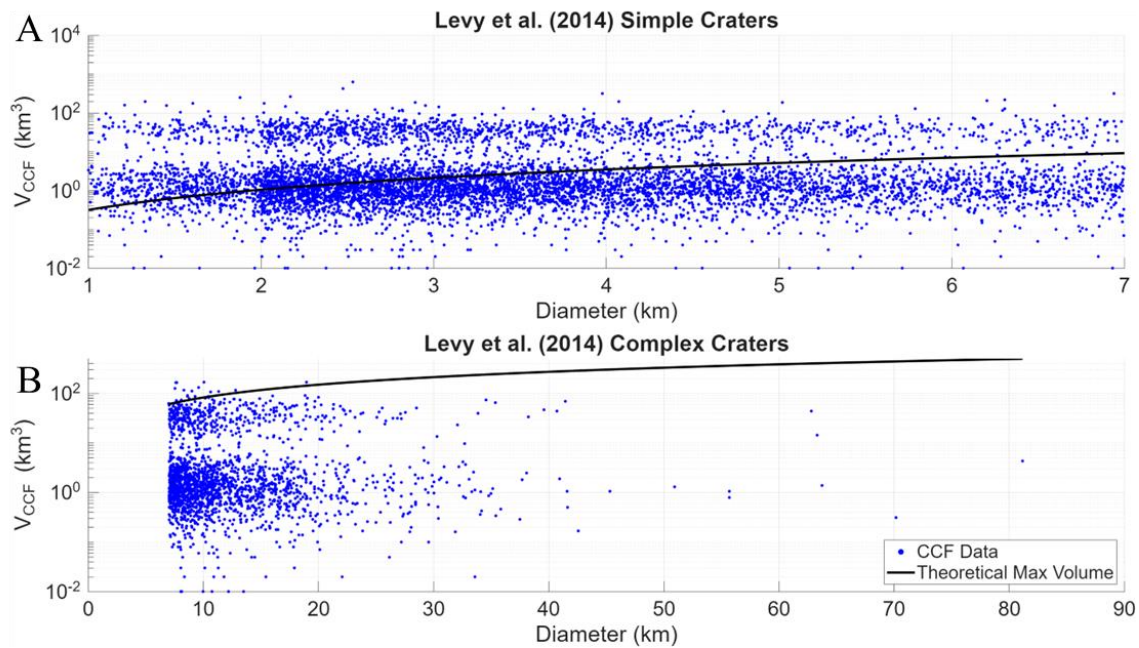


Figure 6.4. CCF volume estimates of Levy et al. (2014) versus host crater diameter. (A) Simple craters ($D < 7$ km), (B) complex craters ($D > 7$ km). Blue points plots their V_{CCF} values for 9,533 CCFs; black lines plot the theoretical volume of pristine, unfilled craters found with Garvin et al.'s (2003) d_i/D power laws.

6.3.2. CCF volume and host crater diameter

Imagine two CCFs at the same latitudes and elevations but differing in D by 10 km (Figure 6.5). At this crater diameter difference, the multilinear regression model in Chapter 5 (Section 5.3.6) predicts a difference in d_a of ~ 300 m. Because $d_a = d_i - d_c$ (Eqn. 6.1), this influence by D on d_a includes the

influence by D on d_t (crater depth), i.e., the d_t – D scaling (e.g., Garvin et al., 2003; Chapter 5). An initial expectation follows directly from crater morphometry. Larger craters possess deeper and more capacious cavities, so V_{CCF} is expected to increase with D even if d_a decreases, because d_t increases more rapidly than the diameter-related increase in d_a . If the two craters in Figure 6.5 were located in the northern hemisphere, a 10 km increase in D would increase d_a by ~ 300 m, yet increase d_t by 310 m. Although this difference is small, the resulting increase in d_c ($d_t - d_a$) with D still outpaces the increase in d_a , yielding a greater V_{CCF} . Larger craters also tend to develop more U-shaped cavity geometries with steeper walls and flatter floors as D increases, which further increases their volumetric capacity. Because the regression model (Section 5.3.6) accounts for $\sim 50\%$ of the variance in d_a , additional volume–depth configurations must be considered. *If* instead the CCFs have identical thicknesses d_c , then the 300-m difference in d_a solely reflects the geometric (scaling) effect of crater size on crater (cavity) depth; note, in this case, that the CCF in the larger crater still retains greater V_{CCF} than the CCF in the small crater. *If* instead the two CCFs have identical V_{CCF} , then the CCF in the larger crater is thinner, with smaller d_c , so the d_a – D relationship encapsulates an added signal from “equal V_{CCF} ”, besides the influence by D on d_t .

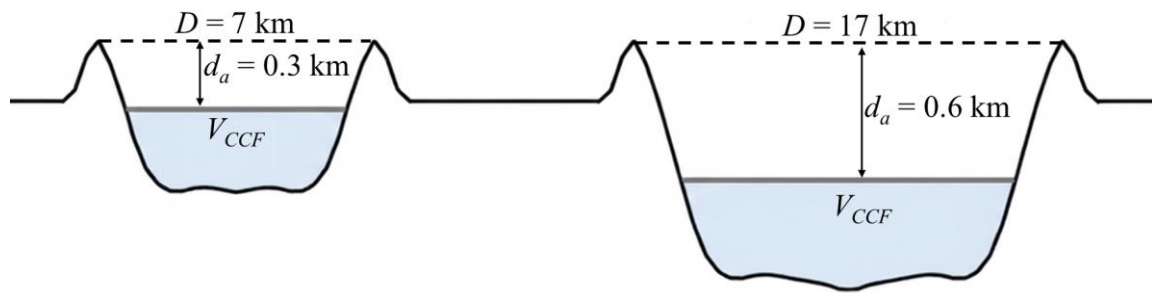


Figure 6.5. Schematic comparison of d_a and V_{CCF} for adjacent craters with differing D . The schematic illustrates the systematic increase of d_a with D , established in Chapter 5, while leaving the corresponding behaviour of V_{CCF} unconstrained.

Thus, exploring how V_{CCF} varies with D and with host-crater capacity motivates the analyses that follow. Section 6.3.2.1 examines fill fraction (V_{CCF} / V_{host}) to determine how frequently CCFs approach rim-full states and whether hemispheric contrasts exist; this metric is also used in subsequent sections because it normalises V_{CCF} by each crater’s capacity, allowing differences in volumetric states to be compared directly across craters of different sizes. This step establishes the extent to which CCFs utilise their available cavity volume. Section 6.3.2.2 then evaluates how V_{CCF} and fill fraction are distributed within discrete D bins in each hemisphere, allowing the upper and lower volumetric limits within each D bin to be assessed. These bin-wise distributions provide the framework for identifying whether CCF volumes scale purely with cavity capacity or whether maximum and minimum volumetric bounds emerge as D increases—bounds that may reflect limits on accumulation or long-term mass-balance stability.

6.3.2.1. Fill fraction of CCFs in host craters

Figure 6.6 plots the V_{CCF}/V_{host} of my sampled CCFs, showing clear hemispheric differences. northern hemisphere fill fractions exhibit a weakly leptokurtic distribution (excess kurtosis $K = 0.22$) with pronounced left skew ($S = -0.78$) and a median of 0.73. In contrast, southern hemisphere fill fractions display a more symmetrical ($S = -0.051$), platykurtic distribution ($K = -0.4$) with a median of 0.52. Thus, volumetrically, northern hemisphere CCFs more frequently approach rim-full states, whereas southern hemisphere CCFs are more evenly distributed across partial infill states. This is consistent with the Chapter 5 finding that southern hemisphere host craters are (i) shallower, as defined by their hemisphere-specific $d_t:D$ ratios (Section 5.3.1), and (ii) have systematically higher d_a values (Section 5.3.4). Both of these factors reduce $d_c (= d_t - d_a)$, thereby lowering the likelihood of rim-full occupation states. The hemispheric contrast is genuine and not an artefact of the input parameters, because histograms of fill fraction calculated using alternative crater-shape and $d_t:D$ parameterisations (see Figure 6.7) closely resemble those in Figure 6.6.

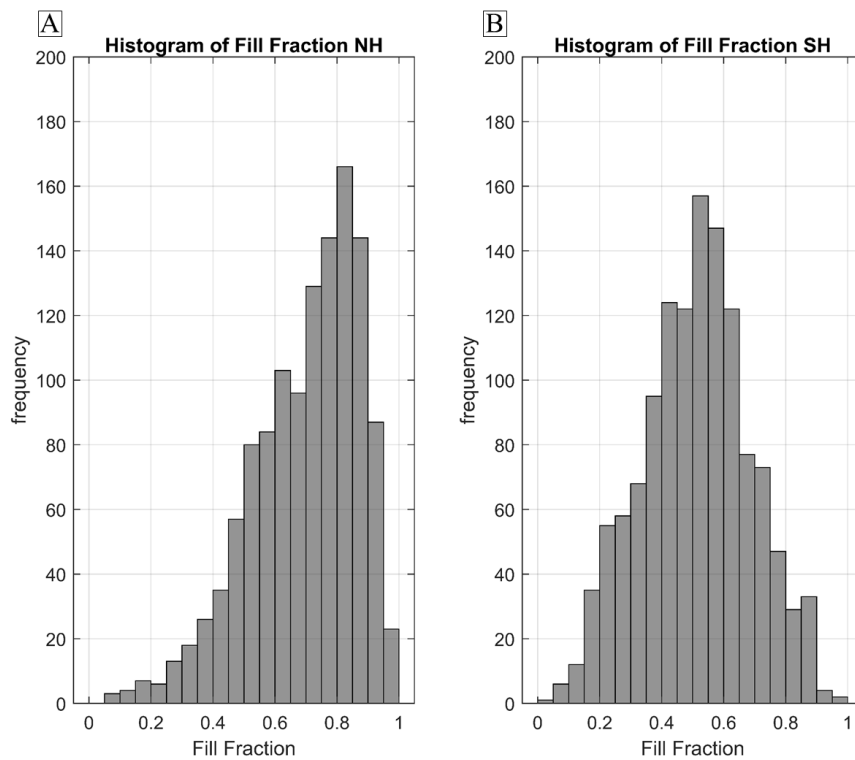


Figure 6.6. Histograms of CCF volume occupation fraction by hemisphere. V_{CCF} occupation expressed as a fraction of V_{host} in 5% bin widths for (A) Northern Hemisphere CCFs, (B) Southern Hemisphere CCFs.

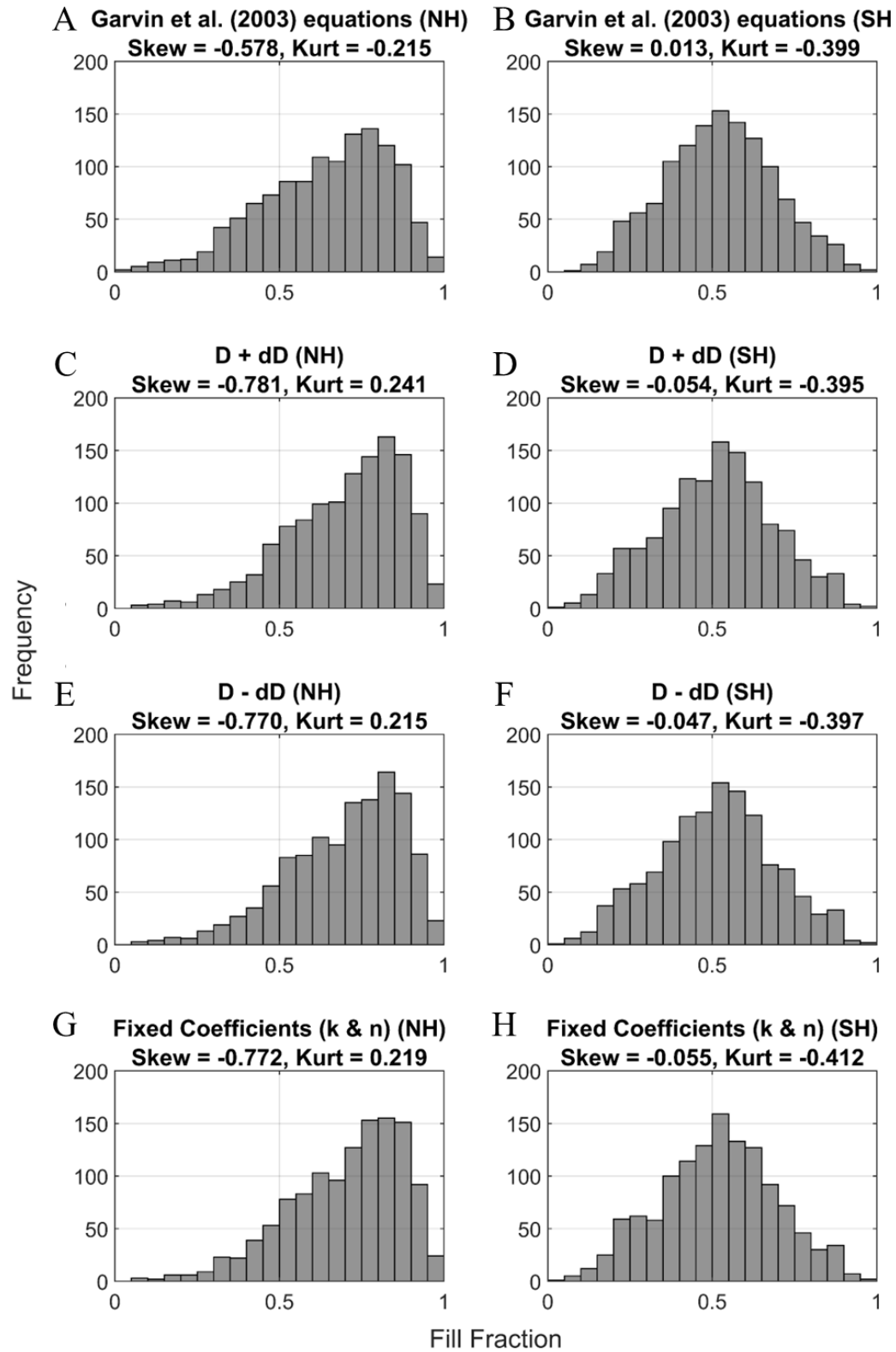


Figure 6.7. Sensitivity of CCF fill fraction distributions to alternative parameterisations. Panels show fill fraction histograms for the Northern (left column) and Southern (right column) Hemispheres. (A–B) d_i derived from Garvin et al. (2003), (C–D) $D + \sigma D$, (E–F) $D - \sigma D$, (G–H) fixed cavity-shape coefficients (k , n) from Garvin et al. (2003).

6.3.2.2. Variations of CCF volume and V_{CCF}/V_{host} with diameter

Figure 6.8 shows the distribution of V_{CCF} within discrete D bins in each hemisphere, plotted as the minimum, median, and maximum values attained within each bin. This binning does not represent uncertainty on individual volume estimates; rather, it is used to characterise the population-level volumetric envelope within each D range and to identify how both the upper and lower bounds of preserved CCF volume evolve with increasing crater size. As expected from basic crater scaling, the minimum, median, and maximum V_{CCF} all increase systematically with D globally and in either hemisphere (Figure 6.8). This behaviour primarily reflects the increasing volumetric capacity of larger crater cavities, consistent with the established d_r - D scaling relationships for Martian craters (e.g., Garvin et al., 2003), and with the geometric arguments developed in Section 6.3.2. However, absolute volume alone cannot distinguish between craters that are near capacity and those that are only weakly infilled. To isolate this effect, the same bin-wise analysis is therefore repeated using the normalised fill fraction, V_{CCF}/V_{host} (Figure 6.9).

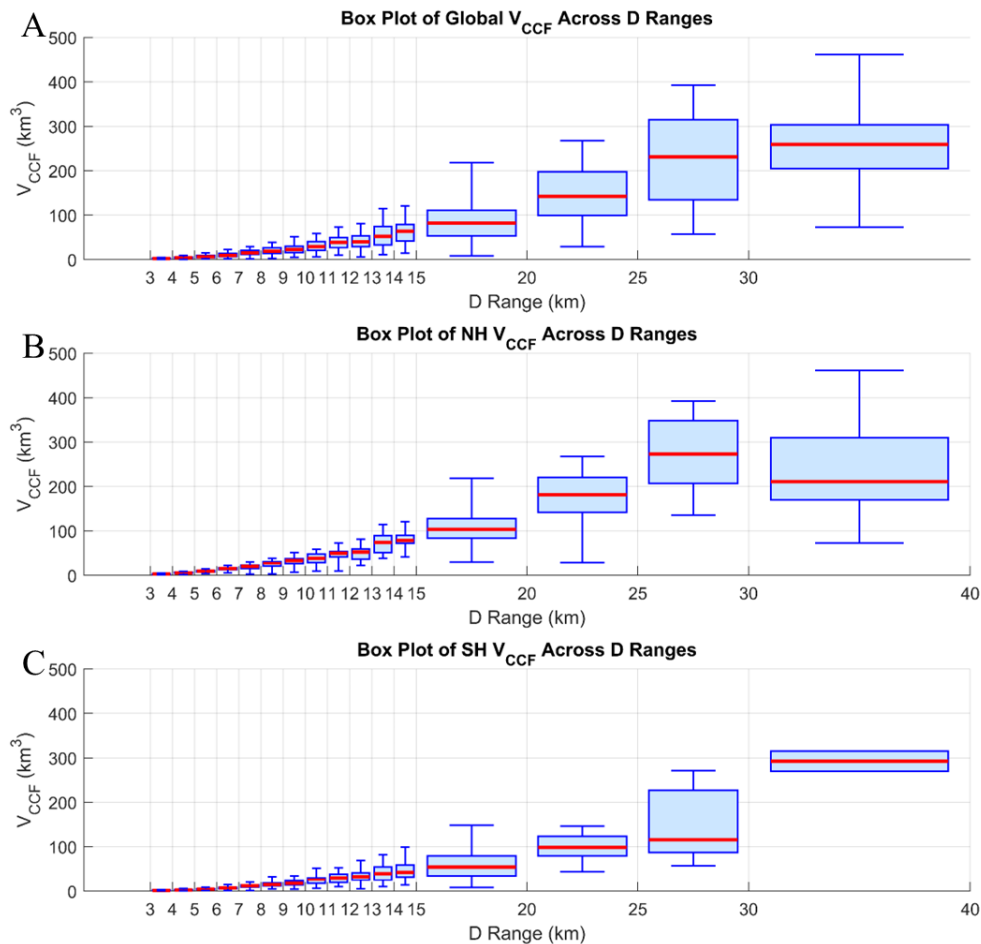


Figure 6.8. Box plot of CCF volume as a function of crater diameter. (A) Box plots of V_{CCF} (km^3) for the global high-confidence CCF population binned by D . (B) As in (A), but for northern hemisphere CCFs only. (C) As in (A), but for southern hemisphere CCFs only.

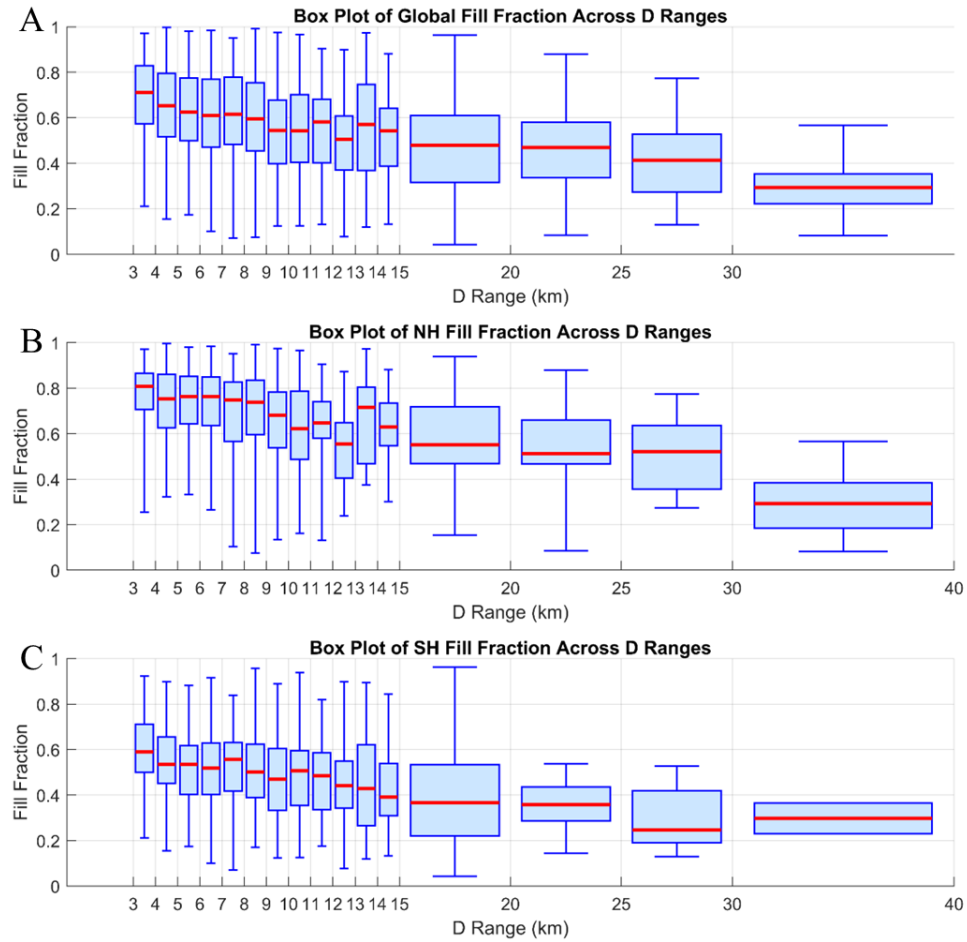


Figure 6.9. Box plots of crater fill fraction as a function of crater diameter. (A) Box plots of fill fraction for the global high-confidence CCF population binned by D . (B) As in (A), but for northern hemisphere CCFs only. (C) As in (A), but for southern hemisphere CCFs only.

When expressed as V_{CCF}/V_{hosts} , clear hemispheric contrasts emerge. In the northern hemisphere (Figure 6.9B; Table 6.3), the upper envelope remains consistently high across small to intermediate D , with maximum fill fractions exceeding ~ 0.9 for all bins up to $D \approx 25$ km. Median fill fractions are likewise high, remaining above ~ 0.7 across most D ranges, before declining in the largest bins, thus indicating that many northern CCFs occupy a large fraction of the available crater cavity volume. The lower envelope in the north, however, shows a different structure. For small craters ($D < \sim 7$ km), minimum fill fractions stabilise near ~ 0.25 – 0.30 , suggesting a relatively well-defined lower bound on preserved infill in this size range. At larger diameters, the lower envelope becomes more dispersed, locally dropping to ~ 0.1 in some bins before recovering to ~ 0.3 – 0.4 at larger D . This tightening of the envelope at higher D indicates that large craters permit a much smaller range of volumetric states. In the southern hemisphere (Figure 6.9C; Table 6.4), CCFs approach rim-full states less frequently, with maximum fill fractions generally between 0.83 – 0.95 . Median values decline more gradually than in the north, from ~ 0.6 in smaller craters to < 0.4 beyond $D \approx 20$ km. Minimum fill fractions remain consistently lower than in the north, rarely exceeding ~ 0.2 across most D bins and in several cases

falling as low as 0.05. Smaller craters ($D < 7$ km) maintain lower bounds of ~ 0.1 – 0.2 , whereas larger craters commonly permit minimum fill fractions below ~ 0.15 and only occasionally reach ~ 0.25 – 0.3 . This confirms that the hemispheric volume contrast identified in Section 6.3.2.1 persists across most of the crater-size spectrum.

These bin-wise envelopes demonstrate that V_{CCF} does not simply scale as a fixed fraction of crater capacity. Instead, both the maximum and minimum attainable volumetric states converge with increasing D , defining a widening admissible range of V_{CCF}/V_{host} . This behaviour is consistent with CCF volumes being bounded by competing constraints: upper limits imposed by accumulation efficiency, cavity geometry, and possible rim overflow, and lower limits imposed by incomplete infilling and/or long-term post-glacial mass loss. D therefore controls not only absolute storage capacity, but also the range of volumetric outcomes that can be preserved.

Having established that hemispheric differences in V_{CCF} and V_{CCF}/V_{host} cannot be explained by crater geometry alone, but instead likely reflect differences in CCF glacial history, the following section examines whether these contrasts also operate at finer spatial scales. By analysing the distribution of V_{CCF} and V_{CCF}/V_{host} as a function of latitude and elevation within D bins, the next stage tests whether the hemispheric asymmetry persists regionally or whether local climatic and topographic factors impose additional structure on CCF volumetric states.

Table 6.3 North Hemisphere V_{CCF} descriptive statistics

D Range (km)	N	Max V_{CCF} (km ³)	Min V_{CCF} (km ³)	Median V_{CCF} (km ³)	Maximum Host Crater Fill Fraction	Minimum Host Crater Fill Fraction	Median Host Crater Fill Fraction
2-4	193	4.79	0.79	2.91	0.97	0.26	0.81
4-5	214	9.07	1.72	5.41	1.00	0.32	0.75
5-6	174	14.66	4.06	9.33	0.98	0.33	0.76
6-7	121	22.44	5.52	15.41	0.98	0.26	0.76
7-8	115	29.59	2.57	19.93	0.95	0.10	0.75
8-9	82	38.75	2.80	26.11	0.99	0.07	0.74
9-10	56	51.33	6.71	33.19	0.97	0.13	0.68
10-11	59	58.56	9.48	38.68	0.97	0.16	0.62
11-12	29	73.11	10.10	49.67	0.90	0.13	0.65
12-13	25	81.15	21.89	52.40	0.87	0.24	0.56
13-14	21	114.44	38.74	73.64	0.97	0.37	0.72
14-15	25	120.78	41.06	79.06	0.88	0.30	0.63
15-20	70	218.44	29.44	103.69	0.94	0.15	0.55
20-25	25	267.47	28.99	181.09	0.88	0.08	0.51
25-30	11	392.45	135.01	272.83	0.77	0.27	0.52
30-40	5	461.28	72.94	211.36	0.57	0.08	0.29

Table 6.4 South Hemisphere V_{CCF} descriptive statistics

D Range (km)	N	Max V_{CCF} (km ³)	Min V_{CCF} (km ³)	Median V_{CCF} (km ³)	Maximum Host Crater Fill Fraction	Minimum Host Crater Fill Fraction	Median Host Crater Fill Fraction
2-4	180	0.57	1.84	3.40	0.92	0.21	0.59
4-5	186	0.75	2.89	6.01	0.90	0.15	0.54
5-6	180	1.72	4.61	8.89	0.88	0.17	0.54
6-7	156	1.28	7.18	15.01	0.92	0.10	0.52
7-8	101	1.75	12.05	20.85	0.84	0.07	0.56
8-9	100	5.04	15.15	32.02	0.96	0.17	0.50
9-10	86	4.87	17.84	34.36	0.89	0.12	0.47
10-11	68	6.21	24.90	51.83	0.94	0.12	0.51
11-12	38	10.21	29.55	52.31	0.82	0.18	0.48
12-13	30	5.83	32.36	69.20	0.90	0.08	0.44
13-14	24	10.66	39.09	81.87	0.90	0.12	0.43
14-15	27	14.45	42.35	99.02	0.84	0.13	0.39
15-20	68	8.60	53.98	148.11	0.96	0.04	0.37
20-25	14	43.59	98.34	146.06	0.54	0.14	0.36
25-30	7	57.43	115.36	271.35	0.53	0.13	0.25
30-40	2	270.14	292.51	314.88	0.36	0.23	0.30

6.4. Geographic variation in CCF volume

Having analysed the total V_{CCF} and its scaling with D , I proceed to examine the global spatial distribution of V_{CCF} and fill fraction (V_{CCF}/V_{host}) and how these quantities vary with latitude and elevation. Of particular interest is whether they show systematic trends with these variables, as found for d_a in Chapter 5. If found, such trends could offer further insights into CCF preservation and glacial history.

6.4.1. Spatial distribution of CCF volume and fill fraction

Figures 6.10 and 6.11 illustrate the hemispheric distribution of V_{CCF} and fill fraction. The spatial variation in V_{CCF} shows no simple environmental gradient, because a single V_{CCF} may arise from

fundamentally different geometric contexts. Any given V_{CCF} value—without explicit knowledge of each crater’s capacity—is ambiguous: it may reflect a low d_a , nearly rim-full deposit in a smaller crater, or a high- d_a , deposit within a much larger crater. As a result, the spatial distribution of V_{CCF} alone cannot reveal environmental controls such as latitude or elevation. Despite this, distinct regional clustering of similar-volume CCFs is evident in either hemisphere, with some regions dominated by low-volume CCFs (e.g., lower Arcadia Planitia, eastern Hellas, northern Promethei Terra) and others by high-volume CCFs (e.g., outer Utopia Planitia, Nilosyrtris Mensae, northern Alba Patera, southern Promethei Terra). Crucially, while high V_{CCF} regions occasionally overlap with areas of high CCF density (Section 4.4.5), this correlation is not universal. Furthermore, high V_{CCF} regions are not uniformly distributed across either hemisphere but instead cluster within discrete regions, implying that these regions offer conditions conducive to both the accumulation of substantial ice volumes and/or their subsequent preservation. The fill fraction (Figure 6.11) exhibits a poleward increase that appears rather striking in the northern hemisphere (where concentrations of high V_{CCF}/V_{host} occur in Arabia Terra, Utopia Planitia, and Nilosyrtris Mensae), and still apparent although less systematic in the southern hemisphere (where modestly high V_{CCF} values populate Terra Cimmeria and Terra Sirenum) – probably due to the narrower latitudinal range of CCFs there. The corresponding latitudinal dependences will be studied in 6.4.2.

Both V_{CCF} and fill fraction display distinct regional contrasts (Figures 6.10 and 6.11), which may indicate specific regional-scale controls/influences on the glacial history of CCFs. For instance, CCF populations in the lower-latitude sectors of Utopia and Arcadia Planitia, are characterised by uniformly low V_{CCF} ($<5 \text{ km}^3$) yet retain high fill fractions (>0.6). This combination reflects low- d_a CCFs confined to small host craters, experiencing minimal post-glacial ablation. The complete absence of high-volume CCFs in these sectors implies that larger craters, if present, do not currently host CCF. This absence suggests that paleo-CCFs either did not accumulate sufficient ice to achieve substantial fill or were unable to preserve ice through interglacial periods, which is suggestive of a supply-limited (either time or ice) accumulation regime.

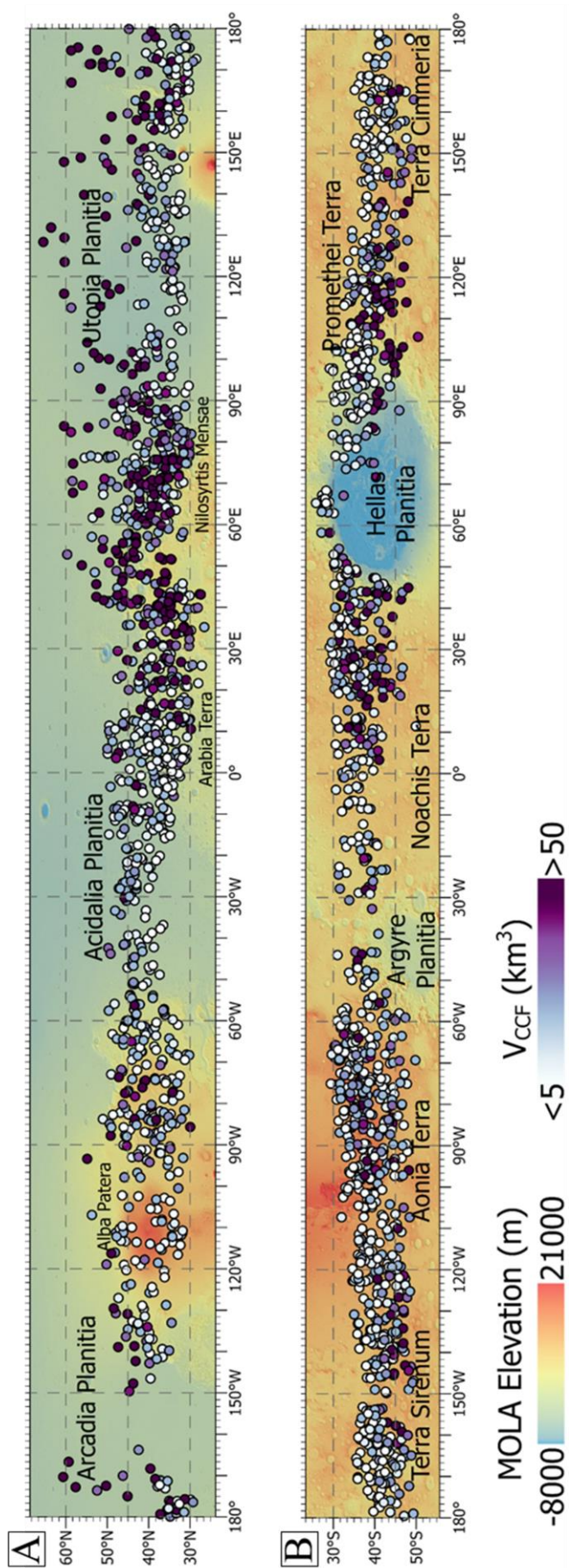


Figure 6.10. Spatial distribution of CCF volumes. (A) Northern hemisphere map showing the distribution of high-confidence CCFs coloured by V_{CCF} magnitude, overlain on MOLA topography. (B) As in (A), but for the southern hemisphere.

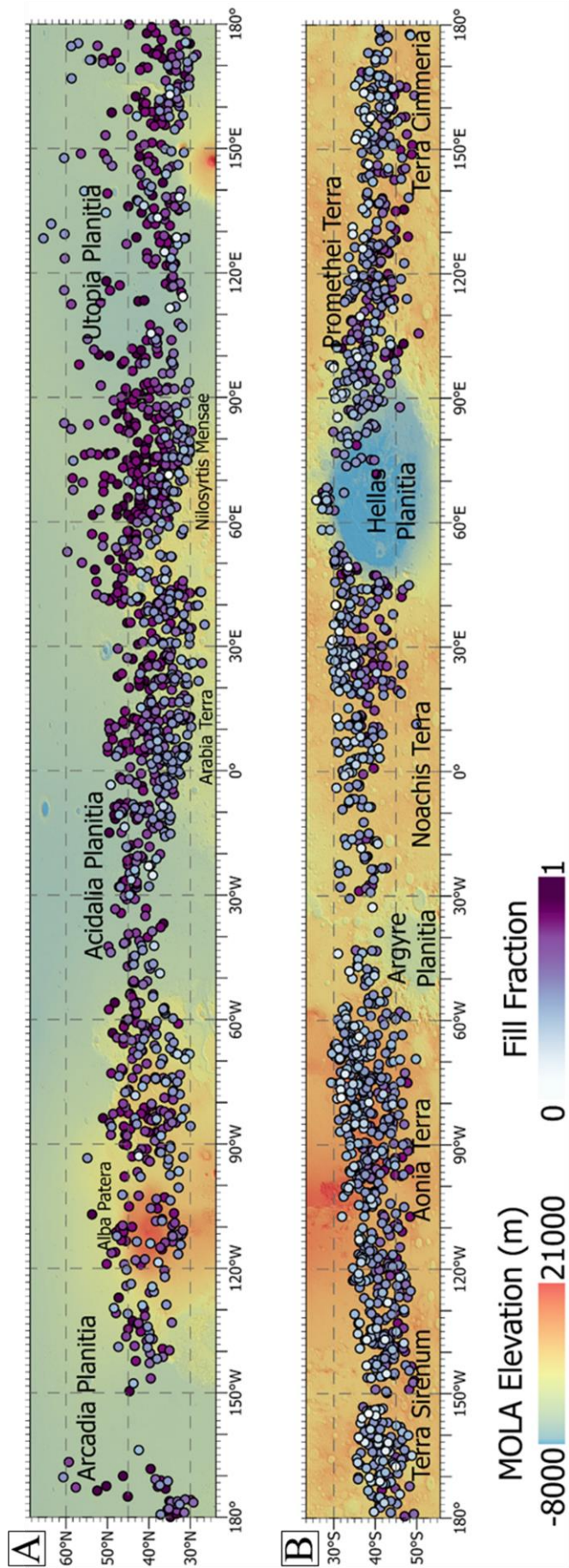


Figure 6.11. Spatial distribution of CCF fill fraction.

(A) Northern hemisphere map showing the distribution of high-confidence CCFs coloured by fill fraction, overlain on MOLA topography. (B) As in (A), but for the southern hemisphere.

In contrast, discrete clusters exhibit a combination of high V_{CCF} and high fill fractions are found in Arabia Terra, western Utopia Planitia, western Arcadia Planitia, and eastern Hellas. These CCFs represent substantial ice bodies, requiring both high net accumulation during mid-latitude glaciation coupled with minimal post-glacial ablation. However, these specific clusters present a striking anomaly: despite the clear evidence for high-volume accumulation and preservation, there is a complete absence of low-volume, high-fill-fraction CCFs within smaller host craters. It is improbable that any palaeo-CCFs formed within smaller craters within these regions experienced substantial post-glacial ablation whilst adjacent larger craters retained near-rim-full ice volumes.

The glacial histories of these clusters and their CCF populations may hold considerable potential to clarify mid-latitude ice evolution and warrant detailed future investigation. For the purposes of this chapter, however, these regional contrasts indicate that crater size—and likely their unknown local accumulation constraints—exerted the primary control on CCF volume and preservation.

6.4.2 Latitude trends in CCF volume and fill fraction

Figures 6.12 and 6.13 plot V_{CCF} and fill fraction against latitude for small ($D < 7$ km) and complex ($D > 7$ km) craters in each hemisphere. V_{CCF} exhibits no statistically significant dependence on latitude in either hemisphere (Figure 6.12), reflecting the dominant influence of crater capacity—larger host craters accommodate far greater absolute volumes irrespective of latitude, producing wide volumetric scatter across all latitudes. By contrast, fill fraction appears to show some latitude-dependent variation (Figure 6.13), with values increasing poleward, though the strength of this increase differs between hemispheres. In the north, fill fraction increases poleward, with statistically significant but weak correlations for both small ($D < 7$ km; $R^2 = 0.20$, $p < 0.001$) and large ($D > 7$ km; $R^2 = 0.06$, $p < 0.01$) host craters, and the range of fill fraction magnitudes narrows poleward, as low-fill CCFs become less common. In the south, fill fraction appears to increase with absolute latitude, yet no statistically significant correlation emerges, likely owing to the narrower latitudinal range of southern hemisphere CCFs. Overall, V_{CCF} remains highly scattered across both hemispheres, yet fill fraction tends to increase poleward, with low-fill CCFs becoming progressively less common at higher latitudes. This is consistent with the latitudinal decrease in d_a identified in Chapter 5.

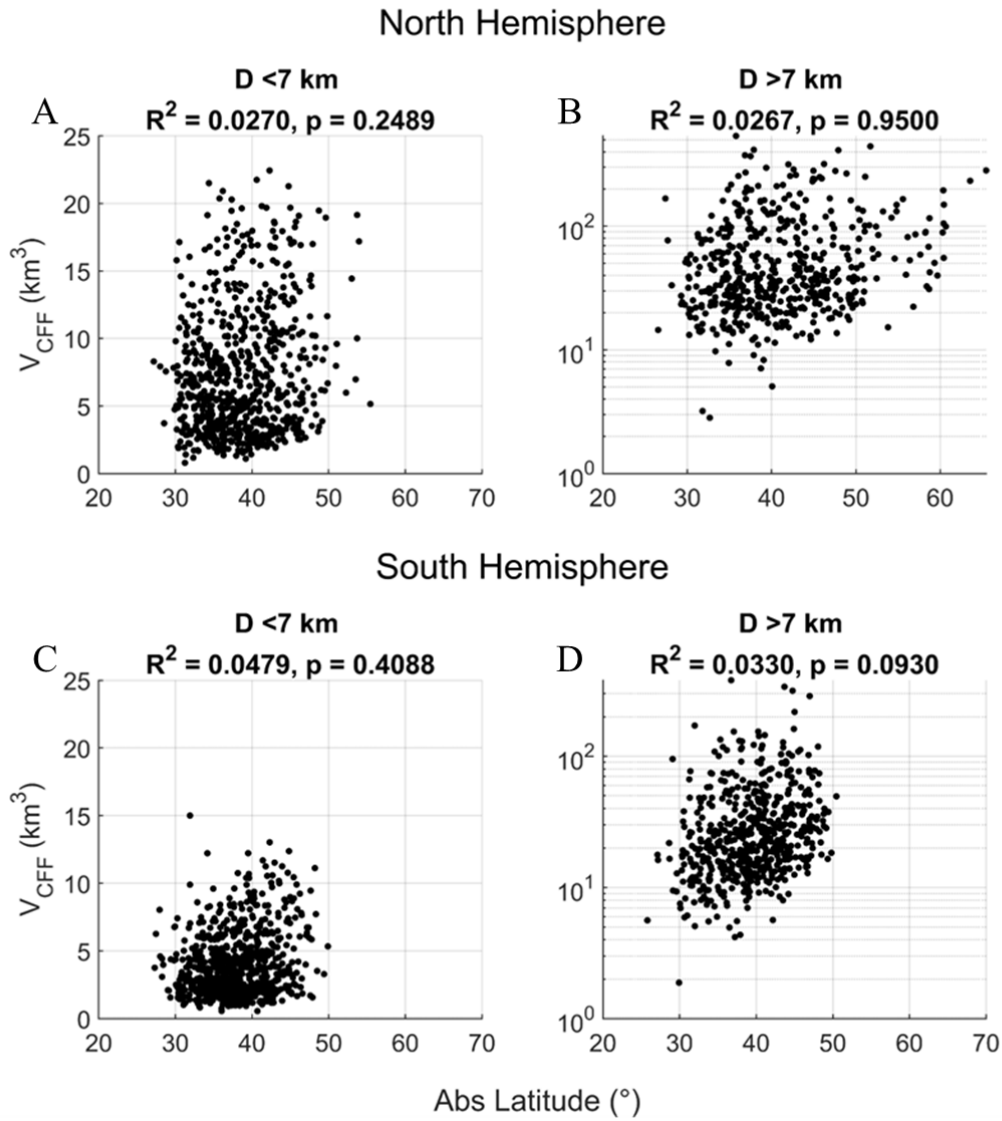


Figure 6.12. Latitudinal variation of V_{CCF} in either hemisphere. V_{CCF} (km^3) plotted against latitude for simple craters ($D < 7 \text{ km}$; A, C) and complex craters ($D \geq 7 \text{ km}$; B, D). Panels A and B show the Northern Hemisphere, and panels C and D show the Southern Hemisphere (absolute latitude). Note that panels B and D use a logarithmic y-axis to accommodate the wider range of V_{CCF} values in complex craters.

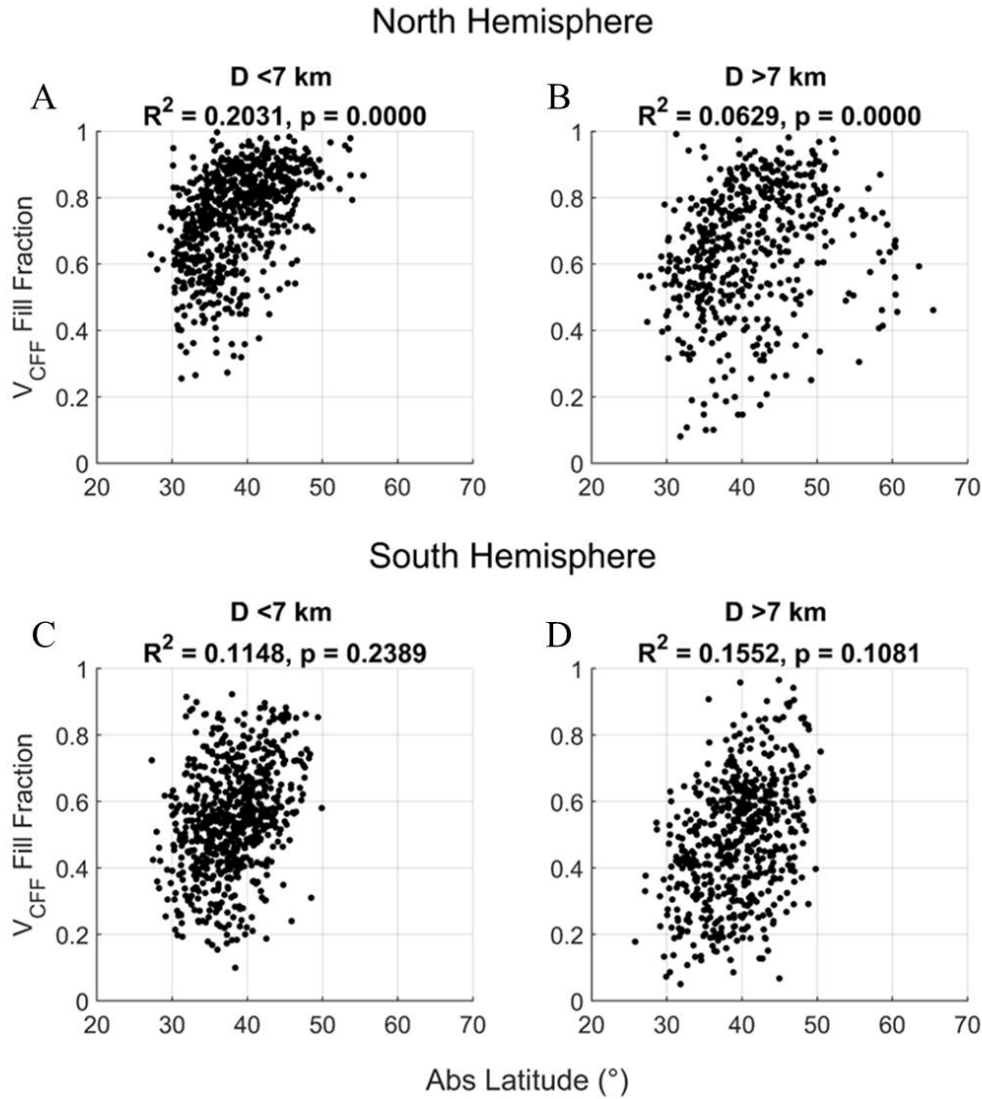


Figure 6.13. Latitudinal variation of CCF fill fraction (V_{CCF} / V_{host}) in either hemisphere. Fill fraction plotted against latitude for simple craters ($D < 7$ km; A, C) and complex craters ($D \geq 7$ km; B, D). Panels A and B show the Northern Hemisphere, and panels C and D show the Southern Hemisphere (absolute latitude)

Next, I break down the analyses of V_{CCF} and fill fraction versus latitude into 16 discrete D bins (the same bins as in Chapter 5), to account for differences in the host-crater capacity, which was shown to influence V_{CCF} and fill fraction systematically (Sec. 6.3.2). The plots for all bins appear in Appendix A. Figs. 6.14 and 6.15 here show a few binned plots to illustrate typical findings. In the north, V_{CCF} increases with latitude in seven D bins ($p < 0.05$), although all but one exhibit minimal explanatory power ($R^2 < 0.1$). The weak correlations likely reflect the influence of increasing crater capacity with D , as even within narrow 1 km D bins, larger craters accommodate substantially greater absolute volumes masking any systematic latitudinal signal. A further limitation is that each D bin mixes data from all longitudes; any latitude-linked variation in V_{CCF} may emerge more clearly within restricted longitudinal sectors, a point explored in Section 6.5. When expressed as fill fraction (Figure 6.15), the latitudinal dependence becomes slightly clearer. Ten bins yield statistically significant trends ($p < 0.05$),

though roughly half remain negligible ($R^2 < 0.1$) and the remainder indicate weak positive relationships ($R^2 \leq 0.32$). Thus, any latitude-linked influence on CCF fill fraction is modest and expressed unevenly across D bins. In the south, V_{CCF} shows no systematic latitudinal dependence (Figure 6.14), with only one D bin (3–4 km) yielding a statistically significant result ($p < 0.05$; $R^2 = 0.01$). When expressed as fill fraction (Figure 6.15), only two bins display statistically significant correlations with latitude, though both are weak ($R^2 = 0.06$ and 0.08). Although statistically significant correlations are absent, the data distributions exhibit a geometric constraint: the lower envelope of points rises poleward, indicating that minimum V_{CCF} tends to increase with latitude even where overall correlations remain weak.

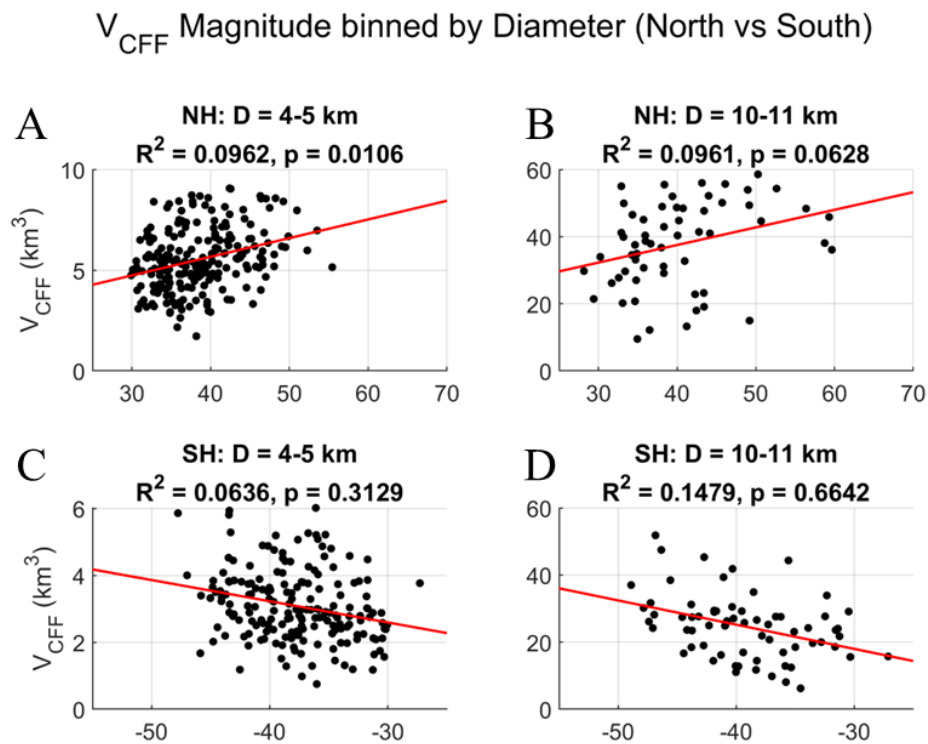


Figure 6.14. Latitudinal variation of V_{CCF} across hemispheric subsets when binned by D . V_{CCF} plotted against latitude for simple craters (D 4–5 km; A, C) and complex craters (D 10–11 km; B, D). Panels A and B show the Northern Hemisphere, and panels C and D show the Southern Hemisphere (absolute latitude). Red lines denote linear regressions, with associated R^2 and p -values provided for each subset.

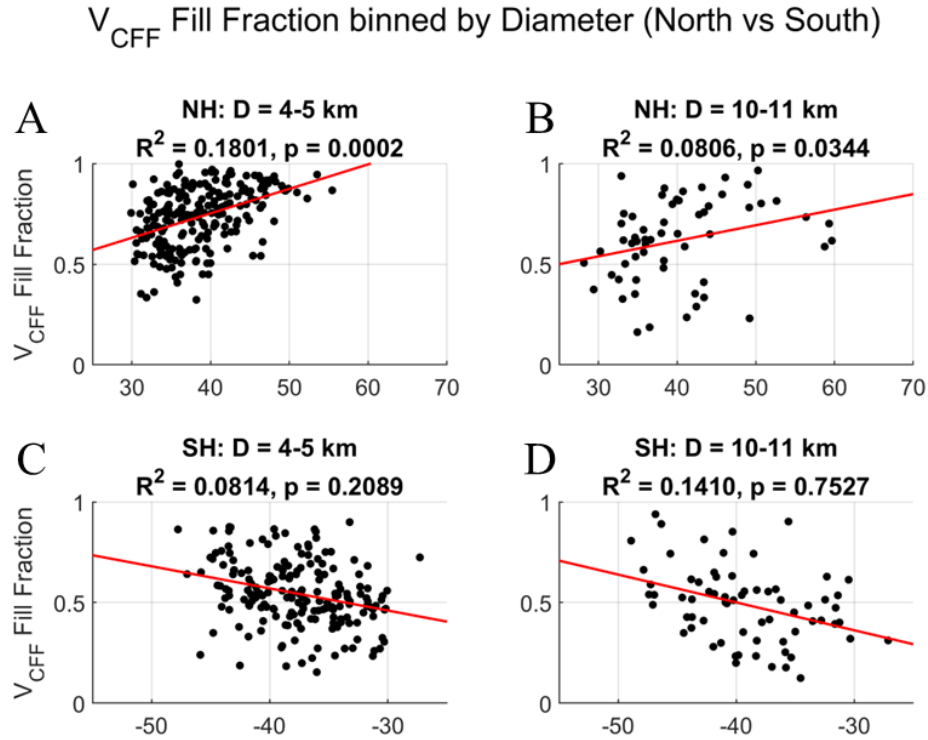


Figure 6.15. latitudinal variation of fill fraction across hemispheric subsets when binned by D . Fill fraction plotted against latitude for simple craters (D 4–5 km; A, C) and complex craters (D 10–11 km; B, D). Panels A and B show the Northern Hemisphere, and panels C and D show the Southern Hemisphere (absolute latitude). Red lines denote linear regressions, with associated R^2 and p -values provided for each subset.

6.4.3. Elevation trends in CCF volume and fill fraction

V_{CCF} and fill fraction show statistically-significant but negligible correlations with elevation in each hemisphere ($p < 0.05$, $R^2 < 0.052$; Figures 6.16–6.17), suggesting that if they carry an elevation signal, it is overwhelmed by the strong effects of crater size (Sec. 6.3.2). But the same analyses for different crater-diameter bins uncover systematic elevation trends for V_{CCF} and fill fraction. Figures 6.18 and 6.19 show the results for a few bins, while Appendix A contain the plots for all bins.

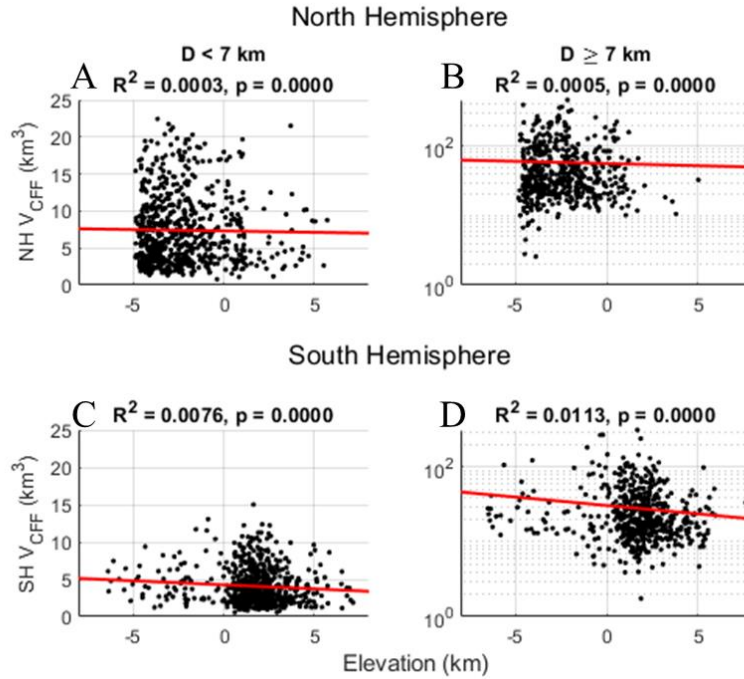


Figure 6.16. Elevation variation of V_{CCF} in either hemisphere. V_{CCF} (km^3) plotted against elevation for simple craters ($D < 7 \text{ km}$; A, C) and complex craters ($D \geq 7 \text{ km}$; B, D). Panels A and B show the Northern Hemisphere, and panels C and D show the Southern Hemisphere. Note that panels B and D use a logarithmic y-axis to accommodate the wider range of V_{CCF} values in complex craters.

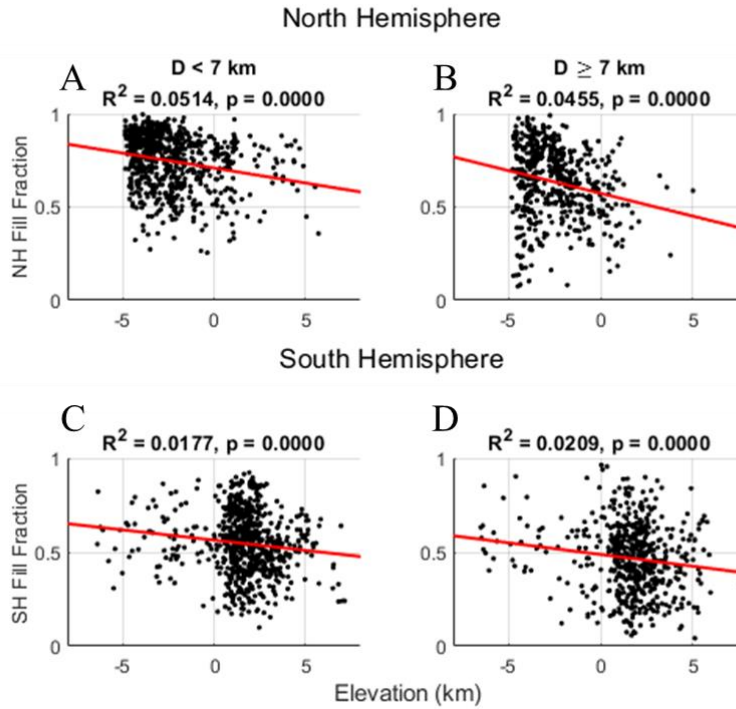


Figure 6.17. Elevation variation of V_{CCF} in either hemisphere. V_{CCF} / V_{host} plotted against elevation for simple craters ($D < 7 \text{ km}$; A, C) and complex craters ($D \geq 7 \text{ km}$; B, D). Panels A and B show the Northern Hemisphere, and panels C and D show the Southern Hemisphere.

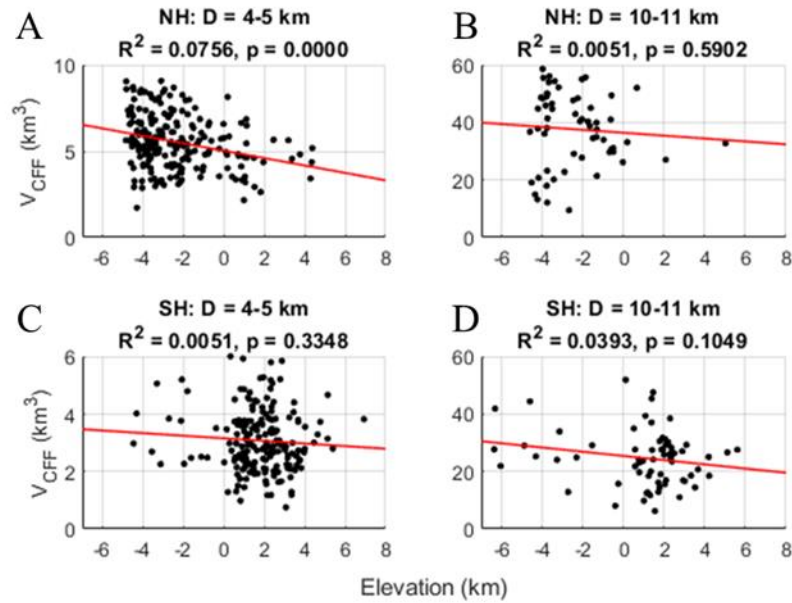


Figure 6.18. Elevation variation of V_{CFF} across hemispheric subsets when binned by D . V_{CFF} plotted against elevation for simple craters (D 4–5 km; A, C) and complex craters (D 10–11 km; B, D). Panels A and B show the Northern Hemisphere, and panels C and D show the Southern Hemisphere (absolute latitude). Red lines denote linear regressions, with associated R^2 and p -values provided for each subset.

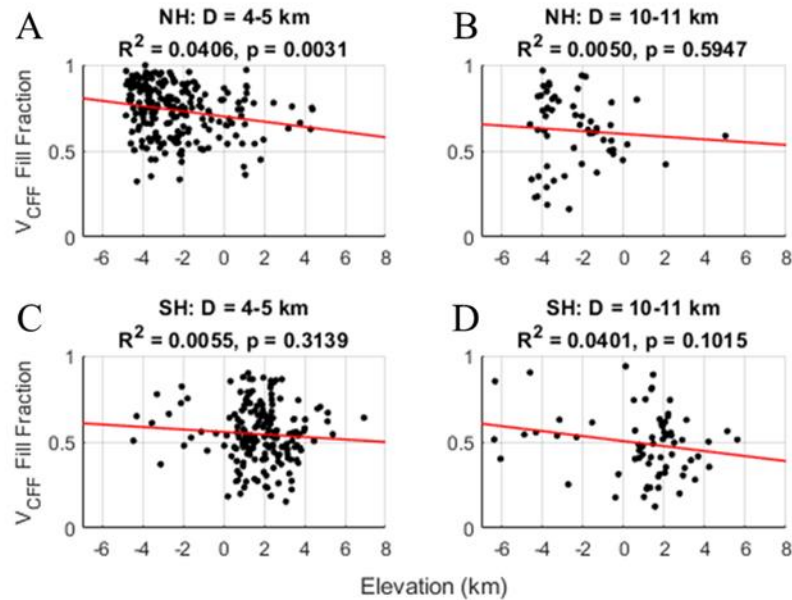


Figure 6.19. Elevation variation of fill fraction across hemispheric subsets when binned by D . Fill fraction plotted against elevation for simple craters (D 4–5 km; A, C) and complex craters (D 10–11 km; B, D). Panels A and B show the Northern Hemisphere, and panels C and D show the Southern Hemisphere (absolute latitude). Red lines denote linear regressions, with associated R^2 and p -values provided for each subset.

Binned by D , V_{CCF} exhibits statistically significant correlations with elevation in both hemispheres ($p < 0.05$). For CCFs in smaller host craters ($D < 10$ km), correlations are weak ($R^2 < 0.1$), whereas CCFs within larger craters ($D > 10$ km) show stronger associations, with R^2 values reaching ~ 0.3 . When expressed as fill fraction, the same overall pattern is observed. Correlations remain statistically significant across most D bins ($p < 0.05$) but are weak for CCFs within small craters ($D < 10$ km; $R^2 < 0.1$) and more variable in larger host craters (R^2 up to 0.4).

In the south, generally across D bins both V_{CCF} and fill fraction decreases with elevation. At the lowest elevations, CCFs retain relatively high but narrowly distributed volumes, indicating uniformly well-filled craters. Across the elevation zone of peak CCF occurrence (0–5 km), volumes become more widely dispersed, reflecting greater variability in infill state, before declining in both magnitude and range at higher elevations. A comparable pattern is observed in the north, where V_{CCF} and fill fraction also tend to decline with increasing elevation.

Generally, CCFs in either hemisphere retain greater and more uniform infill volumes at lower elevations, whilst CCFs at higher elevations possess more variable V_{CCF} and typically less completely filled, reflecting the combined effects of elevation and crater capacity on CCF preservation.

6.4.4. Summary

D remains the dominant control of V_{CCF} , reflecting its geometric role in defining host capacity and d_r – D scaling that underpin all volumetric estimates. By contrast, latitude and elevation, though statistically significant predictors of d_a , show little explanatory power for V_{CCF} at the hemispheric scale. This decoupling suggests that while d_a is sensitive to broad environmental gradients, total preserved V_{CCF} may be more strongly conditioned by site-specific factors such as local topography, accumulation efficiency, or post-depositional modification.

When examined within discrete D bins, weak but spatially variable trends with latitude and elevation emerge—most clearly for larger craters in the northern hemisphere—indicating that some regional dependence exists, but it is neither universal nor systematic. Relative fill fraction, meanwhile, increases modestly poleward, implying enhanced preservation potential or greater initial accumulation at higher latitudes, mirroring the poleward increase of d_a identified in Chapter 5, yet any volumetric trend remains highly scattered and subordinate to local variability.

These results indicate that V_{CCF} does not simply scale with the same first-order climatic or geometric parameters that control d_a but instead may reflect a more complex interplay between crater capacity and regional/local environmental conditions. Consequently, volumetric variability is best interpreted at the regional rather than hemispheric scale—a rationale that underpins the subsequent analysis of isovolumetric contours, which aims to resolve these spatial differences in accumulation and preservation more explicitly.

6.5. Isovols

The concept of *isovolumetric contours* (hereafter *isovols*) represents a novel approach for examining crater-confined ice deposits on Mars. Not found in the existing Mars ice/VFF literature, it is put forward for the first time in this Thesis. An isovol is a curve of constant V_{CCF} in d_a – D space. Different CCFs from different-sized host craters but having the same volume (of material/deposit/ice) lie on an isovol. The system of isovols provides a morphometric framework for comparing CCF volumes in d_a – D space, either across the full population or within specific subsets.

The physical behaviour of isovols reflects the requirement that, for a fixed CCF volume, d_a increases with D as the same V_{CCF} is distributed within a wider and deeper cavity. This can be illustrated using two hypothetical CCF volumes placed within 7, 10, and 20 km diameter host craters. A CCF volume of 23.6 km³ would completely fill the 7 km D crater ($d_a \approx 0$), but its surface would lie progressively farther below the rim in the 10 and 20 km craters as crater depth and cavity capacity increase. Similarly, a smaller CCF volume of 8.25 km³ would only partially fill the 7 km D crater (fill fraction ≈ 0.35); (producing $d_a \approx 0.5$) and would become progressively thinner and more recessed in the 10 and 20 km craters. These examples demonstrate that isovol trajectories extend towards larger d_a values as D increases, while lower-volume trajectories approach the geometric limit at which d_a approaches d_t and the remaining CCF thickness becomes negligible.

Departures from these curves reveal how CCF volumes vary, forming the basis for the analyses developed in this section. CCFs with identical d_a in different-sized craters must possess different volumes, and therefore follow trajectories that are shallower than the isovols, demonstrating that V_{CCF} increases with D . Conversely, CCFs with identical D but differing volumes must display different d_a values, with larger volumes requiring greater d_c within the same cavity and therefore reduced d_a . These principles define the structure of the d_a – D domain and provide the basis for interpreting the distribution of CCFs in subsequent analyses.

Four conceptual end-member scenarios (Figure 6.20) illustrate how observed trajectories may intersect or follow isovols and thus reveal distinct volumetric behaviour. In the first scenario, the trajectory of CCF d_a – D data is shallower than the isovol curves, meaning that V_{CCF} increases with D . In the second, the trajectory rises more steeply than the isovols, meaning that d_a increases more rapidly than expected for constant-volume behaviour; here, V_{CCF} decreases with D . In the third scenario, the data lie along a single isovol, indicating near-constant V_{CCF} across D . The fourth scenario shows a dispersed d_a distribution with no systematic relation to D , producing no coherent alignment with isovols; in this case V_{CCF} is likely controlled by crater scale microclimates and reflects localised differences in accumulation or preservation.

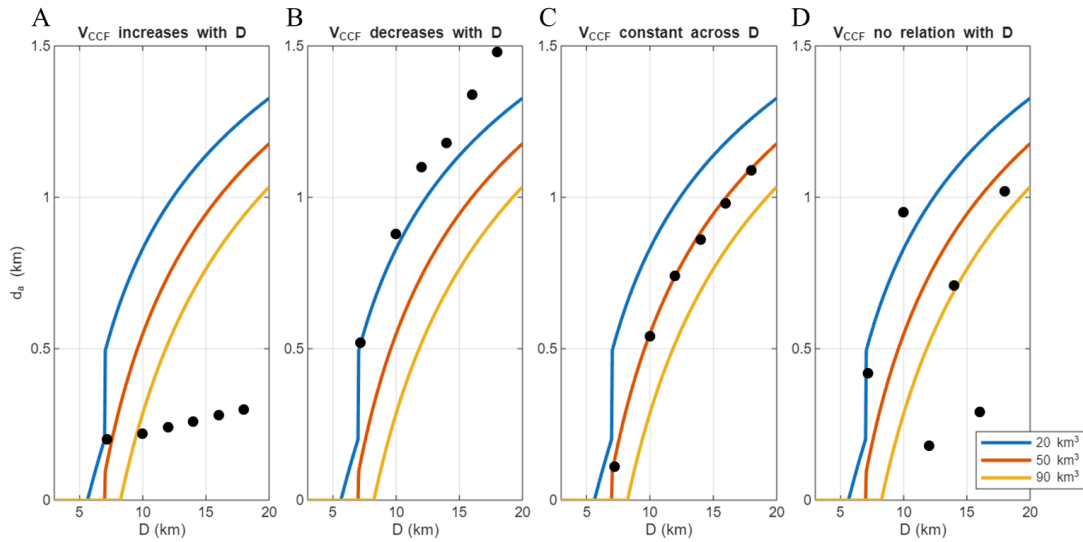


Figure 6.20. Conceptual isovol scenarios illustrating potential volumetric relationships among North hemisphere CCFs in d_a – D space. Coloured curves represent isovols (lines of constant V_{CCF}) derived from the crater cavity geometry. Black dots illustrate hypothetical population-scale trajectories in d_a – D space. Depending on whether these trajectories are shallower than (A), steeper than (B), aligned with (C), or unrelated to the isovols (D), V_{CCF} respectively increases with D , decreases with D , remains approximately constant, or shows no systematic dependence on D .

The rationale for introducing isovols is twofold. The first rationale concerns the identification of volumetric limits on CCF persistence as a function of host-crater size. In this context, isovols enable testing whether the processes and factors determining the current states of CCFs and their evolution towards the current states (e.g. formation, accumulation/ablation, preservation) involve volumetric thresholds or constraints. This rationale is examined by grouping CCFs into D bins and identifying the maximum d_a attained within each bin. Each maximum d_a corresponds to a specific retained volume that can be interpreted in d_a – D space using the isovol framework. If the maxima from multiple D bins align along a single isovol, this would indicate convergence toward a common minimum V_{CCF} across crater sizes. Such behaviour would imply that CCF preservation is governed not by a fixed d_a or d_c —which necessarily vary with crater geometry—but by a minimum volume required to maintain long-term mass balance through repeated glacial–interglacial cycles.

Second, d_a – D data of a population of CCFs within a region can be compared against the isovol curves to assess how the distribution of their V_{CCF} values vary with D , and therefore to infer regional mass-balance histories. Departures of regional CCF populations toward higher or lower isovols indicate systematic differences in retained volume for a given crater size, providing a means to assess spatial variability in accumulation efficiency and long-term mass balance that is not resolvable at the hemispheric scale.

Among these end-member behaviours, trajectories shallower than the isovols—implying that V_{CCF} increases with D —are expected to be the most common outcome, reflecting the fundamental scaling of cavity capacity with crater size. In this case, regional differences are expressed not by convergence toward a fixed volume state, but by systematic offsets between isovols, which indicate differences in retained volume for a given crater size and therefore differences in long-term mass balance and the local controls influencing this. By contrast, trajectories that steepen relative to the isovols, align with a single isovol, or show no coherent relationship to the isovol framework indicate fundamentally different volumetric controls, including supply limitation, strong post-glacial ablation, or dominance of local-scale effects. Accordingly, the position and orientation of regional d_a – D trajectories relative to isovols provides a physically grounded means of diagnosing whether CCF evolution in a given region is governed primarily by geometric capacity, volumetric thresholds, or post-glacial modification.

In the following analysis, this framework is applied in two stages. First, the method is outlined and the procedure for identifying maximum d_a and corresponding volume thresholds is established. Second, isovol relationships are examined across several regions of the northern hemisphere to evaluate the large-scale spatial variability in volumetric scaling of the high confidence CCF population.

6.5.1. Isovols: constant volume contours

By rearranging Eqn. 6.3, we find that a CCF with volume V in an idealised crater of diameter D (with known cavity-shape parameterisation) has the thickness

$$d_c = \left(\frac{V \cdot k \left(\frac{2}{n} \right) \cdot (2+n)}{\pi n} \right)^{\frac{n}{2+n}}. \quad (6.12)$$

The apparent depth is $d_a = d_t - d_c$ (Fig. 6.1), where d_t is given by

$$d_t = k(D/2)^n. \quad (6.13)$$

It follows that for any given pair of D and V values, the corresponding d_a is

$$d_a = k(D/2)^n - \left(\frac{V \cdot k \left(\frac{2}{n} \right) \cdot (2+n)}{\pi n} \right)^{\frac{n}{2+n}} \quad (6.14)$$

For CCFs having a fixed V , this equation defines an isovol on the d_a – D plane.

6.5.2. Maximum apparent depth and volume thresholds

Using the isovol framework, here I examine whether CCF preservation is constrained by a minimum volumetric threshold or by geometric limits on d_a . Comparison of maximum d_a values across successive D bins with their corresponding isovol contours allows assessment of whether consistent volumetric thresholds emerge independent of host crater cavity size.

Across both hemispheres, maximum d_a increases with D up to ~ 7 km, after which d_a values tend to plateau (Figure 6.21). At $D > 5$ km, d_a stabilises within ~ 0.6 – 1.1 km for the southern hemisphere, while in the north, for $D > 7$ km, d_a shows greater variability but remains confined to ~ 0.6 – 1.4 km. This transition suggests that for smaller craters ($D < 7$ km), deeper infill is required to sustain preservation, but for $D > 7$ km the d_a necessary for long-term stability no longer increases substantially with D .

Isovol overlays place this geometric transition in volumetric context. Across both hemispheres, minimum preserved V_{CCF} increase with D , consistent with geometric scaling, yet maximum d_a values in many bins converge within narrow volumetric ranges. This convergence suggests that CCF stability is governed not solely by cavity geometry but also discrete volumetric thresholds necessary to maintain mass-balance equilibrium. In the southern hemisphere (Figure 6.21B), CCFs in small craters ($D < 8$ km) retain low d_a yet occupy a consistent volume range of ~ 1 – 3 km³. For $D > 8$ km, maximum d_a values group into recurring isovol bands: 8–11 km and 12–13 km bins fall within ~ 5 – 7 km³; 11–12, 13–14, and 15–20 km bins align near ~ 11 – 13 km³; and the 14–15 km bin (~ 17 km³) lies just above this envelope. At larger D (20–30 km), maximum volumes cluster around ~ 20 km³. Although not every bin aligns perfectly, successive bins repeatedly converge on a small set of volumetric plateaus, indicating that CCF preservation is regulated by threshold volumes rather than by smooth geometric scaling alone.

In the north (Figure 6.21B), the maximum d_a distribution is scattered across D bins, yet the volumetric trend remains coherent. V_{CCF} still increases with D , and successive bins show alignment within recurrent isovol ranges. Small craters ($D \approx 4$ – 9 km) cluster along two narrow volume bands (~ 2.8 – 3.2 km³ and ~ 4.9 – 5.5 km³), defining a consistent lower bound on preserved V_{CCF} . At intermediate diameters ($D \approx 10$ – 25 km), bins group into higher but similarly constrained ranges: 10–12 km craters at ~ 11 – 12 km³, 13–15 km at ~ 45 – 49 km³, and 15–25 km at ~ 36 – 38 km³. Some D bins (e.g., 9–10, 25–30, and 30–40 km) lie outside these envelopes; however, their volumes still increase with D in line with geometric expectations.

Across both hemispheres, present-day CCFs require minimum volumes that increase with D , yet maximum preserved examples frequently stabilise within discrete volumetric bands rather than scaling smoothly with cavity size (Figure 6.21). For $D > \sim 7$ km maximum d_a varies only weakly with D (~ 0.6 – 1.4 km), indicating that long-term stability is governed primarily by achieving a threshold d_a thickness sufficient to maintain mass-balance equilibrium under post-glacial ablation, with V_{CCF} rising secondarily to sustain this depth in increasing host crater cavities.

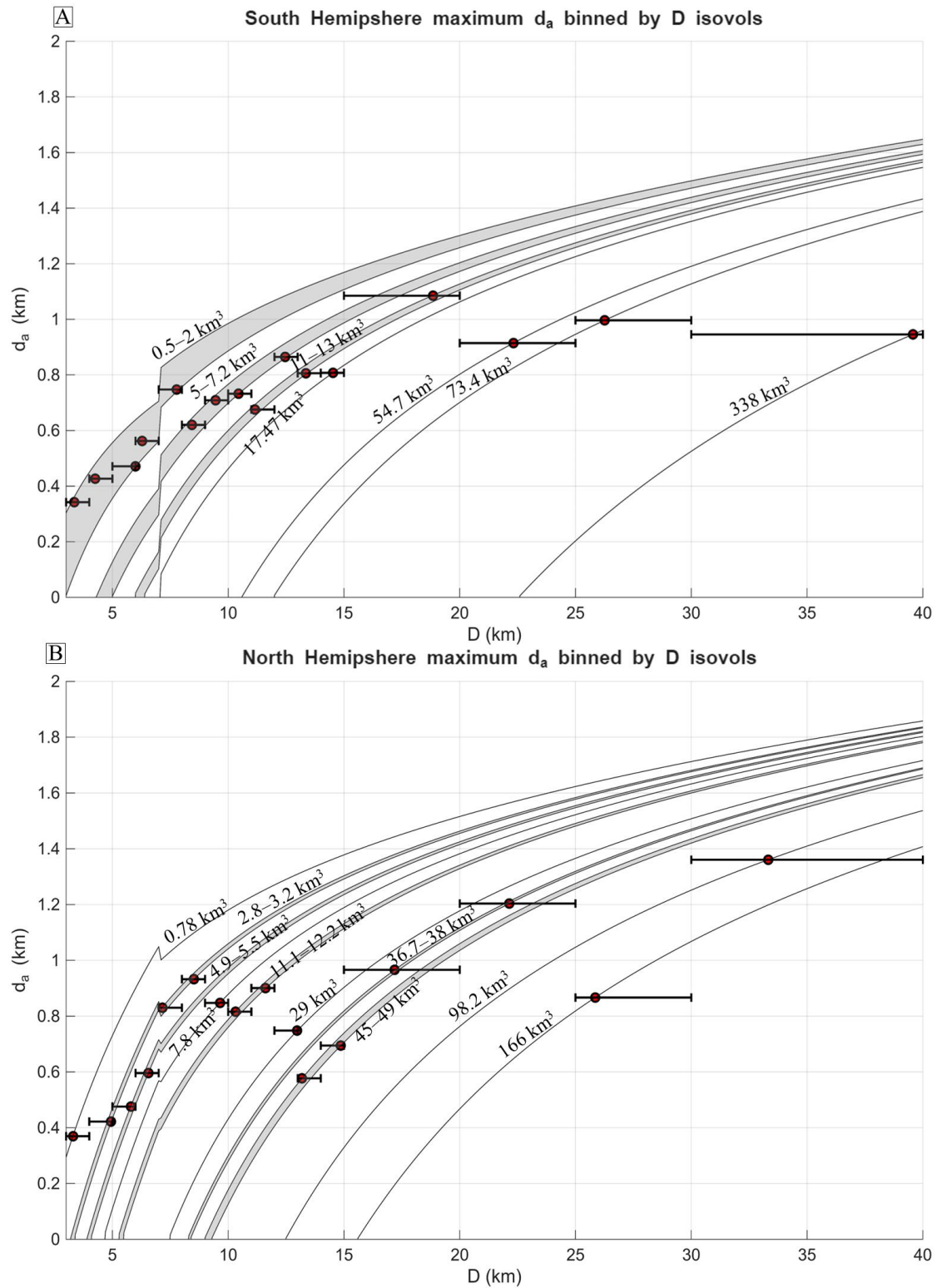


Figure 6.21. Maximum apparent depth and corresponding isovol thresholds across crater diameter bins. Maximum d_a values measured in successive D bins together with the associated isovol contours for (A) the Northern Hemisphere and (B) the Southern Hemisphere.

6.5.3. Regional analysis of V_{CCF} against isovols

This analysis applies the isovol framework to assess regional variability in CCF d_a – D trajectories across the northern mid-latitudes, plotting each region against a standardised set of isovol contours to evaluate how V_{CCF} scale with D . Across all regions (Figure 6.22), CCF d_a – D trajectories are systematically shallower than the isovol system, confirming that V_{CCF} increases with D rather than remaining constant. However, the steepness of these trajectories differs markedly between regions (Figure 6.22), indicating that volumetric scaling and preservation efficacy are not spatially uniform. These variations demonstrate that regional climatic and geomorphic conditions have modified CCF volumes beyond purely geometric expectations. Comparison of regional slopes—both relative to one another and to the mean northern-hemisphere trajectory (0.0229)—provides a diagnostic means of identifying where such influences have enhanced or suppressed post-glacial preservation.

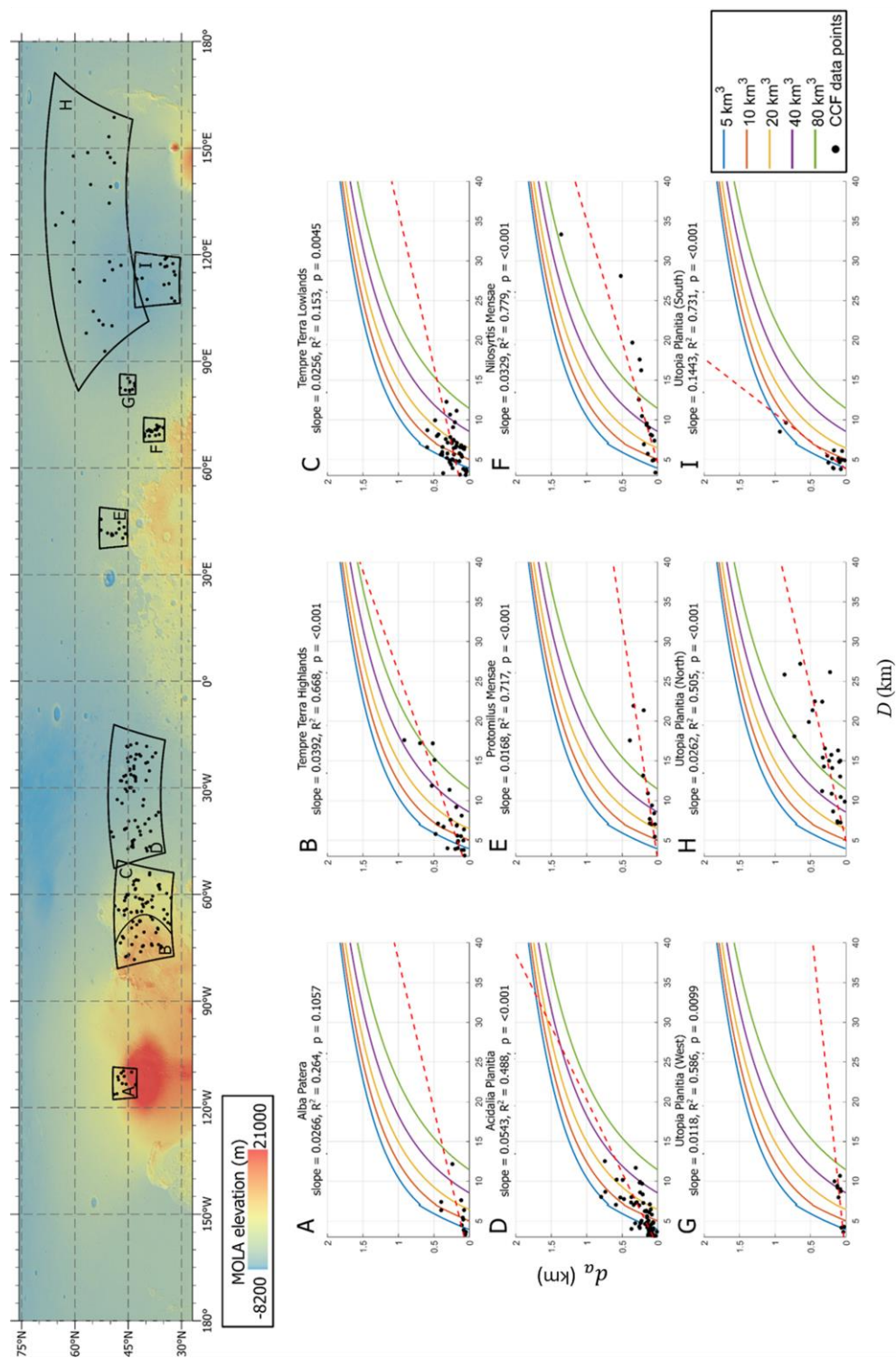


Figure 6.22. Regional isovol analysis of the northern hemisphere. (Top panel) Spatial distribution of nine northern hemisphere regions plotted over MOLA gridded topography and hillshade, showing the local groupings of CCFs mapped within specified regional boundaries (A–I). (A–I) Regional d_v - D distributions for the nine northern hemisphere subdivisions plotted against constant-volume isovols (5–80 km³). Red dashed lines indicate regional linear trends with annotated slopes, R^2 , and p -values. Each regional distribution illustrates that V_{CCF} increases with D .

Among the tested regions, Protonilus Mensae (Figure 2.22E) and Utopia Planitia (West) (Figure 6.22G) exhibit the shallowest d_a – D slopes (0.0168 and 0.012 respectively), both falling below the northern-hemisphere mean (0.0229). These trajectories indicate reduced regional post-glacial ablation and near-equilibrium preservation of V_{CCF} , regardless of D . The subdued slopes may reflect local conditions that favour long-term stability, such as enhanced debris availability forming sublimation lags, increased self-shading from background slopes, or higher accumulation efficiency during glacial phases.

Utopia Planitia (South) and Arcadia Planitia exhibit the steepest d_a – D trajectories (0.144 and 0.050) in the dataset, exceeding the hemispheric mean by an order of magnitude in the case of Utopia South. Both regions coincide with areas of high CCF occupancy (Chapter 4), indicating glacial provinces that once supported extensive CCF formation but now record strong post-glacial ablation.

A comparable intra-regional contrast is observed within Tempe Terra (Figure 6.22B, C), where subtle topographic variation exerts a measurable influence on CCF preservation. The highlands and lowlands differ in mean elevation by only ~ 1.8 km yet display distinct d_a – D trajectories: highland CCFs follow a steeper slope (0.039) and lowland CCFs a shallower trajectory (0.026). The resulting slope difference ($\Delta = 0.013$, equivalent to ~ 13 m km $^{-1}$) aligns closely with the ~ 14 m offset predicted from elevation alone, based on the MLR-derived gradient of 0.0079 per kilometre of elevation change (Chapter 5). This agreement indicates that the divergence between the two subregions primarily reflects the minor but statistically significant elevation gradient identified in Chapter 5, reaffirming the weak yet consistent influence of altitude on CCF d_a .

Across the regional analyses, all tested northern-hemisphere sectors exhibit the same fundamental trend, that V_{CCF} increases with D , although not all associations are statistically significant. This directional consistency confirms that volumetric scaling is broadly consistent with the hemispheric pattern established in Chapter 5. Regional differences in the steepness of these trajectories reflect the same environmental gradients identified previously in Chapter 5 (latitude and elevation).

When examined at a more localised scale, however, additional spatial variability emerges (Figure 6.23). Subdivision of Arcadia Planitia, into smaller CCF clusters reveals that while most local populations maintain the general trend of increasing V_{CCF} with D , some deviate from this behaviour. In two lower-latitude clusters, d_a – D trajectories steepen beyond the isovol system, indicating reduced volumetric preservation and stronger ablation influence—likely reflecting the latitudinal gradient in d_a identified in Chapter 5. These findings demonstrate that, although regional averages capture the dominant scaling trends, localised clusters may record distinct accumulation–ablation histories that are sensitive to subtle topographic or climatic variations. Exploring these intra-regional patterns lies beyond the scope of this Thesis but provides a promising avenue for future work, particularly in assessing whether neighbouring CCF populations exhibit coherent shifts through the isovol framework indicative of local mass-balance variation.

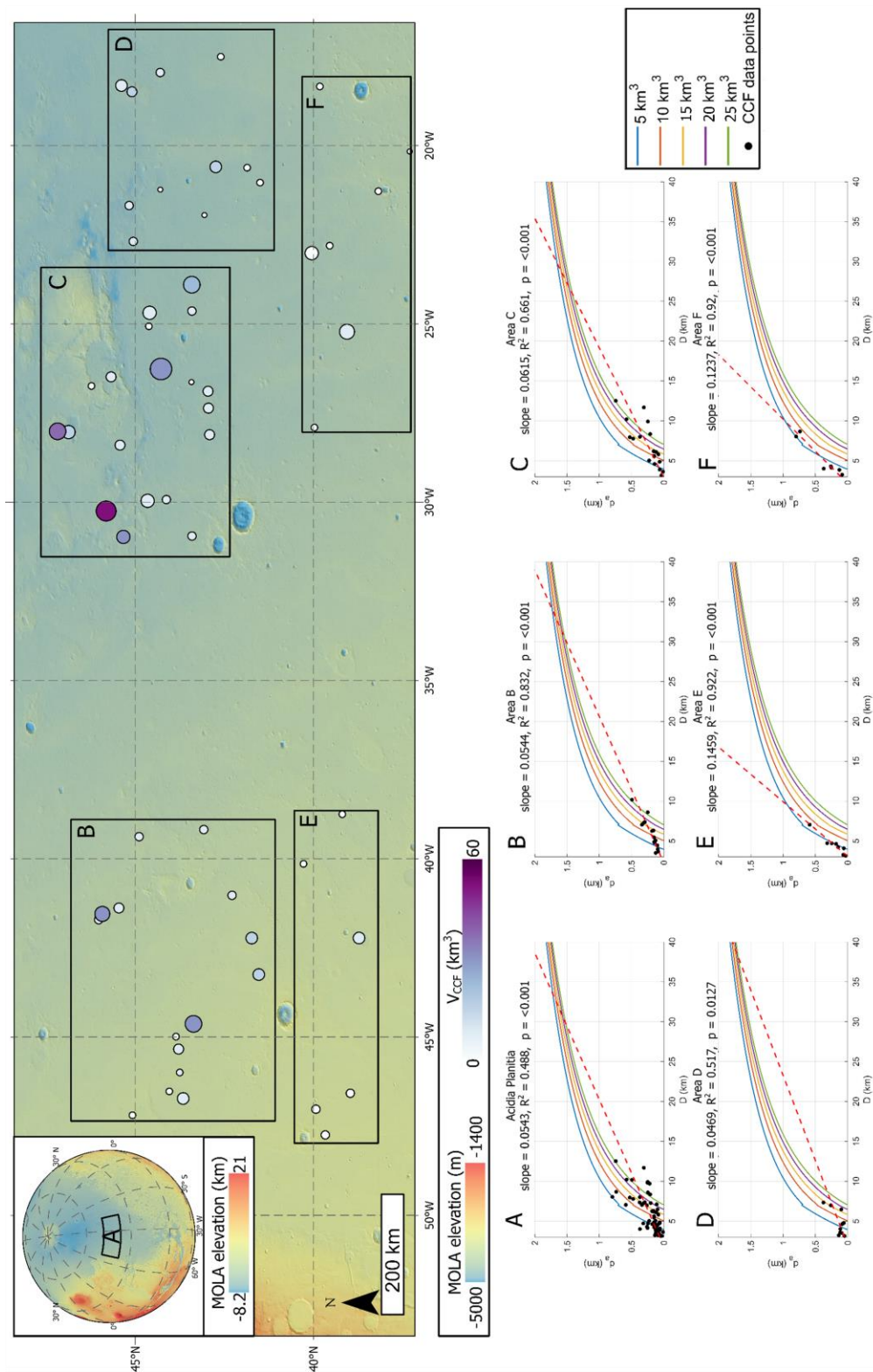


Figure 6.23. Isovol analysis of Arcadia Planitia. (Top panel) Spatial distribution of the five Arcadia clusters plotted over MOLA gridded topography and hillshade, showing the local groupings of CCFs sized by D and coloured by V_{CCF} . (A) Regional d_e - D isovol analysis for the entirety of Arcadia Planitia (Figure 6.22D). (B-F) d_e - D distributions for the five subdivisions of Arcadia Planitia plotted against constant-volume isovols (5–25 km³). Red dashed lines indicate regional linear trends with annotated slopes, R^2 , and p -values. Each cluster illustrates local deviations from, or adherence to, the broader hemispheric trend of increasing V_{CCF} with D .

6.5.4. Summary

The isovol analysis tested V_{CCF} using both maximum d_a values across D bins and regional d_a – D trajectories in the northern hemisphere. The results show that V_{CCF} increases with D and that preserved CCFs exhibit little alignment with isovol contours. Across both hemispheres, maximum d_a values rise with D and converge within recurring volumetric bands, indicating that preserved CCFs occupy discrete V_{CCF} ranges rather than aligning along constant-volume contours. These bands may demonstrate that long-term preservation requires threshold volumes that increase with D , with larger cavities sustaining greater V_{CCF} even though d_a stabilises above ~ 7 km. As a result, the maximum- d_a trends show only limited isovol alignment but a clear increase in preserved V_{CCF} with D .

Regional analysis shows that CCFs in all tested northern-hemisphere provinces follow the same fundamental trend: their d_a – D trajectories cross the isovol system, confirming that V_{CCF} increases with D rather than remaining constant. However, the steepness of these trajectories varies between regions, indicating that volumetric preservation has been differentially modified by local environmental conditions. When regions are subdivided into smaller spatial clusters, additional variability emerges, suggesting that some local populations record distinct accumulation–ablation histories superimposed on the broader geometric trend. Although detailed exploration of these local controls lies beyond the scope of this chapter, their expression within the isovol framework highlights the potential for future work to explore fine-scale mass-balance variation across the mid-latitudes.

6.6. Conclusion

This chapter has reassessed the volumetric significance of mid-latitude CCFs, focusing on three core aspects: quantifying their cumulative ice content, evaluating the degree to which they fill available crater cavity, and testing how these relationships vary across scales and between hemispheres. In addition, the chapter introduced an isovol framework, allowing preservation patterns to be recast in terms of volumetric thresholds rather than d_a limits. This approach not only refines the scale-dependence of CCF stability but also provides new insight into mid-latitude glaciation.

Here, I return to the key questions outlined in the introduction and summarise the principal findings:

1. What is the total volume and GEL of CCFs?

This Thesis' CCF sample has an estimated volume of $6.22 \pm 0.27 \times 10^4$ km³, equivalent to 0.39 ± 0.026 m GEL at 90% ice purity. This represents a modest downward revision compared to earlier global estimates of 9533 CCFs (Levy et al., 2014), likely reflecting both refined d_a : D ratios and the use of D -dependent cavity equations in addition to a smaller sample size.

2. What proportion of host crater capacity is typically filled by CCFs?

When expressed relative to host crater capacity, CCF fill fraction reveal strong hemispheric contrasts. In the northern hemisphere, CCFs frequently approach rim-full states, with median fill fractions typically above ~ 0.7 for small-to-moderate craters and maxima exceeding 0.9 up to D of ~ 25 km. Minimum values, however, are more variable, ranging from ~ 0.25 – 0.30 in the smallest craters to ~ 0.1 in some larger bins. By contrast, the southern hemisphere exhibits lower overall occupation: maximum fill fractions rarely exceed 0.95, medians decline steadily from ~ 0.6 in small craters to < 0.4 at larger D , and minima are consistently subdued, commonly between 0.05–0.2. These hemispheric differences in V_{CCF} reflect the d_a offset identified in Chapter 5, where southern CCFs exhibit systematically higher d_a at a given D , attributed to elevation-dependent sublimation rates (Chapter 5). Consequently, northern CCFs more readily achieve higher fill fractions, whereas southern CCFs—despite shallower crater cavities—preserve proportionally smaller fill fractions of their total capacity.

3. How does this volume vary by crater diameter, latitude, hemisphere, and across different regions?

As expected, D exerts the dominant first-order control, with median V_{CCF} increasing systematically with D in either hemisphere, consistent with geometric scaling. When examined against latitude, no clear population-wide trend emerges, whether considered in absolute terms or as fill fraction, although some weak correlations appear in the northern hemisphere when data are binned by D . Similarly, elevation shows little explanatory power at the population scale ($R^2 < 0.05$), but clearer relationships emerge in the north when restricted to larger craters, with fill fraction decreasing with elevation, particularly in the 13–20 km D range (R^2 up to 0.32).

In the south, V_{CCF} shows no meaningful latitudinal dependence, with only one D bin yielding a statistically significant but negligible result ($R^2 \approx 0.01$). When expressed as fill fraction, two bins show weak latitude trends ($R^2 < 0.1$). Elevation exerts a clearer influence, with both V_{CCF} and fill fraction generally decreasing with elevation in addition to low-elevation CCFs retaining higher and more uniform volumes. This parallels the northern pattern, indicating that elevation exerts a weak but consistent control on infill state in either hemisphere.

Whilst large-scale controls are relatively weak, hemispheric asymmetries further reinforce the conclusion that ice accumulation and preservation conditions differ systematically between north and south.

4. Do CCFs within a given region occupy similar ice volumes? Does the isovolumetric framework reveal thresholds for CCF preservation?

CCFs within a given region do not converge on uniform volumes nor display isovol alignment. Regional d_a – D trajectories consistently increase with D and cross the isovol system,

indicating that preserved V_{CCF} scales upward with host-crater size rather than clustering along constant-volume contours.

Isovol analysis, however, offers tentative evidence for preservation thresholds. Maximum d_a values in successive D bins define the minimum preserved V_{CCF} , and these minima rise with D along trajectories shallower than the isovol curves, confirming that minimum V_{CCF} increases with diameter. Although CCFs do not align with any single isovol, maximum d_a values often stabilise within narrow volumetric bands—most clearly for $D \gtrsim 7$ km—suggesting that long-term preservation may require exceeding discrete volumetric limits, even if these limits are only weakly expressed.

5. To what extent can volumetric thresholds be used to explore and reconstruct the accumulation and preservation history of CCFs?

Volumetric thresholds provide a robust means of reconstructing accumulation and preservation histories. Regional d_a – D trajectories consistently cross the isovol system, demonstrating that V_{CCF} increases with D but that the slope of this trajectory varies between regions, recording differences in environmental modification. These regional slope differences record the same latitude- and elevation-dependent gradients identified in Chapter 5, reinforcing that the environmental controls previously shown to influence d_a also shape its volumetric scaling across regions. Regions with shallow trajectories indicate efficient preservation and reduced post-glacial ablation, whereas regions with steep trajectories show enhanced mass loss. Subdivision into local crater clusters reveals further variability, identifying clusters where trajectories exceed the isovol system and therefore record stronger ablation or reduced initial accumulation. These spatial variations demonstrate that volumetric thresholds offer an effective diagnostic tool for distinguishing geometric scaling from environmental overprinting and for inferring how accumulation efficiency and ablation intensity have varied across the mid-latitudes.

Beyond their significance for assessing present-day ice storage, the V_{CCF} estimates derived in this chapter provide a quantitative basis for evaluating the duration and pacing of mid-latitude glaciation on Mars. By constraining both the absolute volumes preserved within individual craters and the degree to which these volumes scale with host-crater capacity, the results presented here place first-order limits on the amount of ice that must have accumulated during glacial episodes. These volumetric constraints are directly exploitable in simple scaling arguments and in comparison, with numerical ice-flow models, allowing the plausibility of single-cycle versus multi-cycle emplacement to be tested. In Chapter 8, these V_{CCF} values are therefore used to examine accumulation timescales and to assess whether the observed CCF volumes can be reconciled with formation during a single orbital excursion or instead require cumulative growth across multiple glacial cycles.

6.7 References

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Chapter 7: CCF Surface Slope and Aspect

7.1. Introduction and background

A completely different avenue of probing the factors behind CCF evolution is to study their surface slopes and aspects—such morphometric analysis being a natural extension of the work on apparent depth in Chapter 5. Orientation and slope measurements have been widely used in global analyses of VFFs to infer patterns of ice accumulation and preservation. Studies of VFF slope and aspect have so far focused mainly on GLFs and crater-confined VFFs. Analyses of GLFs surface slope and aspect show strong poleward-facing orientation preferences in both hemispheres, commonly accompanied by a secondary east-facing bias (Milliken et al., 2003; Souness et al., 2012). For crater-hosted VFFs, Dickson et al. (2012) inferred preferred flow directions from the observable pattern of surface ridges and lobate morphology (Figure 7.1), finding that the crater-confined VFF population exhibits a latitudinal transition from crater-full CCFs at higher latitudes to pole-facing LDAs at lower latitudes that incompletely fill crater floors. These poleward biases have been interpreted as reflecting dominant insolation regimes and directional controls on ice stability, although those studies did not determine whether they reflect asymmetries in initial VFF accumulation or reflect their preservation, and the cause of the eastward bias remains unresolved. Similar analyses have been conducted on smaller-scale ice-related features such as gullies (e.g., Berman et al., 2005, 2009; Noblet et al., 2024), crater wall ice deposits (e.g., Hecht, 2002; Conway et al., 2018), crater-confined LDA like flow features (Sinha and Vijayan, 2017), and combined studies of CCFs alongside pole-facing flow features (Dickson et al., 2012). However, the surface slope and internal aspect bias of CCFs—which were previously categorised as concentric and thus orientation-neutral by Dickson et al. (2012) (Figure 7.1B)—have not yet been systematically analysed.

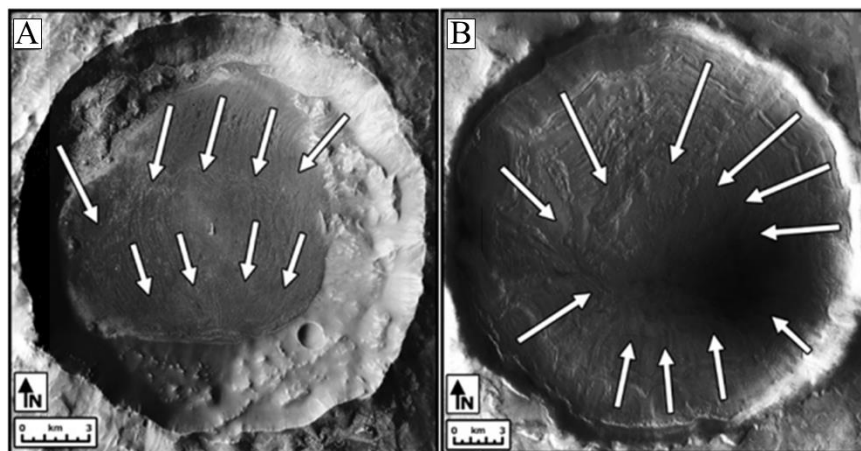


Figure 7.1. Examples of flow-orientation interpretations for crater-hosted ice features. Examples illustrating the inference of ice-flow orientation from surface morphology after Dickson et al. (2012). (A) Pole-facing flow interpretation for a crater at 31.8° N. (B) Concentric flow interpretation for a crater at 46.3° N. White lines indicate the interpreted flow-orientation patterns. Redrawn from Dickson et al. (2012).

Initial inspection of CCF surface geometry using a simple north–south transect across the same crater illustrated in Figure 7.1B reveals a small but systematic surface tilt, with a best-fitting slope of -0.2766° (Figure 7.2). Although the transect contains two opposing local slope segments—reflecting the concave nature of the CCF surface—the overall profile is not horizontal but exhibits a clear first-order gradient and therefore a definable preferential aspect. This raises the question of whether a similar slope or aspectual bias emerges when this analysis is applied to the wider CCF surface rather than a single transect segment. Importantly, this finding does not contradict the classification of such deposits as “concentric” by Dickson et al. (2012); a concave surface indicates inward flow from all margins, yet this concentric internal flow geometry does not require the final preserved surface to be perfectly level. This study addresses that gap by investigating fundamental outstanding questions about their morphology: Are CCF surfaces sloped? If so, what is the magnitude of these slopes? Do they exhibit systematic directional preference i.e. preferential aspect? If found, does this preferential direction align with other VFFs? Using this Thesis’ subset of high-confidence CCFs (Chapter 3), I explore these matters in this chapter.

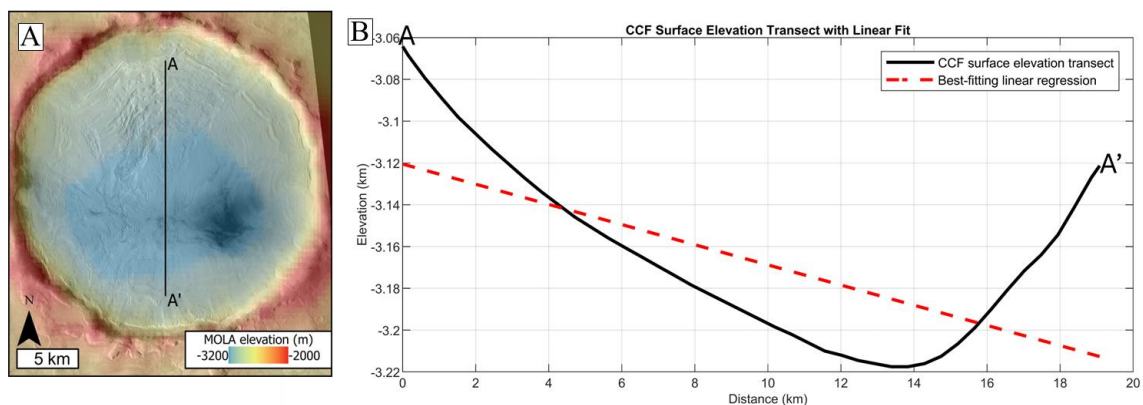


Figure 7.2. Example of CCF surface transect used to illustrate first-order surface slope. (A) The same CCF shown in Figure 7.1B, with the location of the north–south transect indicated by a black line (55.5° E, 46.3° N), overlain on MOLA elevation and CTX image: P19_008308_2265. (B) Elevation profile extracted along the transect in (A), showing the CCF surface topography (black line) and the best-fitting linear regression (red dashed line), illustrating a small but systematic first-order surface tilt.

This chapter first collects measurements to understand the fundamental characteristics of CCF slopes and aspects. Then, the chapter examines key systematic patterns in this data, proposing different scenarios for the processes and history of CCF accumulation, modification, and preservation to explain these patterns. If CCFs are found to possess statistically significant non-zero slopes, then some process must have driven this slope development, whether through slope, aspect, elevation or position-dependent variations in mass balance during their initial emplacement within their host crater (during glacial conditions), or via subsequent ablation (during interglacial periods) that exhibited such dependence. Even the absence of statistically significant slope will be interesting and might help in understanding CCF evolution, for example it may indicate that accumulation or preservation occurred without any spatial gradient that had directional preference, or that any initial slope has since been

obscured by the buildup of insulating overburden, deposited evenly across the surface and flattening any initial slope.

If CCF surface aspects cluster around specific directions, this may reflect the influence of insolation and wind regimes, and/or orogenic effects on ice deposition. Conversely, the absence of a regional trend could manifest in two distinct physical scenarios. First, the population might exhibit a stochastic aspect distribution, where individual CCFs possess measurable slopes but their orientations vary, with no observable bias. This may imply that insolation was modulated primarily by local microclimates—such as variable host crater rim shading—which effectively decoupled mass balance from cardinal direction, making individual crater geometry the dominant control. Second, the population might be characterised by a-directional uniformity (near-zero slope). In this case, the lack of aspect reflects a physical absence of surface gradient. This could signal a primary climatic signature, where conditions permitted spatially uniform mass balance across the entire CCF surface. Alternatively, this horizontality may result from secondary non-glacial modification: a sufficiently thick overburden of dust and debris could effectively mask the underlying ice, diffuse surface roughness and resetting the observed slope to zero. Regional or hemispheric aspect patterns, if present, could help identify spatial patterns in accumulation or preservation, refining our understanding of Martian glacial processes. Agreement with aspect patterns seen in other VFFs and other ice-related landforms and terrains (e.g., gullies) would support interpretations of shared climatic forcing. Significant divergence between the two populations may highlight how topographic confinement within a closed crater basin uniquely modulates CCF preservation, perhaps shielding the confined ice deposits from the environmental drivers that dictate the orientation of non-crater confined VFFs in the same region.

The aim of this chapter is to determine whether CCFs exhibit statistically significant surface slopes at the feature-scale—representing the macro-topography of the CCF surface as a whole, rather than localised or intra-CCF slope variations. If found, I then analyse how these slope magnitudes and directions vary regionally, hemispherically and latitudinally, using any systematic patterns found to derive insight into CCF evolution and glacial history. Sections 7.2 and 7.3, respectively, describe my methods and the results. In Section 7.4 I construct and examine different potential mechanisms behind the slope and aspect patterns I observe; the conceptual models I present variously invoke spatially variable mass balance during initial accumulation, modification through interglacial processes, and long-term preservation. To contextualise the work, I start in the following sections by reviewing relevant insights from terrestrial glaciology and published slope and aspect studies of Martian VFFs.

7.1.1. Terrestrial literature

Whilst global studies of VFF slope and aspect remain relatively limited, the influence of slope and aspect on glacier formation and persistence is well documented in terrestrial glaciology (e.g., Kruss and Hastenrath, 1987; Mölg et al., 2003; Kaser et al., 2004; Baroni et al., 2004; Evans 2006; Evans 2020),

with long-standing evidence that both factors interact with insolation and wind redistribution to influence mass-balance distributions (e.g. Evans., 1977; Evans, 2006). Terrestrial glaciers consistently show a strong preference for pole-facing aspects, due to reduced solar radiation receipt that favour net ice accumulation (Evans, 2006). Generally, discounting local topographical shading effects, glaciers formed on pole facing slopes receive less solar radiation (Ahmad and Rais, 1999), recede less and are expected to react more slowly to climate change than the equator-facing glaciers (Bhambri 2011). Additionally, rapid recession observed for west facing glaciers has been attributed to west-facing surfaces receiving maximum solar radiation during the mid-afternoon (Garg et al., 2017), and due to diurnal variation in cloud cover (Evans et al., 2006). This insolation-driven preference for pole-facing aspects is particularly evident in mid-latitude zones, at higher altitudes, and on steeper slopes (Evans and Cox, 2005; Evans, 2006).

Formation and survival of terrestrial glaciers is known to be regulated, in part, by prevailing wind directions, with glaciers often developing on the sheltered leeward sides of topographic obstacles relative to the dominant winds, where increased snow accumulation supports growth and melt rates are reduced (Evans, 1977; Sanders et al., 2013). Wind-driven processes, together with avalanches, enhance winter accumulation by redistributing snowfall across slopes and encouraging deposition in sheltered areas (Kuhn, 1995, 2003), particularly within small cirque glaciers situated in bowl-shaped depressions (Hoffmann et al., 2007). Some models of wind-driven remobilisation and deposition of snow in cirques, based solely on differential slope angle, have accounted for up to 65% of the spatial variation in winter accumulation within individual glaciers (e.g., Andrews Glacier; see Hoffmann et al., 2007 for discussion).

However, the relationship between wind and glacial accumulation is multifaceted; while leeward slopes benefit from drifting snow, windward slopes may experience higher accumulation if they are the primary interceptors of orographic precipitation—incoming water vapor carried by prevailing winds that condenses as air is forced upward (Elder et al., 1991; Zhang et al., 2017). This creates a distinction between wind-sheltered drift accumulation (leeward) and moisture-driven supply (windward), both of which can bias glacier distribution depending on the local climatic regime.

Glacier orientation is influenced not only by climatic factors but also by host topography and topographic shading interactions, where asymmetry or variations in landscape can heavily bias glacier aspects (Evans, 2006; Arnold et al., 2006). Host topographic slope asserts a degree of control on glacier distribution patterns, with slopes between 15–30° generally most conducive to ice cover (López-Moreno et al., 2006), whilst steeper slopes within glaciated altitudinal ranges can impede glacier development by preventing sufficient snow accumulation (Brozovic et al., 1997; Mittaz et al., 2002). The sensitivity of glacier mass balance to slope and aspect is latitude-dependent; at high latitudes, the low solar zenith angle minimises the radiation differential between sunward and poleward slopes,

causing regional latitudinal cooling to override local topographic shading as the primary control (Arnold et al., 2006)

Topographic shading induces spatial variations in surface mass balance of glacier surfaces by affecting the direct solar radiation available at the surface and by the proportion of the sky visible from any given location (Arnold et al., 2006). Numerous studies have quantified and assessed the decrease in shortwave radiative flux due to shading, for instance finding difference of up to 10% on Morteratsch Glacier in Switzerland (Klok and Oerlemans, 2002) and a 5.2% decrease for Haut Glacier d'Arolla (Arnold et al., 1996). Such reductions can influence both ablation and accumulation patterns thus exerting a significant influence on glacier behaviour, advance and survival.

Decades of terrestrial research have established a clear framework: slope and aspect are key factors of glacier accumulation, preservation, and climate sensitivity. Regionally consistent distributions in glacier aspects suggest regional-scale controls on the patterns of ice accumulation and preservation, whereas restricted aspectual distributions point to stronger controls from insolation—modulated in part by topographic shading—or wind. If comparable glaciological processes operated on Mars during mid-latitude glaciation, similar relationships between slope, aspect, and ice occurrence should be detectable in Martian VFFs. This assumes that latitudinal patterns of insolation remained consistent; however, extreme shifts in Martian obliquity can fundamentally alter these drivers (e.g., Kreslavsky, 2006), potentially leading to regionally systematic orientations that diverge from terrestrial directional expectations. The following section reviews how these variables have been explored in global VFF studies.

7.1.2. Previous VFF slope and aspect studies

The influence of slope and aspect on the distribution and preservation of Martian VFFs is well established, particularly for GLFs (e.g., Milliken et al., 2003; Souness et al., 2012; Brough et al., 2019), where orientation strongly modulates solar radiation receipt and hence ice stability. These studies demonstrate that aspect-dependent energy balance can exert a first-order control on where ice-related landforms can form and persist at the Martian surface. By contrast, crater-hosted ice bodies—and CCFs in particular—have not been examined in the same way. The only global study to include CCFs (Dickson et al., 2012) did not analyse surface slopes or aspects directly but instead inferred flow direction from surface lineations and lobe morphologies and treated CCFs together with other crater-confined ice deposits. As a result, CCF aspects were characterised primarily as concentric, and thus any potential climatic or glaciological signal encoded in their actual surface slopes and aspects remains unexplored. This chapter therefore addresses a clear gap: it presents the first global, CCF-specific analysis of surface slope and aspect, treating these quantities as potential recorders of CCF formation and/or modification processes, rather than inferring behaviour indirectly from surface textures or flow indicators.

In the first global inventory of GLFs, Souness et al. (2012) found that a pronounced poleward preference with a minor, eastward skew is evident in both hemispheres (Figure 7.3). In the northern hemisphere, GLFs aspects display a broad spread of orientations, yet still exhibit an overall bias toward pole-facing aspects, with a secondary eastward bias. In contrast, the poleward preference is much stronger in the southern hemisphere, where GLF aspects are tightly clustered between southwest and southeast. Souness et al. (2012) interpreted the poleward biases as resulting from insolation differences, much like terrestrial glaciers, with pole-facing alcoves receiving less solar radiation and therefore offering more favourable conditions for ice accumulation and/or preservation. However, they did not propose a mechanism or definitive explanation for the eastward bias, nor did they examine whether GLF aspects varied geographically or with latitude. Subsequent studies of GLF aspect (e.g., Brough et al., 2019) have likewise not resolved the cause of this bias, nor explained the difference in orientation distributions between the hemispheres.

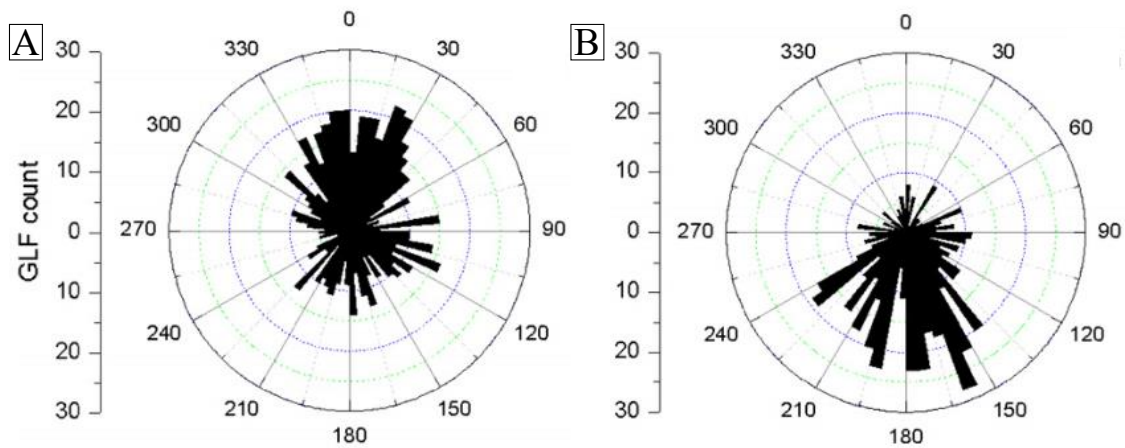


Figure 7.3. GLF rose plots of aspect orientation. (A) Northern hemisphere GLFs, (B) southern hemisphere. GLFs in either hemisphere show preference toward a polar aspect alongside a secondary easterly bias. Redrawn and modified from Souness et al., 2012.

The finding of poleward bias was complicated by a targeted analysis of GLFs by Brough et al. (2019), who focused on a subset of 436 GLFs which display strong evidence of having undergone substantial mass loss and recession (termed as recessional GLFs). These regression GLFs did not display an equatorward-facing bias, which would be expected if orientation had strongly governed ablation. Instead, a persistent poleward preference was observed in both hemispheres, though the distribution of these orientations was significantly more diffuse in the south than in the north (Figure 7.4). Crucially, the persistence of this poleward bias—even in features undergoing recession—suggests that the preservation signal is robust across different feature subsets. Brough et al. (2019) proposed that these patterns reflect modification under a high obliquity regime ($\geq 45^\circ$), under which, pole-facing slopes in the mid-latitudes receive higher peak summer insolation than equator-facing slopes (Costard et al., 2002). This inverted energy balance means that the protective nature of poleward aspects is lost during

these periods, explaining why recessional features might record a more complex, diffuse orientation history than features formed under lower obliquity.

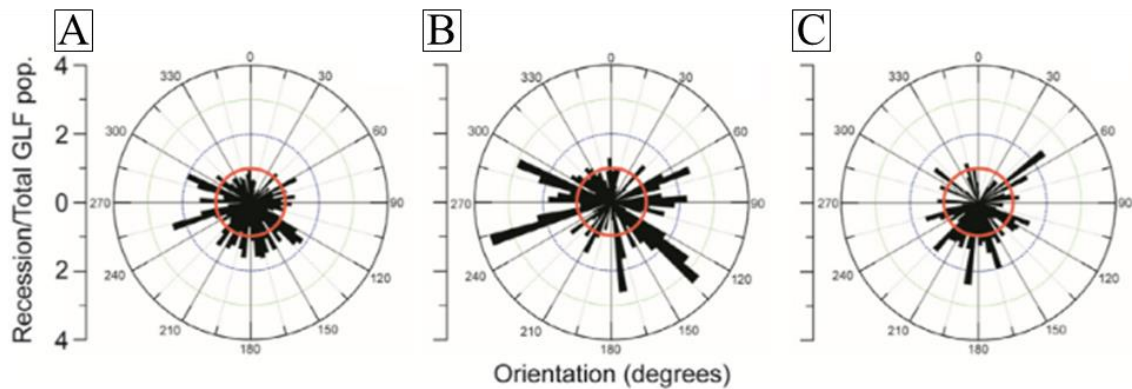


Figure 7.4. Ratio plots of normalised concentration of recessional GLFs relative to the normalised concentration of the total GLF population. (A) global comparison, (B) northern hemisphere and (C) southern hemisphere. Redrawn and modified from Brough et al. (2014)

Another study, which did not investigate surface slopes but instead inferred aspect from the orientation of surface lineations in mid-latitude crater-confined ice features, including CCFs, was that of Dickson et al. (2012). For each feature in their sample, they inferred a flow direction or flow-direction type from the observable pattern of surface ridges and lobe morphologies (using CTX images), rather than measuring the surface slope and aspect directly. Their global study combined three morphologically distinct populations: (1) CCFs, which typically cover the entire crater floor, (2) low-definition CCFs (e.g., Levy et al., 2010; Chapter 3), and (3) lobate deposits extending downslope from pole-facing crater walls, morphologically similar to crater-confined LDAs (Sinha and Vijayan, 2017). Their analysis identified a bimodal latitudinal pattern in flow direction, with pole-facing flows dominating equatorward of 45° N/S, and more concentric, non-directional orientations between 45° and 60° , in both hemispheres (Figure 7.5).

This distribution pattern deviates from the dominant poleward and eastward biases typically reported in studies of smaller VFFs (GLFs; e.g., Milliken et al., 2003; Souness et al., 2012). This divergence likely stems from feature scale, as GLFs are significantly more sensitive to local topographic shading than the macro-topographic gradients of bulk CCFs. Instead, the bimodal latitudinal pattern in flow direction aligns more closely with those observed in crater wall gullies and slope-confined flows (e.g., Berman et al., 2005; 2009; Noblet et al., 2024), where the preferred aspect of hillslopes is latitude-dependent, exhibiting a shift from pole-facing to equator-facing orientations at approximately 40° – 45° N/S. In addition to this latitudinal pattern, Dickson et al. (2012) observed a regional consistency in flow pattern orientation between CCFs in neighbouring craters, suggesting a degree of local climatic or topographic control.

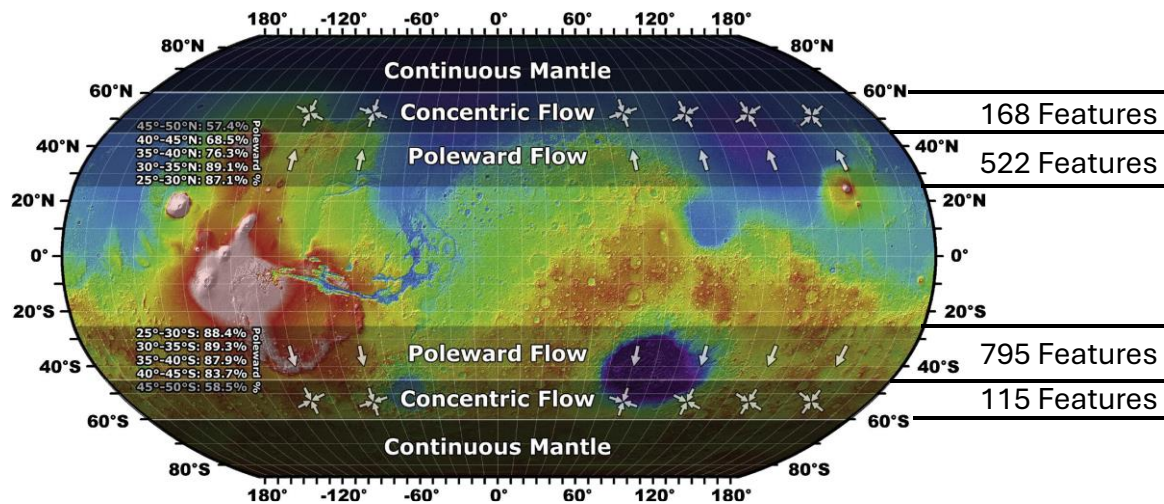


Figure 7.5. Bimodal latitudinal variation in flow orientation of crater-hosted ice-related features identified by Dickson et al. (2012). Flow orientations inferred from surface lineations show a bimodal distribution with latitude. Pole-facing flow dominates between $\sim 25^\circ$ and 45° in both hemispheres, whereas concentric flow orientations become prevalent between $\sim 45^\circ$ and 60° . The mapped population comprises 910 features in the Southern Hemisphere and 690 in the Northern Hemisphere, with the majority of features occurring between 30° and 45° latitude (North: 75.8%; South: 82.3%). Figure redrawn and modified from Dickson et al. (2012).

This latitudinal pattern was attributed by Dickson et al. (2012) to insolation forcing and slope-scale microclimatic effects. Under intermediate obliquity ($\sim 35^\circ$), pole-facing interior crater walls receive less daily insolation than equator-facing slopes at all latitudes, but this insolation asymmetry becomes a dominant influence equatorward of $\sim 45^\circ$. The disparity in insolation enhances accumulation on colder, pole-facing slopes, with increased ablation on the warmer, equator-facing slopes (Dickson et al., 2012). This relationship is, however, strongly modulated by obliquity and season. Energy-balance modelling by Kreslavsky et al. (2006) demonstrates that while equator-facing slopes receive more annual-mean insolation under moderate obliquity, pole-facing slopes at mid-latitudes become the most strongly illuminated surfaces during summer at high obliquity ($>35\text{--}40^\circ$), with slope–insolation contrasts reversing in sign and magnitude depending on latitude and season (Figure 7.6).

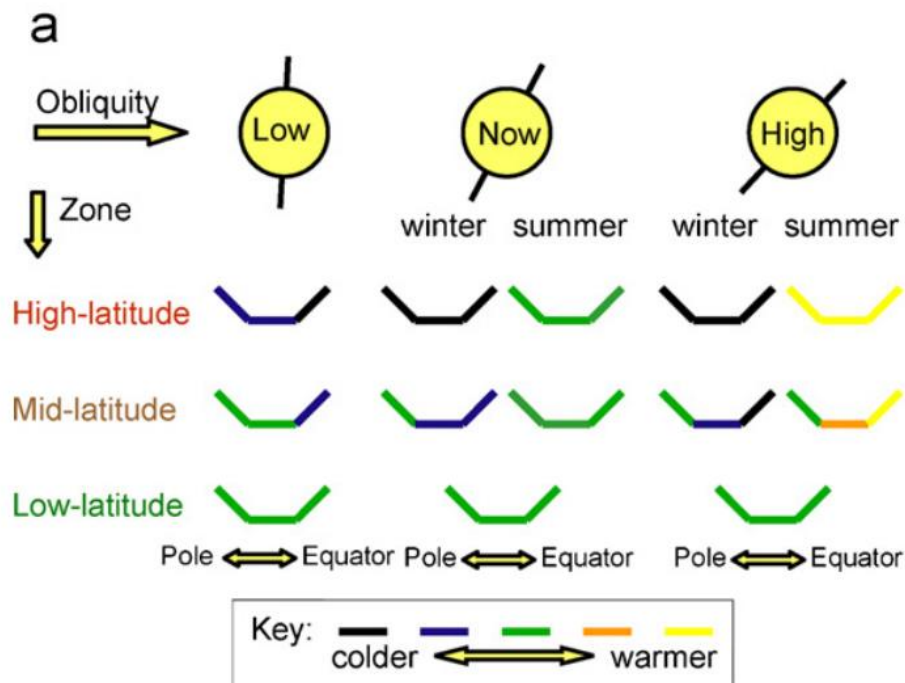


Figure 7.6. Schematic illustration of the dependence of insolation on obliquity, latitude, and slope orientation. Conceptual diagram showing the first-order controls of orbital obliquity and surface slope orientation on received solar radiation at different latitudes. In the mid-latitudes, pole-facing and equator-facing slopes experience systematically different seasonal insolation regimes, leading to persistent contrasts in surface temperature and energy balance over the course of the year. This qualitative framework highlights the tendency for pole-facing slopes to remain colder than equator-facing slopes under typical Martian obliquity states. Redrawn from Kreslavsky et al. (2006).

Regarding the prevalence of concentric flow orientation for CCFs at higher latitudes, Dickson et al. (2012) proposed two explanations: (1) under the intermediate obliquity usually assumed for mid-latitude glaciation to occur (e.g. modelling by Madeleine et al., 2009), insolation generally decreases poleward (Ward, 1974). Consequently, at higher latitudes, slope orientation becomes less of a determining factor for ice accumulation, allowing CCFs to accumulate regardless of slope orientation. Alternatively (2) the observed flow patterns indicate a more extensive regional ice-sheet accumulation of the past, independent of local topographic setting, leading to the formation of concentric distribution of aspects. In this scenario, some CCFs poleward of 45° may not originate from in-crater alcove accumulation (as latter modelled by Weitz et al., 2018) but instead reflect the intrusion of ice from broader regional ice sheets flowing into crater interiors through rim breaches. Once inside, steep interior walls would shed debris onto the ice surface, preserving evidence of flow (e.g., Dickson et al., 2012). At face value, such an origin might be expected to produce preferentially unidirectional flow signatures reflecting the regional ice-flow direction. Dickson et al. (2012), however, argued that concentricity is maintained because the crater rim acts as a circumferential (360°) source of debris: this continuous armouring, combined with topographic confinement within the crater bowl, forces ice-surface flow to

converge toward the centre, thereby overprinting any initial regional flow direction with a concentric surface morphology. However, because Dickson et al. (2012) aggregated CCFs with other crater-hosted ice-flow features (Figure 7.1), it remains unclear whether there are aspect dependencies within the population of concentric CCFs themselves, particularly given that such morphologies extend equatorward to $\sim 25^\circ$ latitude in the present dataset (Figure 3.9). This chapter addresses this by examining global patterns in CCF surface slope and aspect independently of other crater ice features.

7.2. Data and methods

7.2.1. Obtaining CCF slope and aspect

Throughout this chapter, I quantify the feature-scale slope and aspect of each CCF surface. Here, slope refers to the rise-over-run gradient of a single plane fitted to all elevation points within the mapped CCF boundary. This metric therefore describes the first-order inclination of the CCF surface as a whole, rather than the mean of local pixel-scale slopes associated with tracelines, lobes, or surface roughness. The associated aspect represents the downslope direction of the fitted plane. For each CCF, slope and aspect values were derived from the Digital Terrain Model (DTM) data for each CCF, which was extracted from gridded MOLA data at a horizontal resolution of 463 m and previously described in Section 5.3 (encompassing the CCF and its host crater). To isolate the CCF surface, the DTM was clipped to its CCF outline.

Specifically in MATLAB, linear regression was used to fit the plane equation:

$$z = Ax + By + C, \quad (7.1)$$

to the DTM, where x and y are grid (projected) coordinates and k is the scale factor (as detailed in Section 4.2.1). Then, the CCF's mean slope (defined as rise over run, dimensionless) is given by

$$Slope = k \cdot \sqrt{A^2 + B^2}, \quad (7.2)$$

and its aspect measured in degrees clockwise from the y -axis (working the range -180° to 180°) is given by:

$$\theta = \frac{180}{\pi} \cdot \text{atan2}(-A, -B). \quad (7.3).$$

Finally, a simple rotation to account for the angle between the y -axis (grid north) and geographical or true north on the LCC plane, based on the longitude of the CCF's centroid (as used in Chapter 5), gives the CCF's aspect measured from geographical north.

Since the method approximates CCF surfaces as a plane, the mean slope and aspect results may be sensitive to elevation noise or artifacts within the DTM data, which can skew the plane parameters,

particularly in areas with uneven surfaces or minimal data coverage, such as on smaller CCFs. The amount of uncertainty on the slope estimate therefore matters; I address it further below in Sec. 7.2.4.

7.2.2. Validating the slope and aspect results

To verify the slopes and aspects calculated in MATLAB, the coefficients of Eq. (7.1) were first independently calculated in ArcGIS Pro, by using its ‘Linear Trend’ tool to perform polynomial regression fitting to the point data of each CCF’s (clipped) DTM. I found that the MATLAB and ArcGIS Pro fitting gave near-identical coefficients that differed only by 10^{-14} to 10^{-17} . Thus, the MATLAB-fitted coefficients are valid.

Next, I tested whether my MATLAB algorithm (last section) and ArcGIS Pro — specifically its ‘Slope’ and ‘Aspect’ tools — produced identical CCF slope and aspect values from those coefficients. The ArcGIS Pro tools apply a 3-by-3 cell moving window to process input data and gives the user a choice of a planar method (assuming a two-dimensional surface) or a supposedly more accurate, geodesic method (accounting for the planet’s curvature). Given that CCF slopes are low, precision is critical, so the latter method was chosen. The ArcGIS Pro ‘Aspect tool’ automatically projected values to correct any distortion introduced by the non-conformal output coordinate system environment, ensuring accurate orientation measurements even in a non-planar projection.

The test outcome (Figure 7.7) reveals non-negligible differences between CCF surface slopes and aspects from MATLAB and ArcGIS Pro. Although slope comparison yielded ratios of 0.87 and 0.89 for the Northern and Southern hemispheres, respectively, with R^2 values of 0.996 and 0.998, individual slope values disagree by as much as 0.01-0.02. Manual validation using rise-over-run calculations from elevation profiles on MOLA data, aligned with CCF outlines, confirmed the MATLAB-derived slopes as accurate. Residual analysis showed that discrepancies decrease with crater diameter, with ArcPro generally reporting steeper slopes.

Why such large, unexpected differences in the results? A likely culprit is the moving 3×3 cell window algorithm in ArcPro, which omits regions with fewer than seven valid cells. In fact, Figure. 7.7 shows that CCF slopes from ArcPro almost always exceed those from my MATLAB algorithm (except at 1 data point); this likely arises because the ArcPro slope tool averages the absolute slope values from different 3x3 patches on a CCF and therefore cannot yield the CCF mean slope being sought; e.g. the narrow ridge of an arête has zero mean slope according to my definition, but ArcPro will calculate a positive slope for it. Aspect results for individual CCFs also show significant differences, likely for the same reason: the ArcPro algorithm’s local 3x3 patch averaging does not equate to computing a single, mean surface normal direction for the entire CCF. Additionally, ArcPro failed to process 313 CCFs with more than 25 valid cells, for reasons not fully understood. Given these issues, and that the ArcPro algorithm is ultimately not fully transparent, MATLAB-derived slope and aspect values are used in this chapter.

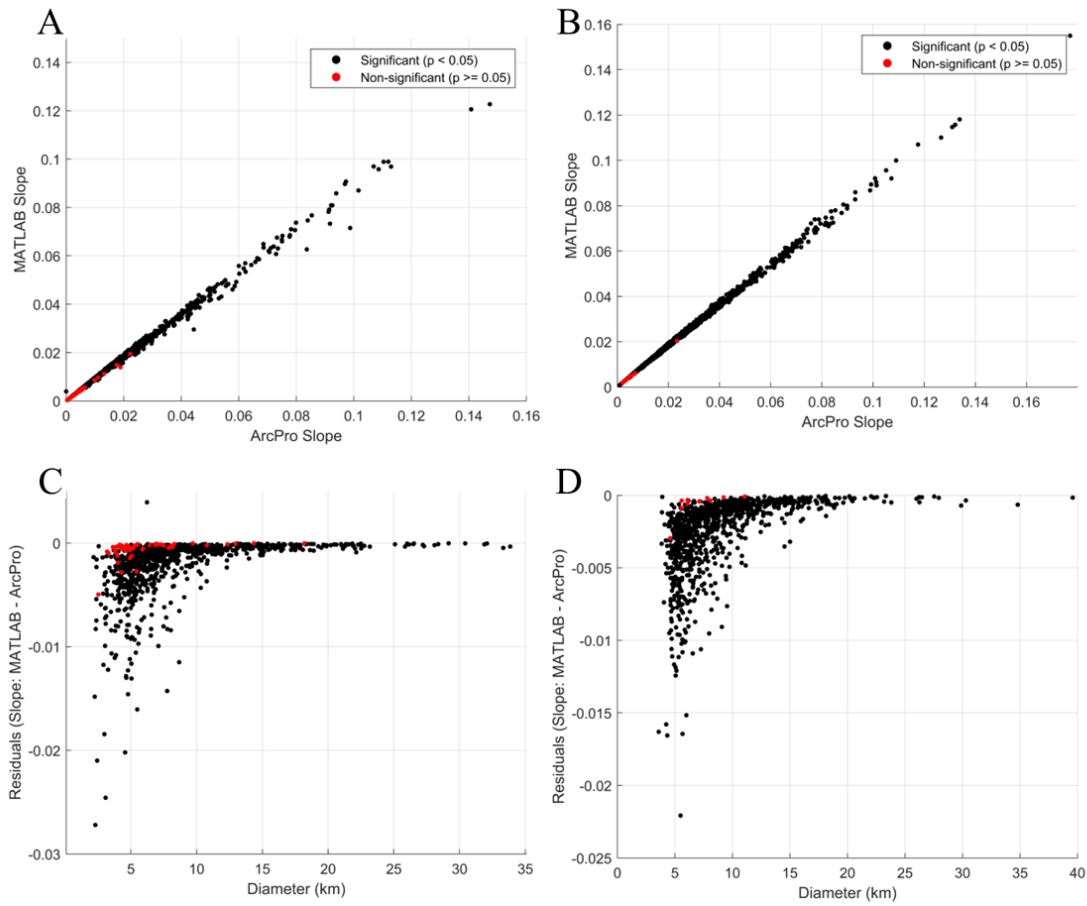


Figure 7.7. Comparison of CCF surface slope results derived from MATLAB and ArcGIS Pro. (A–B) Scatter plots comparing ArcGIS Pro-derived CCF slopes (x-axis) to MATLAB-derived slopes (y-axis) for the Northern (A) and Southern (B) Hemispheres. Data points with statistically significant plane fits ($P \leq 0.05$) are shown in black; non-significant cases ($P > 0.05$) are shown in red. Slope ratios are 0.87 (NH) and 0.89 (SH), with R^2 values of 0.996 and 0.998, respectively. (C–D) Residual slope differences (MATLAB minus ArcPro) plotted against host crater diameter for the Northern (C) and Southern (D) Hemispheres. Residuals decrease with increasing crater diameter, with ArcPro generally reporting slightly steeper slopes across all size bins.

7.2.3. Slope standard error

As Figure 7.7 indicates, CCF surfaces tend to have very shallow slope (~ 0.12 or less). A first question is whether one can regard a given CCF as sloping, as opposed to horizontal. Evaluating this requires knowing the uncertainty/error on its slope estimate. A CCF whose absolute slope is below uncertainty may be either horizontal or sloping; this indeterminacy means that it cannot be analysed. Also, the aspect calculated for a CCF will not be meaningful if the CCF is effectively horizontal (the surface normal is then vertical), and so the aspect is valid only if its estimated slope exceeds the slope uncertainty.

The slope standard error (SSE) is used for the determination. MATLAB gives the SSE alongside the slope coefficient, but only in its single-variable regression function, not in multivariable regression fits to 3D point data. To find the SSE of each CCF's slope, in MATLAB I re-fitted its DTM data – in a

second pass on rotated coordinates. Using the CCF's aspect angle θ , calculated in the first pass, the plane of Eq. (7.1) was rotated by θ so that the new x-axis aligns with the aspectual direction. Besides finding the SSE, I checked that the slope calculated in the second pass was equal that in the first pass. Both passes also gave a p -value for the slope, whose use is described in the next section.

7.2.4. Refinement of CCF dataset

To avoid skewing the analysis, the sample of CCFs that are non-horizontal (statistically speaking) based on the SSE criterion needs to be identified and retained, but small CCFs with too few elevation data also need to be excluded from study. Larger CCFs are intersected by numerous MOLA laser shots that, reliably capture their surface geometry. Mouginis-Mark et al. (2018) found that craters with $D \leq 5$ km typically have fewer than 10 lines of raw MOLA laser shots, while larger craters have better coverage e.g., more than 15 profiles at $D = 15$ –20 km. They recommend a minimum crater D of 5 km for reliable geometric analysis from MOLA data. However, Tornabene et al. (2018) noted significant deviations between MOLA PEDR and HRSC DTMs for craters with $D < 10$ km, questioning the reliability of MOLA for smaller features. Excluding CCFs with $D < 10$ km reduces my sample to 550 CCFs before further filtering. While this level of restriction may enhance data reliability, it would eliminate a large portion of the inventory and risk discarding potentially valuable climatic information from the CCFs in smaller host craters.

To balance coverage with data quality, I decided on $D = 5$ km as the threshold, consistent with the lower-bound recommendations of Mouginis-Mark et al. (2018). However, host-crater diameter is an imperfect proxy for the quality of the geometric data, as the analysable CCF surface is often significantly smaller than the crater floor—particularly where mantling obscures the underlying CCF (as retraced in Section 7.2.1). Because this analysis utilises gridded MOLA data (463 m/pixel) rather than raw PEDR profiles, a secondary filter was applied to exclude any CCF represented by fewer than 25 grid cells. This criterion effectively imposes a minimum surface area threshold (~ 5.4 km²), ensuring that the visible CCF deposit retains sufficient spatial extent to yield a statistically robust mean slope, regardless of D . Importantly, this is a gridded DEM cell-count criterion rather than a direct measure of raw MOLA shot density. The 25-cell threshold ensures that each CCF surface is represented by enough raster cells for plane fitting, but it does not confirm that those cells are equally constrained by original MOLA laser-altimetry points. MOLA point spacing varies with latitude and local track distribution, so interpolation within the gridded DEM may still affect individual slope and aspect estimates. This limitation is partly mitigated by the mid-latitude focus of the analysis, rather than an equator-to-pole comparison, but it remains most relevant for smaller CCFs where fewer observations constrain the gridded surface.

For this chapter, CCFs were excluded if they failed any of the following criteria: non-statistically significant p -values for the slope, host craters with $D \leq 5$ km, CCF DTMs (post clipping) with fewer than 25 MOLA cells, or slopes not statistically significantly different from zero (absolute slope minus

SSE below 0.001). Figure 7.8 summarises my workflow. As in Chapter 5, data for D derive from the crater diameters of the Robbins and Hynek (2012) global crater database.

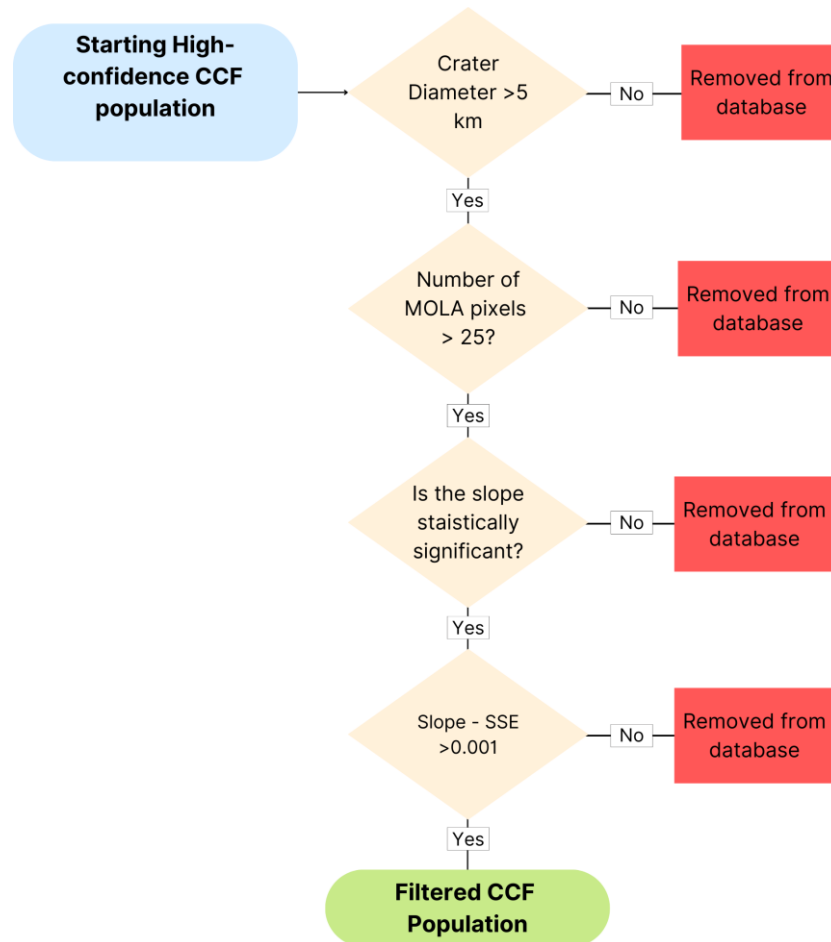


Figure 7.8. Filtering process for identifying CCFs for reliable surface slope and aspect analyses.

Flowchart illustrating the criteria used to exclude CCFs from the analysis. CCFs were removed if they: (i) were hosted in craters smaller than 5 km in diameter, (ii) had fewer than 25 MOLA data points, (iii) had non-significant slope fits ($p > 0.05$), or (iv) had slopes not statistically distinguishable from zero (slope minus its standard error < 0.001).

The criteria on D , MOLA grid-cell count and p -value excluded 758 CCFs in the northern hemisphere and 534 CCFs in the southern hemisphere, leaving 820 Northern and 906 Southern CCFs before the final SSE criterion. That final criterion further removed 28 CCFs in the North and 4 CCFs in the South, leaving 792 and 902 CCFs in those hemispheres respectively, representing 51% and 62% of the initial CCF inventory in each hemisphere for analysis. Their minimum numbers of MOLA grid cells are 29 and 26 respectively. While this represents a substantial reduction from the initial sample (Chapter 3), precision was prioritised due to the low slope values under investigation.

7.2.5. Analysis of aspect clustering

Statistical tests can be used to determine if CCF surface orientations show a statistically significant directional clustering or uniformity, which can be visualised using a rose diagram. CCF aspect values in both hemispheres were tested for conformity to a von Mises distribution, the circular analogue of the normal distribution (Landler et al., 2018), with both hemispheres found to be multimodal and non-conforming. Given that the commonly used Rayleigh test, which assesses the clustering tendency of orientations, exhibits low power to detect multimodal deviations from uniformity, particularly in cases where multiple concentrations of data exist around the circle (Batschelet, 1981; Landler et al., 2018), this method was deemed inappropriate for the current analysis.

Landler et al. (2018) analysed different circular statistical methods, testing their performance under different power simulations for various multimodal distributions, finding that there is no specific multimodal scenario in which any alternative test demonstrates substantially better efficacy than the Hermans-Rasson test. Thus, for the purposes of this chapter, the Hermans-Rasson test was employed to provide a quantitative measure of clustering significance, allowing for an objective identification of directional patterns across the CCF dataset, with no assumptions on the symmetry of the distribution.

In this test, the null hypothesis indicates that the population is uniformly distributed around the circle with no significant variation, while the alternative hypothesis suggests that the population is *not* uniformly distributed, indicating the presence of at least one significant cluster. A *p*-value below a threshold of 0.001 would provide sufficient evidence to reject the null hypothesis, confirming that clustering of aspect orientations is statistically significant within the dataset.

7.3. Results and Analysis

This section describes and examines the slopes and aspect results, assessing first how many CCFs have non-zero slopes. The analysis explores how the slopes vary with latitude, crater diameter, and apparent depth, in order to identify potential controls on CCF slope development. Subsequent sections focus on the aspect distribution, hemispheric and regional biases in that distribution, their relationship to host crater diameter or broader climatic gradients.

7.3.1. CCF surface slopes

Of the 1726 CCFs measured, 1694 possess statistically non-zero slope (refinement step see Figure 7.8). Only 32 CCFs lack statistically non-zero slope — 28 are in the North Hemisphere and 4 in the South Hemisphere. Thus, a first result is that despite their shallowness, many CCFs are sloping. Their surface slopes are gentle. Maximum slope does not exceed 0.16 ($\sim 9^\circ$), with median slope values of 0.0121 in the northern hemisphere ($\sim 0.7^\circ$) and 0.0191 in the southern hemisphere ($\sim 1.1^\circ$). Upper quartile thresholds remain below 0.0206 in the North Hemisphere and 0.0324 in the South Hemisphere. Shallow surface gradients then appear to be a defining property of CCFs globally.

7.3.1.1. Slope and latitude

Slope decreases with (absolute) latitude in each hemisphere (Figure 7.9), with steeper slopes concentrated between 25° to 40° latitude. Beyond $\sim\pm 45^\circ\text{N/S}$, slopes rarely exceed 0.025. Spearman's rank correlation reveals a weak but significant negative correlation between slope and latitude ($\rho = -0.30$, $p < 0.001$ in the northern hemisphere; $\rho = -0.26$, $p < 0.001$ in the southern hemisphere). Irrespective of latitude, most CCFs have slopes below ~ 0.03 , and steeper CCFs are infrequent (Figure 7.9).

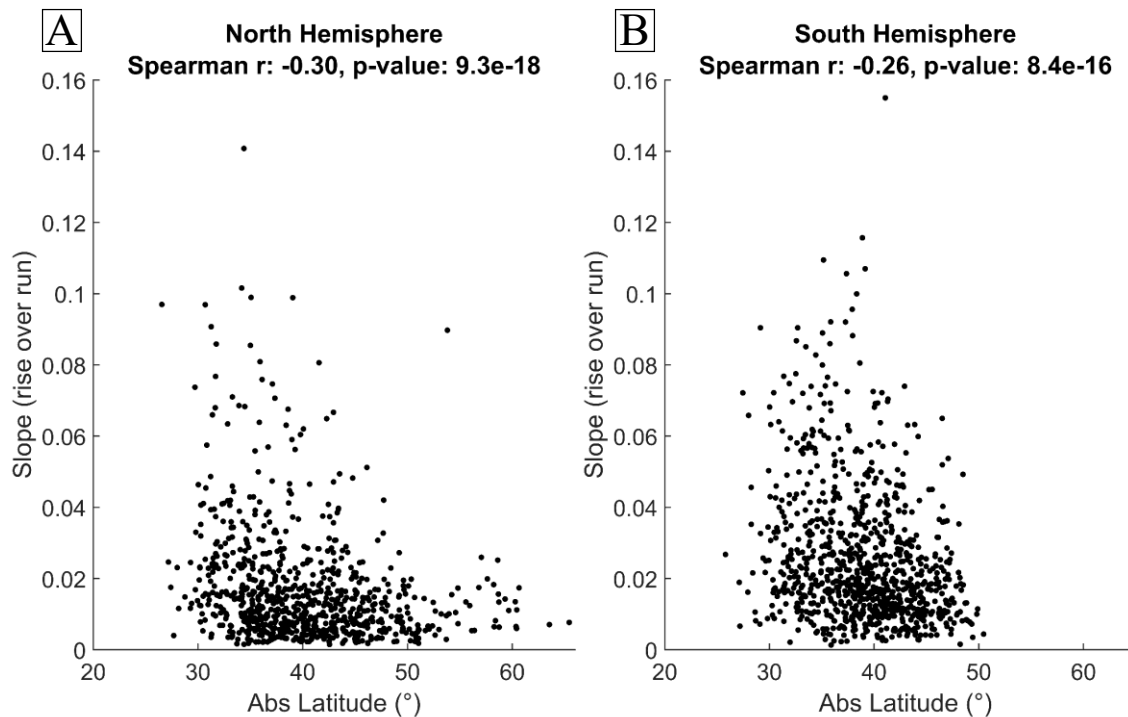


Figure 7.9. Scatterplots of CCF Slope vs. Latitude for the Northern and Southern Hemispheres. (a) Northern Hemisphere CCF population; (b) Southern Hemisphere CCF population. Both hemispheres display a weak but statistically significant negative correlation between slope and latitude, with slopes decreasing poleward and most CCFs concentrated at lower slope values (~ 0 – 0.03).

This poleward decline in slope aligns with other latitudinal trends found in this Thesis, notably decreasing apparent depth (Chapter 5) and increasing estimated CCF volume (Chapter 6) poleward. Together, these findings indicate that CCFs accumulation and/or preservation increases with latitude in both hemispheres. The latitude-dependent reduction in slope may reflect increased mass balance during CCF active glaciation, or enhanced preservation due to lower ablation rates during post glacial periods.

7.3.1.2. Slope and host crater diameter

In both hemispheres, slope exhibits a negative relationship with D (Figure 7.10), supported by weak, statistically significant negative correlations (North: $\rho = -0.31$, $p < 0.001$; South: $\rho = -0.26$, $p < 0.001$).

The data are characterised by wedge-shaped distributions where the upper envelope of maximum CCF slope decreases continuously with increasing D . The steepest slopes are concentrated in smaller craters ($D < 10$ km), where CCF slopes range from very low values (~ 0.002) to ~ 0.14 . Steeper slopes (e.g., > 0.05) are almost exclusively restricted to craters with $D \leq 15$ km in diameter. In the largest craters ($D > 15$ km), this declining upper flank has already constrained slope values to primarily below 0.04 and typically < 0.02 . The plotted data in the two hemispheres have similar steeply declining upper flanks, indicating that D strongly limits the maximum CCF surface slope.

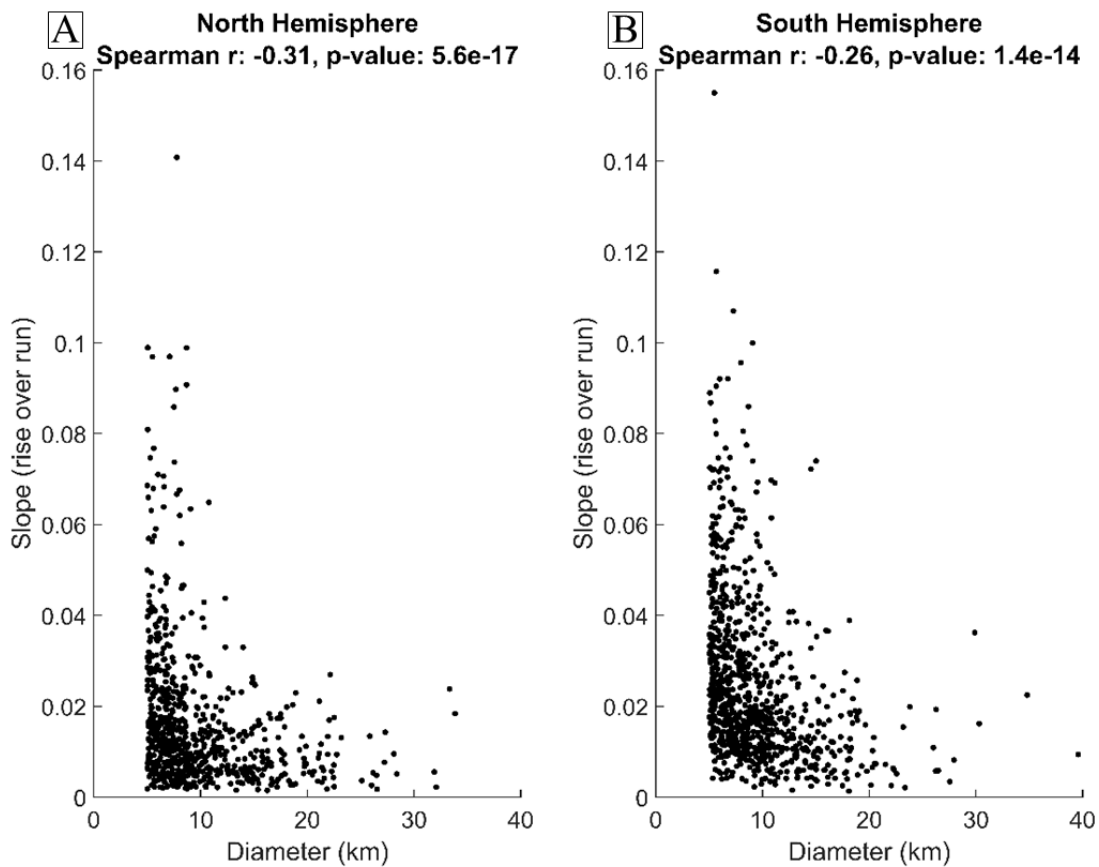


Figure 7.10. Scatterplots of CCF Slope vs. Diameter for the Northern and Southern Hemispheres. (A) Northern Hemisphere CCF population; (B) Southern Hemisphere CCF population. Both hemispheres display a weak but statistically significant negative correlation between slope and latitude, with slopes decreasing with diameter

7.3.1.3. Slope and d_a : origin of maximum slope

For a given host crater with a diameter D , the maximum CCF slope can be deduced from host crater geometry. Viewing the system in vertical section (Figure 7.11), a CCF with d_a (mean crater rim elevation – mean CCF surface elevation) has a mean surface slope up to $\approx 2d_a/D$ (Figure 7.11), not more (because one edge of the CCF would have to lie above the crater rim in such case). Thus, a CCF with low d_a , i.e.

with a mean surface near the mean rim, must have low mean slope. This theoretical bound ignores the slight discrepancy between the CCF surface diameter and the host crater diameter, as well as the three-dimensional intersection between the tilted CCF surface and crater walls. Additionally, I found a poleward-decreasing upper limit to d_a in Chapter 5, for craters of similar D ; this implies an upper limit on CCF slope that decreases poleward in both hemispheres.

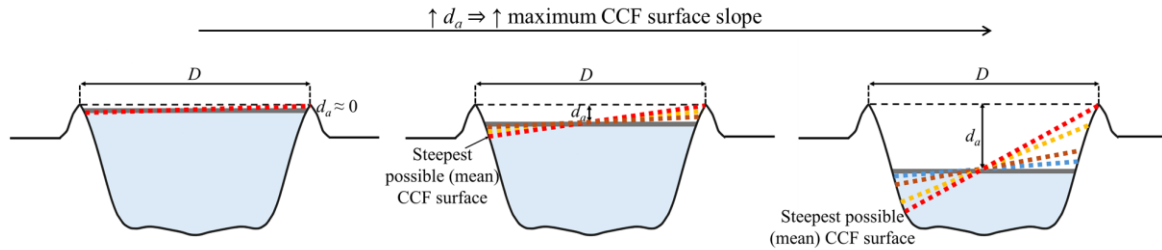


Figure 7.11. Schematic of the upper limit of CCF slope. Three craters with increasing d_a illustrate that both the maximum slope and the allowable range of slope values increase with d_a . The thick grey line marks the mean CCF surface, dotted lines indicate possible surface profiles, and the red line shows the steepest (mean) CCF surface, approximated as $\frac{2 \cdot d_a}{n}$.

To test the idea, Figures 7.12 and 7.13 plot CCF surface slope against d_a , binned by D , for both hemispheres. The red line in each panel plots the theoretical maximum slope, $\frac{2 \cdot d_a}{D}$, where D is taken as each bin's mid diameter. As expected, CCFs with lower surface slopes generally correspond to lower d_a , by nature of being near rim-full leaving little room for significant slope development. CCFs within larger host craters ($D > 11$ km), exhibit consistently low slopes, rarely exceeding 0.05 irrespective of d_a .

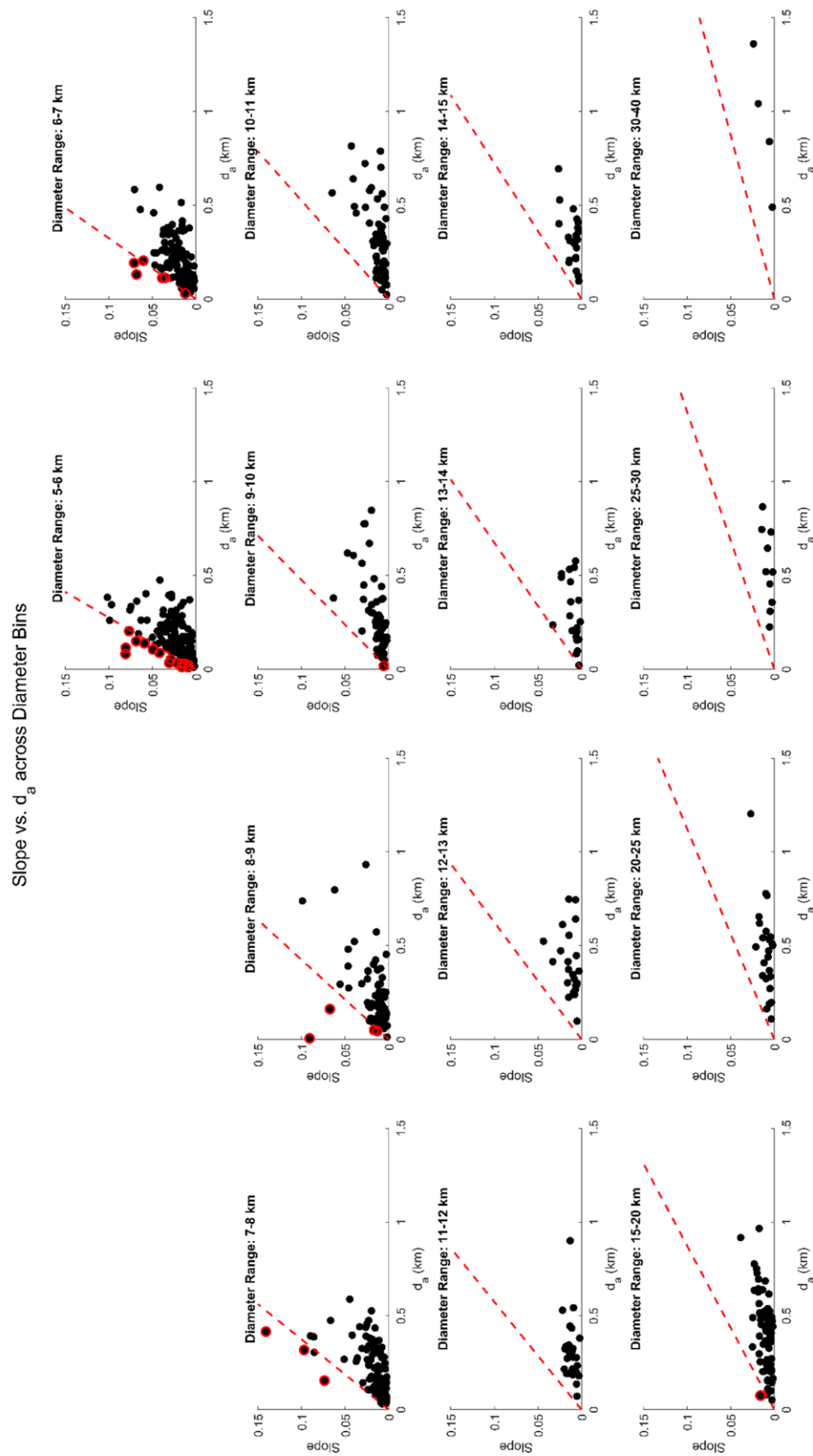


Figure 7.12. Scatterplots of CCF Slope vs. Apparent Depth Across Diameter Bins (Northern Hemisphere). Each panel represents a binned diameter range, with a red dashed line indicating the approximate maximum slope at the midpoint of each bin. CCFs outlined in red exceed their respective slope thresholds.

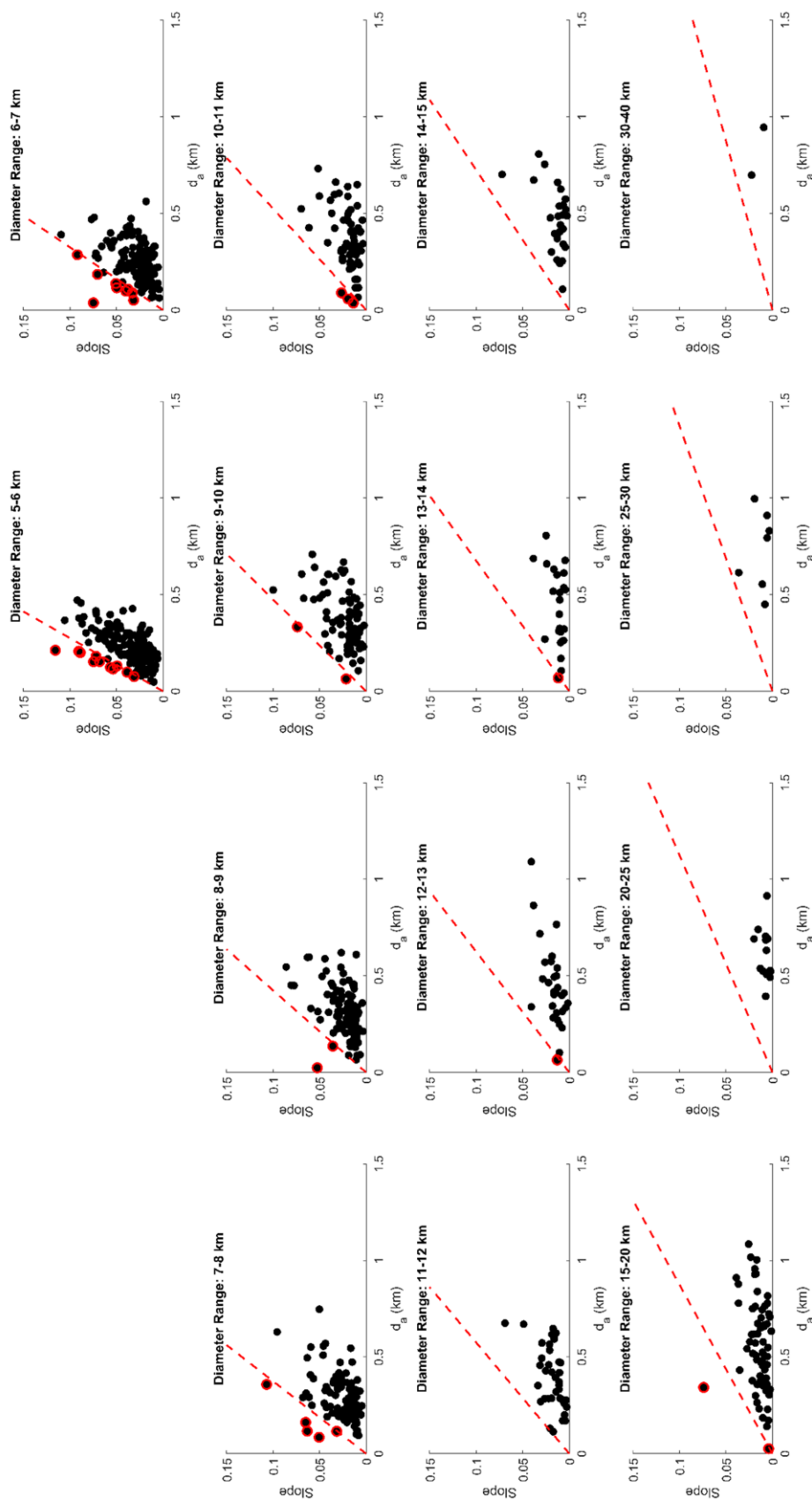


Figure 7.13. Scatterplots of CCF Slope vs. Apparent Depth Across Diameter Bins (Southern Hemisphere). Each panel represents a binned diameter range, with a red dashed line indicating the approximate maximum slope at the midpoint of each bin. CCFs outlined in red exceed their respective slope thresholds.

Figures 7.12 and 7.13 thus confirm the operation of an upper bound (in slope) that not only increases with d_a , but also decreases with D . This empirical bound is well described by the theoretical bound. Most CCFs fall significantly below their theoretical maximum slope, although 30 Northern and 36 Southern CCFs exceed it (highlighted red, Figures 7.12 and 7.13). Images analysis reveals that ~75% of the outlier CCFs are associated with complex background topographic features, most commonly being situated within larger craters (Figure 7.14A–C), which likely contributes to the development of steep CCF surfaces—perhaps limiting dust availability for the overburden or sublimation lag. Few CCFs (~10%) are linear or divergent CCFs (Figure 7.14D), but more curiously, some are classic CCF (Figure 7.14E, F).

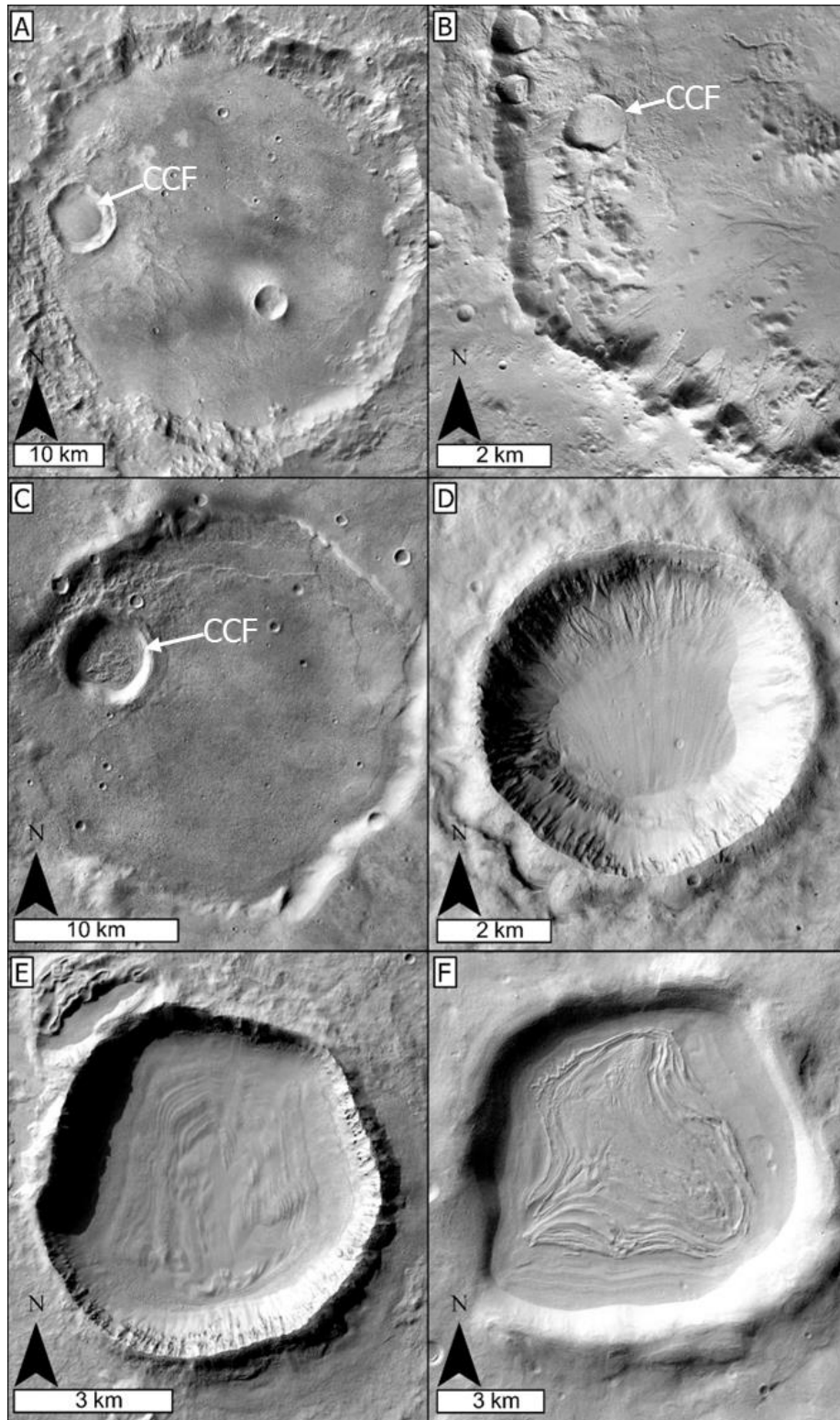


Figure 7.14. CTX Images of Steepest Slope CCFs and Influencing Factors. Steepest slope CCFs are associated with complex background topography (A, B), larger host craters (C, D), highly eroded rims (E, F). (A) 39.09° S, 142.18° W; CTX images: G17_024996_1406 and F08_038974_1405. (B) 37.68° S, 159.25° W; CTX images: D08_030324_1416, F14_041269_1417, P06_003319_1414. (C) 32.78° S, 22.65° E; CTX images: B02_010406_1472 and P07_003787_1467. (D) 37.25° S, 151.97° E; CTX image: B11_014067_1426. (E) 35.82° S, 28.66° E; CTX image: G18_025227_1437. (F) 32.66° S, 52.03° E; CTX image: F21_043872_1480

7.3.1.4. CCF sub-grouping by slope

For the ensuing analysis, including CCF aspects, it is useful to address different sub-groups of CCFs within the admitted set of CCFs (those with statistically-significant non-zero slopes). I define these groups by using quartiles on the slope frequency histograms. In the northern hemisphere, slope values corresponding to the lower quartile, median, and upper quartile are 0.0066, 0.0121, and 0.0206, respectively, for the southern hemisphere, 0.0116, 0.0191, and 0.0324. The lower quartile and median values are shown in Fig. 7.15.

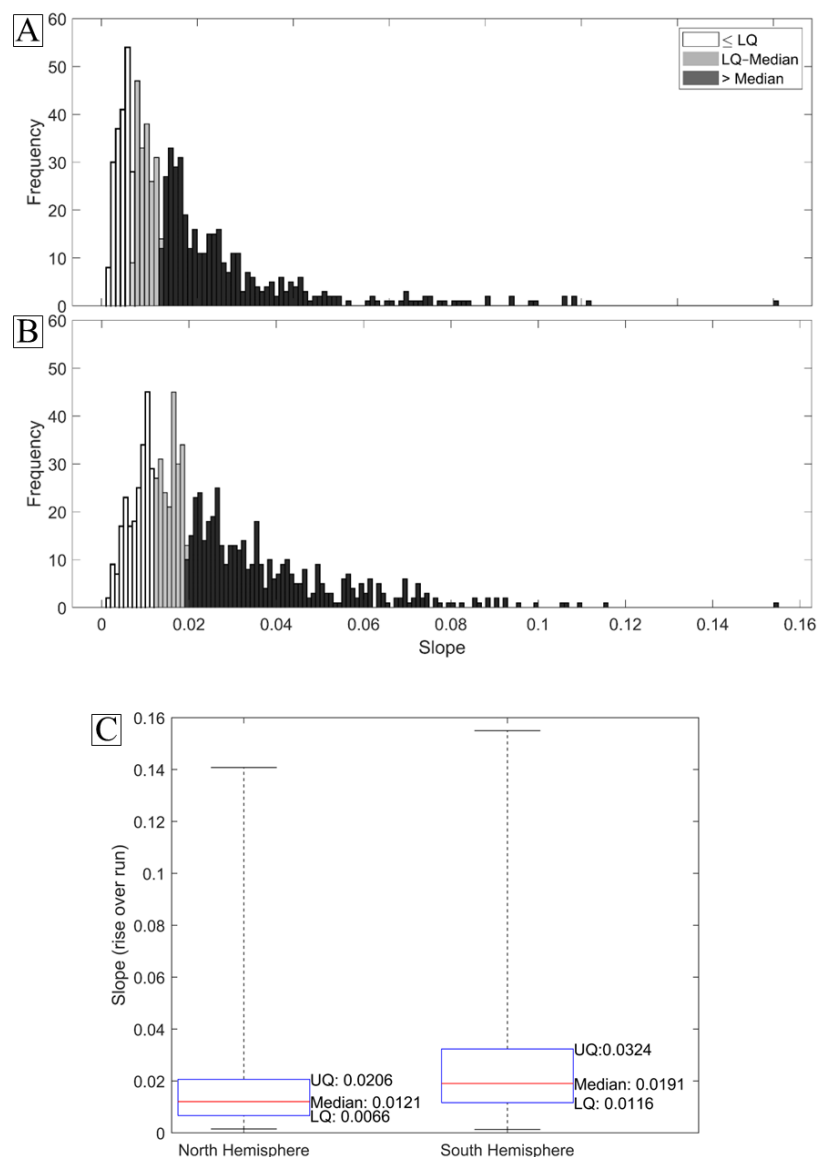


Figure 7.15. Distribution of CCF surface slopes in the northern and southern hemispheres. (A, B) Histograms showing the frequency of CCF slope magnitudes for the Northern (A) and Southern (B) Hemispheres. Bars are shaded by slope magnitude: white (slope $<$ lower quartile), light grey (between lower quartile and median), and dark grey (slope $>$ median). (C) Box plot summarising slope distributions. Boxes show the interquartile range (IQR), red lines indicate the median, and whiskers represent the full range of values.

For the remainder of this chapter, only CCFs with surface slopes exceeding the lower quartile will be included. This selection totals 594 in the northern hemisphere and 677 CCFs in the southern hemisphere which possess statistically significant slope. In future plots, the full population of 792 CCFs in the northern hemisphere and 902 CCFs in the southern hemisphere will be represented as black outlines with white infill for context but not factored into statistical analysis.

CCFs with slopes above the lower quartile will be shaded in grey, while those with slopes above the median will be shown in black (shown in Figure 7.15). This approach ensures that low-slope CCFs, which could introduce uncertainty into the analysis, are suitably excluded from statistical analysis, providing a more accurate understanding of CCF surface aspect.

7.3.2. CCF aspects

This section examines the aspect results on rose plots. Poleward and easterly biases in CCF aspect are evident in both hemispheres, though with varying strengths (Figure 7.16). In the northern hemisphere, CCF aspects exhibit a broad distribution (Figure 7.16A), with many CCFs oriented between northeast and southeast and alongside two south- and west-facing clusters. Northwest-facing aspects are near absent. The overall northward bias is weak. In contrast, southern hemisphere aspects are more concisely distributed (Figure 7.16B), with a strong and consistent pole-facing (south to southeast) bias.

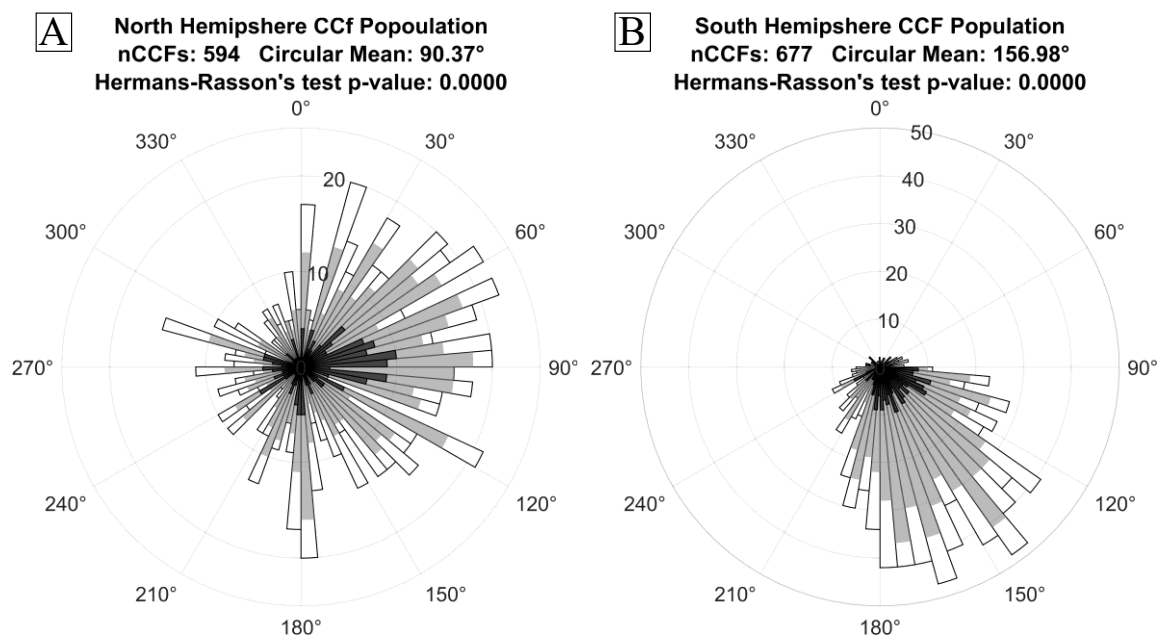


Figure 7.16 Compass-rose plots showing CCF Aspect distribution separated by hemisphere. Orientation is divided into 5° bins, and each concentric ring represents an increase of 10 in frequency. (A) Northern Hemisphere CCF population, with a concentric distribution but strong Northeast to Southeast facing trend and a mean orientation of 90.37°. (B) Southern Hemisphere CCF population, displaying a strong South and Southeast facing bias. Both populations display statistically significant clustering as per the Hermans-Rasson's test.

As shown by the subgroup shading (Sec. 7.3.1.4), CCFs with above-median slopes mirror these biases. In the northern hemisphere, they maintain a strong easterly bias while also retaining distinct west and south-facing clusters. Conversely, in the southern hemisphere, the dominance of south and southeast orientation persists across steeper slope ranges, with a near-total absence of steep equatorward-facing CCFs ($270^\circ - 90^\circ$), but with a notable reduction in southwest orientated steep slopes.

7.3.2.1 Spatial variation in CCF aspect

To investigate where CCF aspects are clustered or dispersed on Mars, a global map was made of them (Figure 7.17), with the direction symbols coloured again by slope quartiles. In the northern hemisphere, aspects appear highly variable, with no clear regional clusters evident. By contrast, southern hemisphere CCFs display preferential clustering, a pattern that is examined in more detail using density heatmaps. To visualise clustering, CCFs were grouped into 45-degree orientation bins, with density heatmaps computed for each (Figure 7.18 for the northern hemisphere; Figure 7.19 for the southern hemisphere). In both cases, the first panel presents the overall distribution of all CCFs, categorised by slope quartile. Subsequent panels plot the density of CCFs with above-lower-quartile slope within each orientation bin.

In the northern hemisphere, a northeast-to-southeast aspect bias is apparent, particularly around $0^\circ - 90^\circ$ longitude in Arabia Terra. However, the distribution is broad, with no single orientation dominating, and these biases aren't confined to specific regions. In contrast, the southern hemisphere shows a more consistent and spatially constrained clustering. A dominant south- and southeast-facing bias is evident, with southwest, south, and southeast aspects tending to cluster within the same regions. While smaller, localised clusters of other orientations exist, such as on the western flank of Hellas (likely influenced by background topography), the overall aspect clustering is more pronounced and spatially consistent than in the north.

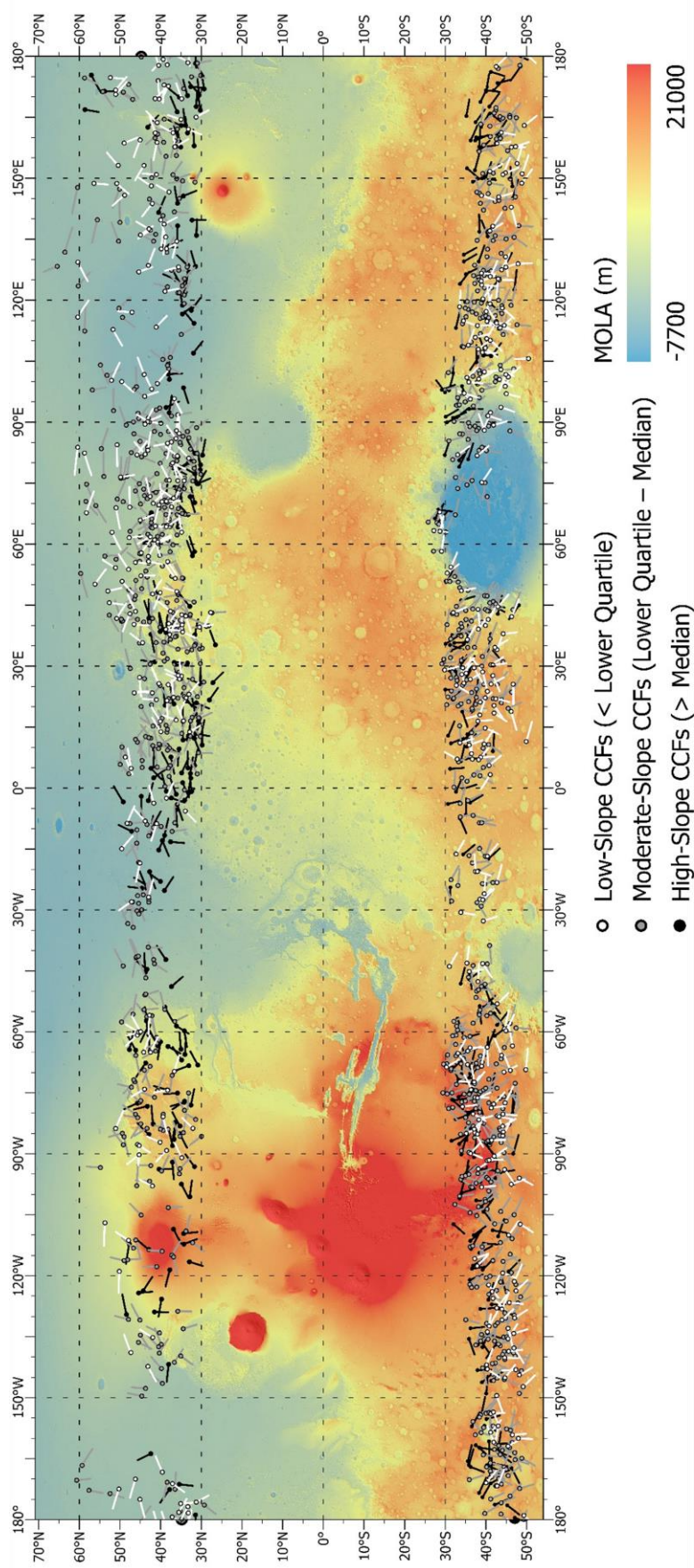


Figure 7.17. Map of CCF Aspect Orientations with Slope Magnitude Classification. MOLA gridded topography overlain on MOLA-derived hillshade, showing CCF aspect distribution with aspect directions represented by bearing lines. CCFs are colour-coded by slope: white (low slopes), grey (lower quartile to median), and black (above median). The Northern Hemisphere exhibits a diffused distribution with no discernible orientation clustering, while the Southern Hemisphere shows more localised clustering with a consistent south-facing bias and a pronounced southeast orientation trend.

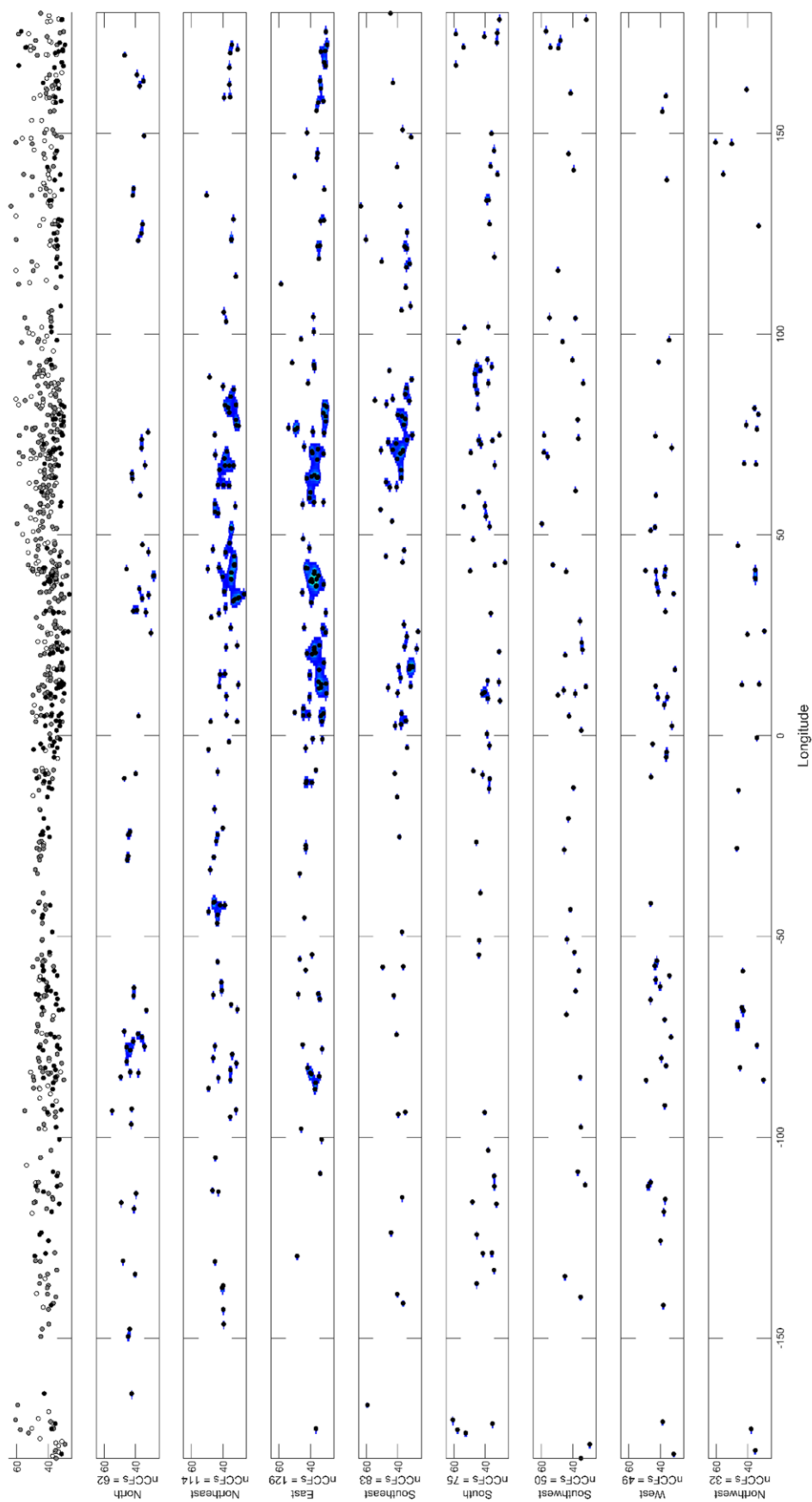


Figure 7.18. Northern Hemisphere CCF Aspect Density Heatmap. CCFs are binned into 45° aspect intervals, with density maps highlighting regional clustering of aspects. The first panel displays all CCFs categorised by slope quartiles, subsequent panels show the density distributions of CCFs with slopes above the lower quartile, highlighting orientation concentrations in 45 -degree bins, from north to west, in a clockwise direction.

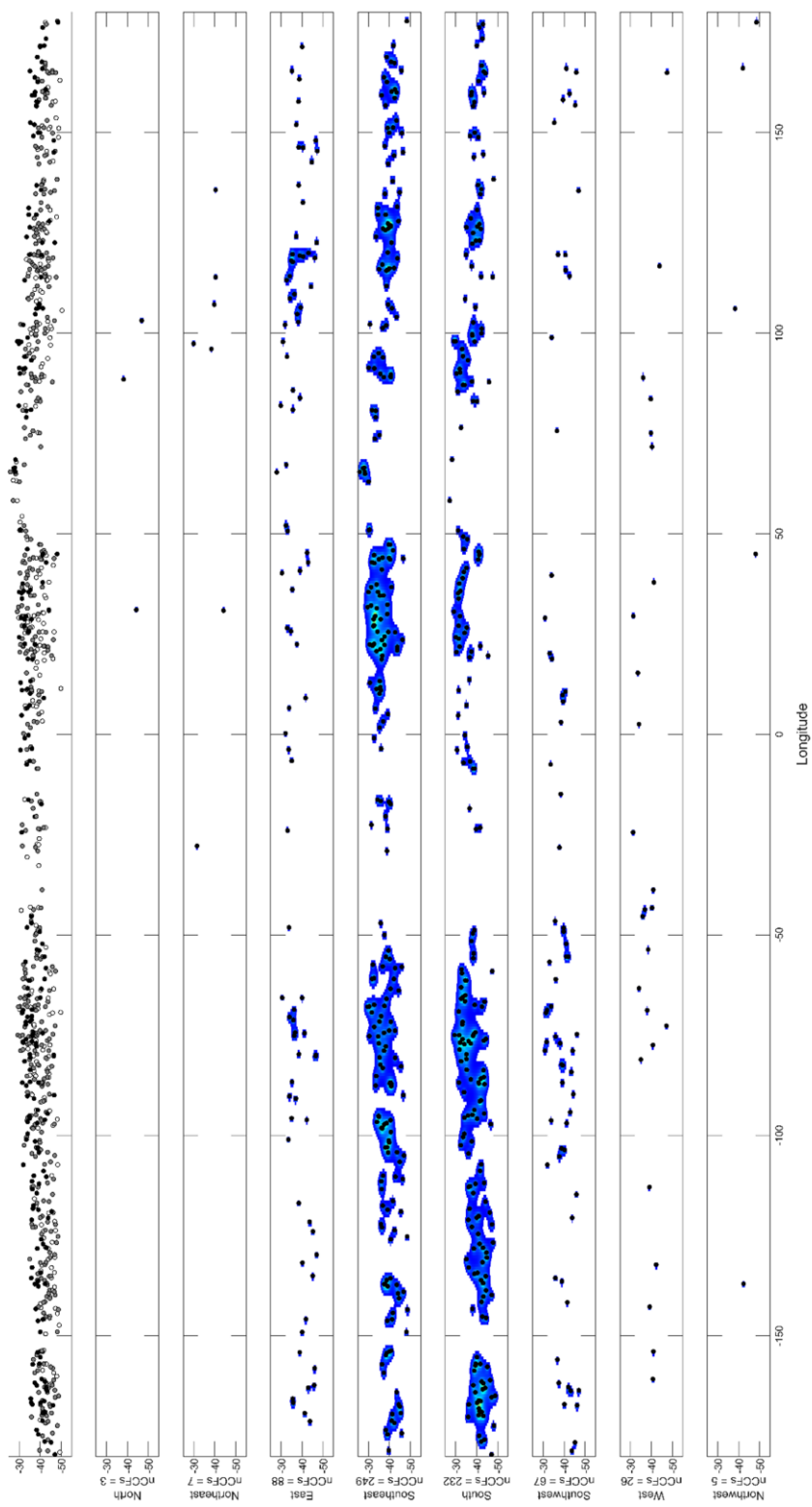


Figure 7.19. Southern Hemisphere CCF Aspect Density Heatmap. CCFs are binned into 45° aspect intervals, with density maps highlighting regional clustering of aspects. The first panel displays all CCFs categorised by slope quartiles, subsequent panels show the density distributions of CCFs with slopes above the lower quartile, highlighting orientation concentrations in 45° -degree bins, from north to west, in a clockwise direction.

7.3.2.2. CCF Aspect in different regions

What do the rose plots of aspects (and accompanying statistics) tell us about CCF aspects in different regions across each hemisphere? To explore this, I adopt a regional framework based on a previous global study of gullies by Noblet et al. (2024), who focused on 11 topographically distinct regions across the North and South mid latitudes. Noblet et al. (2024) found that the aspect statistics of gully hillslopes are constrained by latitude and have high regional variation, with aspect distributions reflecting the orientations of large reliefs and extensive geological features specific to each region.

It is worth exploring whether the CCFs in this chapter's CCF sample display any form of regional biases, which could suggest the influence of background topographic relief or superposed climatic effects acting at regional to global scales. To investigate aspect variations across the entire CCF population without altering the total sample size, I adopt the same 11 regions used by Noblet et al. (2024) which encompass the basins of Hellas and Argyre, the plains of Arcadia, Utopia, and Acidalia Planitia, and the highlands of Arabia, Aonia, Sirenum, Noachis, Tempe, and Cimmeria Terrae.

Analysis of the regional framework reveals contrasting patterns of aspect variability between hemispheres. In the northern hemisphere (Figure 7.20), regional CCF aspect patterns vary in clustering strength and directional bias. Arcadia Planitia and Tempe Terra (Figure 7.20A, B) exhibit uniform aspect distributions with no statistically significant clustering (*Hermans-Rasson*, $p = 0.287$ and $p = 0.042$, respectively). While the low number of CCFs in Arcadia Planitia ($n = 35$) may limit the emergence of a preferential orientation, Acidalia Planitia (Figure 7.20C) displays a northeast orientation bias with statistically significant clustering (*Hermans-Rasson*, $p = 0.001$) despite a similarly small population ($n = 47$). Arabia Terra ($n = 246$, Figure 7.20D) and Utopia Planitia ($n = 162$, Figure 7.20E) both show statistically significant clustering (*Hermans-Rasson*, $p < 0.001$), though their overall distributions remain broadly dispersed. In both regions, circular means are east-facing, with CCFs with surface slopes above the regional median being predominantly east-facing.

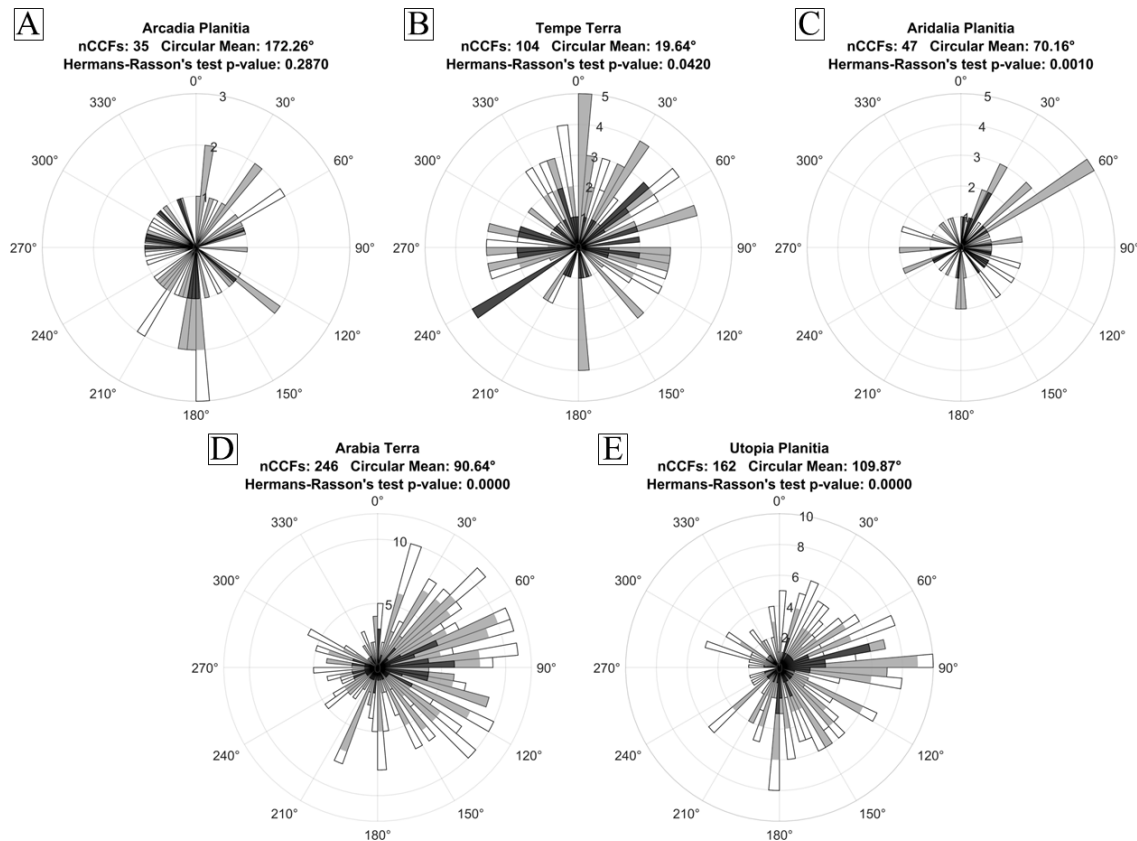


Figure 7.20. Compass-Rose Plots of regional Northern Hemisphere CCF Aspect Distributions: (A) Arcadia Planitia (-180° to -120°), (B) Tempe Terra (-120° to -50°), (C) Acidalia Planitia (-50° to -5°), (D) Arabia Terra (-5° to 75°), and (E) Utopia Planitia (75° to 180°). Aspect orientations are binned at 5° intervals. Arabia Terra and Utopia Planitia exhibit statistically significant clustering ($p < 0.001$), with Arabia Terra and Utopia Planitia showing a northeast to southeast bias. In other regions, orientations are more dispersed, with no statistically significant clustering.

All regions in the southern hemisphere were found to have statistically significant clustering of CCF surface aspects (*Hermans-Rasson*, $p < 0.001$) (Figure 7.21), with minor variation between regions. Terra Sirenum ($n = 126$, Figure 7.21A), Aonia Terra ($n = 171$, Figure 7.21B), and Argyre Basin ($n = 33$, Figure 7.21C) all exhibit south-facing circular means. Among these, Terra Sirenum shows the greatest variation in CCF aspect, containing the majority of the southern hemisphere's Southwest-oriented CCFs. This may indicate a regional influence that maintains the broader pole-facing bias while allowing for localised deviations toward Southwest orientations. Noachis Terra ($n = 120$, Figure 7.21D), Hellas Basin ($n = 86$ Figure 7.21E), and Cimmeria Terra ($n = 141$ Figure 7.21F) all exhibit southeast circular means. Unlike the northern hemisphere, poleward dominance remains consistent and statistically significant across southern hemisphere regions.

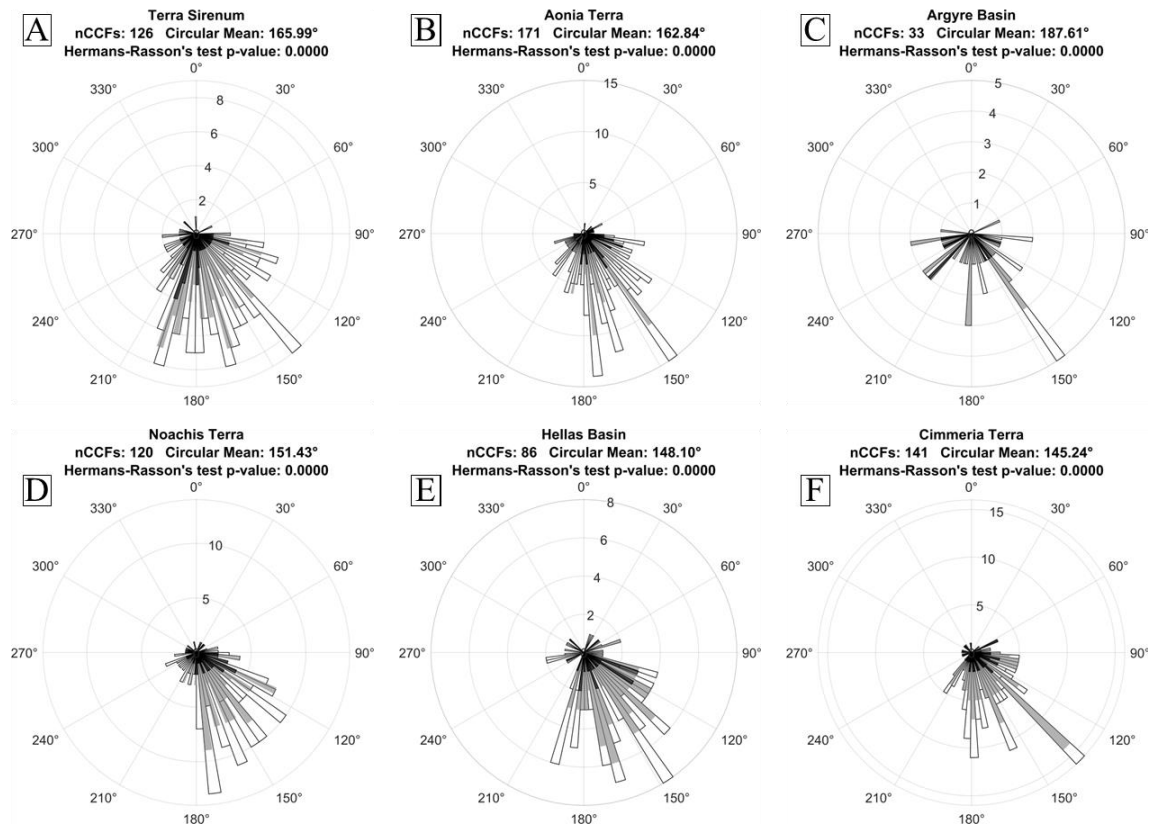


Figure 7.21. Compass-Rose Plots of regional Southern Hemisphere CCF Aspect Distributions: (A) Terra Sirenum (-180° to -120°), (Bb) Aonia Terra (-120° to -60°), (C) Argyre Basin (-60° to -25°), (D) Noachis Terra (-25° to 43°), (E) Hellas Basin (43° to 102°), and (F) Cimmeria Terra (102° to 180°). Aspect orientations are binned at 5° intervals. All regions exhibit statistically significant clustering, with a predominant poleward bias.

In summary, regional CCF aspect distributions differ markedly between hemispheres. In the northern hemisphere, aspect patterns vary substantially between regions and range from uniform to weakly clustered, indicating no single, coherent hemispheric-scale preferred orientation. In contrast, the southern hemisphere exhibits a consistently poleward-oriented and statistically significant aspect bias across all regions, implying that a hemispheric-scale control operates there with sufficient strength to override regional variability, even if local influences remain present.

7.3.2.3. Latitudinal variations in CCF aspect

In the northern hemisphere (Figure 7.22), CCF aspect distributions vary with latitude systematically. Aspects at latitudes <40°N (Figure 7.22A, B) show statistically significant clustering (Hermans-Rasson $p < 0.001$), with a strong eastward bias, particularly for steeply-sloping CCFs equatorward of 35°N. This eastward clustering weakens at higher latitude (becoming statistically non-significant: Hermans-Rasson $p > 0.001$), where a wider range of orientations appears, including more west- and south-facing CCFs.

In the southern hemisphere (Figure 7.23), south to southeast aspects dominate across latitudes, with statistically significant clustering (*Hermans-Rasson* $p < 0.001$) and minimal variation in the mean aspect. Circular means consistently trend southeast, varying by no more than 10° between latitude bins. Although a broader range of orientations emerges south of 45°S , the poleward bias persists. In each bin, only rare instances of non-poleward orientations occur, involving no more than a few CCFs.

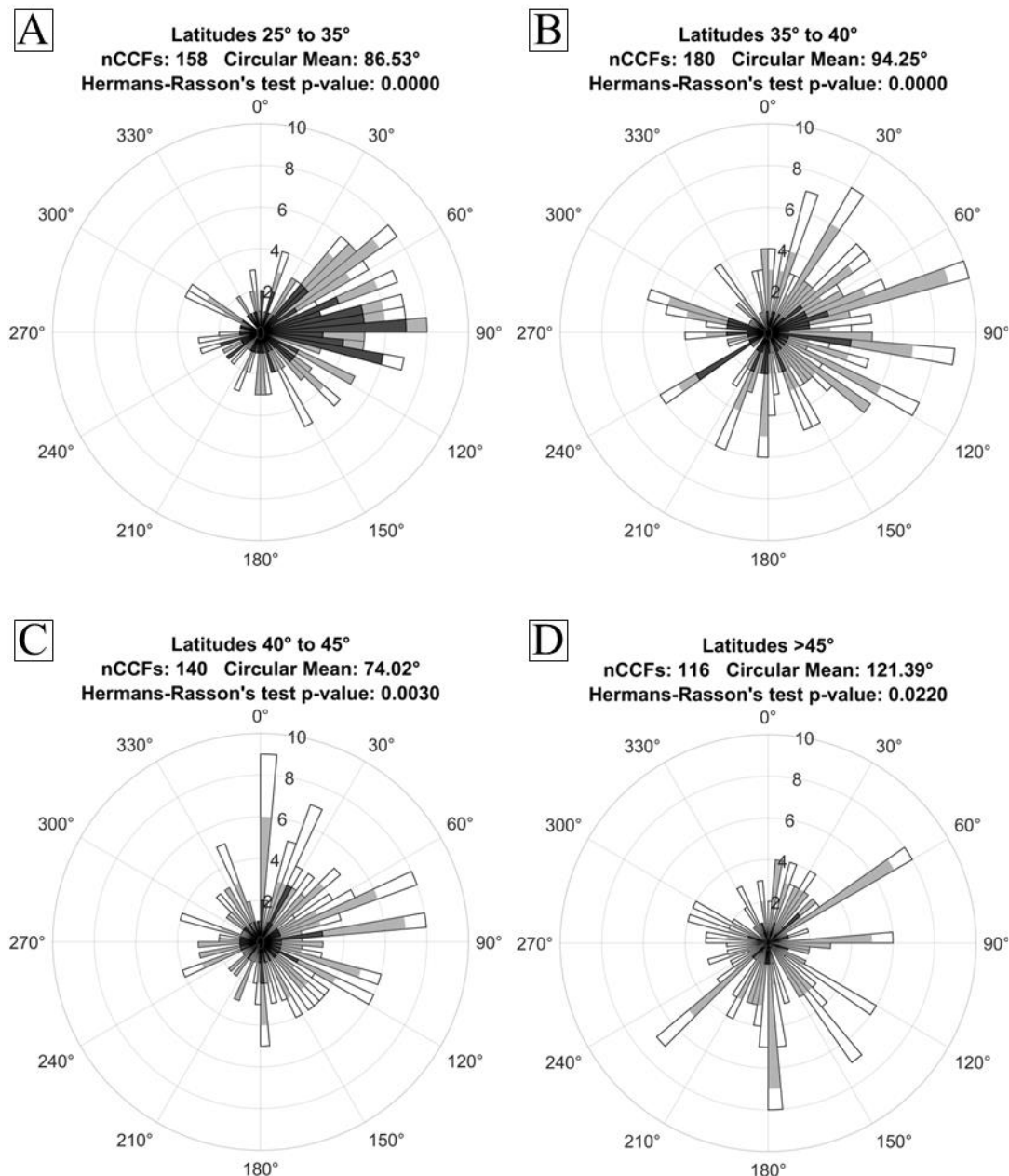


Figure 7.22. Compass-Rose Plots of Northern Hemisphere CCF Aspect Distributions for different latitudes: (A) 25–35°, (B) 35–40°, (C) 40–45°, and (D) >45°. Aspect orientations are binned at 5° intervals. A strong eastward bias is observed at lower latitudes, with significant clustering in (A) (86.53°) and (B) (94.25°) ($p < 0.001$). At higher latitudes, orientation distributions broaden, incorporating more west- and south-facing aspects. Clustering is non-significant above 40° , where steep slopes become less frequent.

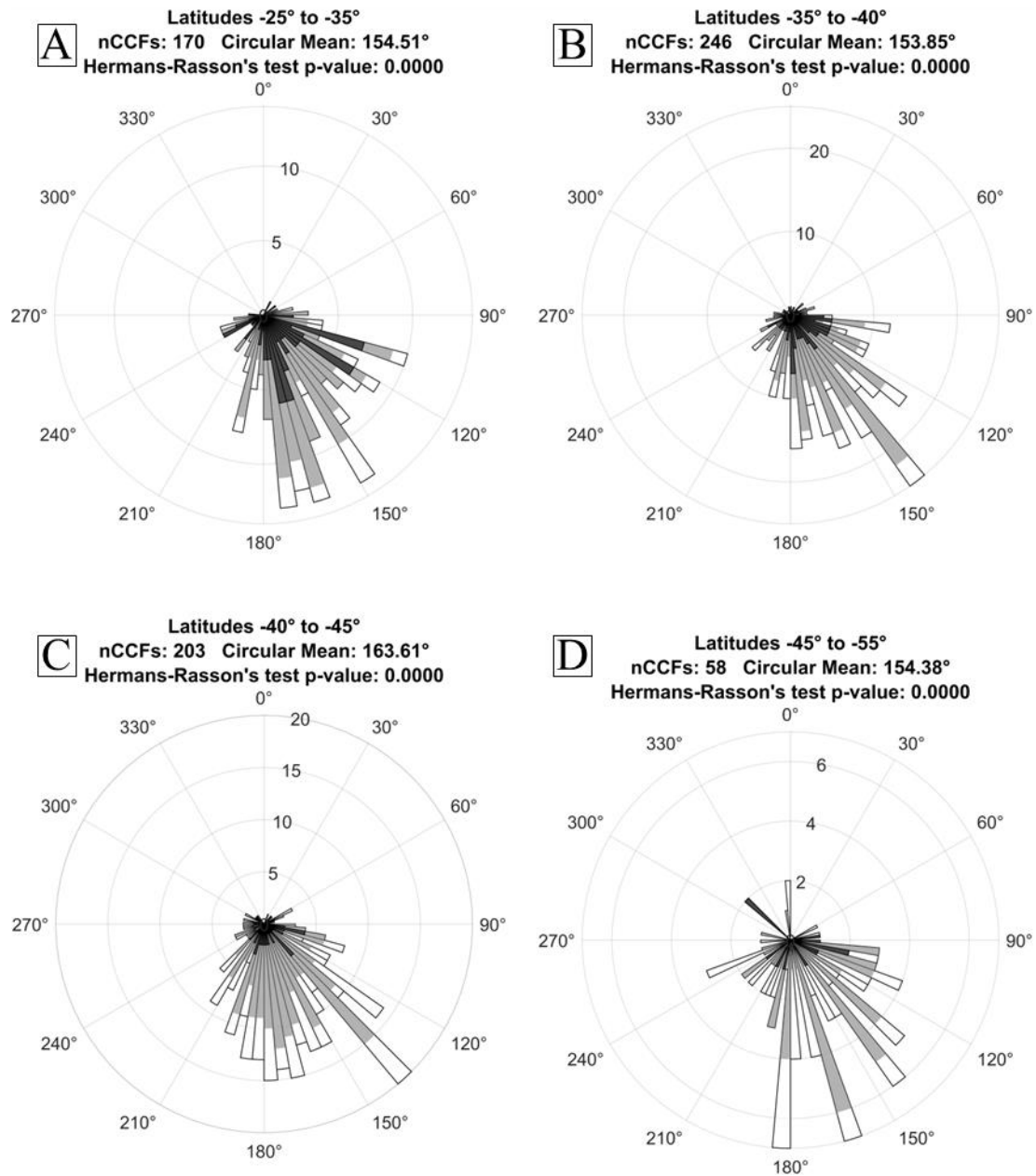


Figure 7.23. Compass-Rose Plots of Southern Hemisphere CCF Aspect Distribution for different latitudes: (A) -25° to -35°, (B) -35° to -40°, (C) -40° to -45°, and (D) -45° to -55°. Aspect orientations are binned at 5° intervals. A strong poleward bias is observed at lower latitudes, with significant clustering across all bins ($p < 0.001$). Circular mean values remain consistent, with (A) 154.51°, (B) 153.85°, (C) 163°, and (D) 154.38°.

7.3.2.4. Aspect and host crater diameter

Figures 7.24 and 7.25 display the relationship between CCF surface aspect and D in the Northern and Southern hemispheres, respectively.

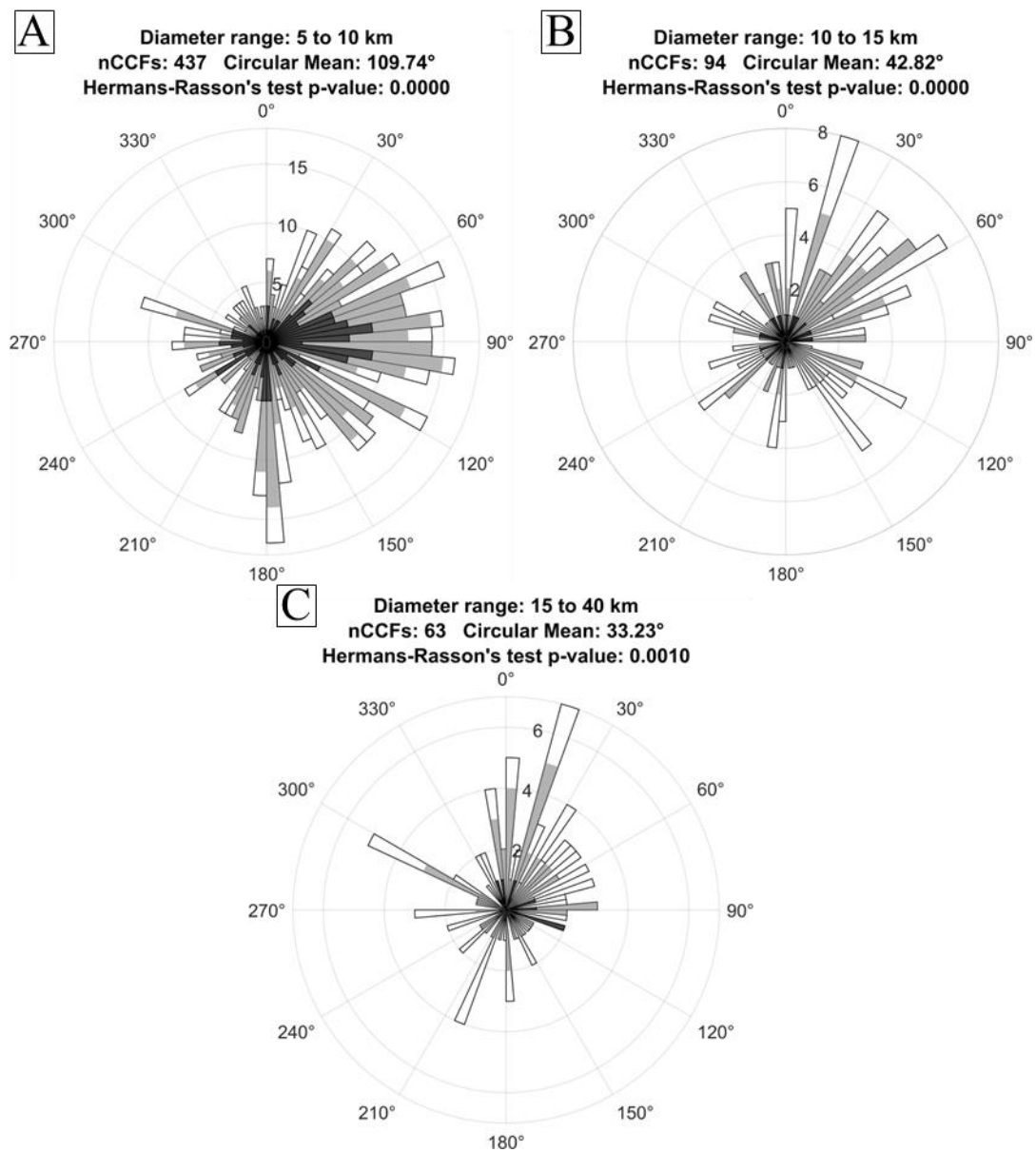


Figure 7.24. Compass-Rose Plots of Northern Hemisphere CCF Aspect Distribution Across Host Crater Diameter Ranges. Northern Hemisphere compass-rose diagrams depicting CCF aspect distribution across three host crater diameter ranges: (A) 5–10 km, (B) 10–15 km, and (C) >15 km. Aspect orientations are binned at 5° intervals.

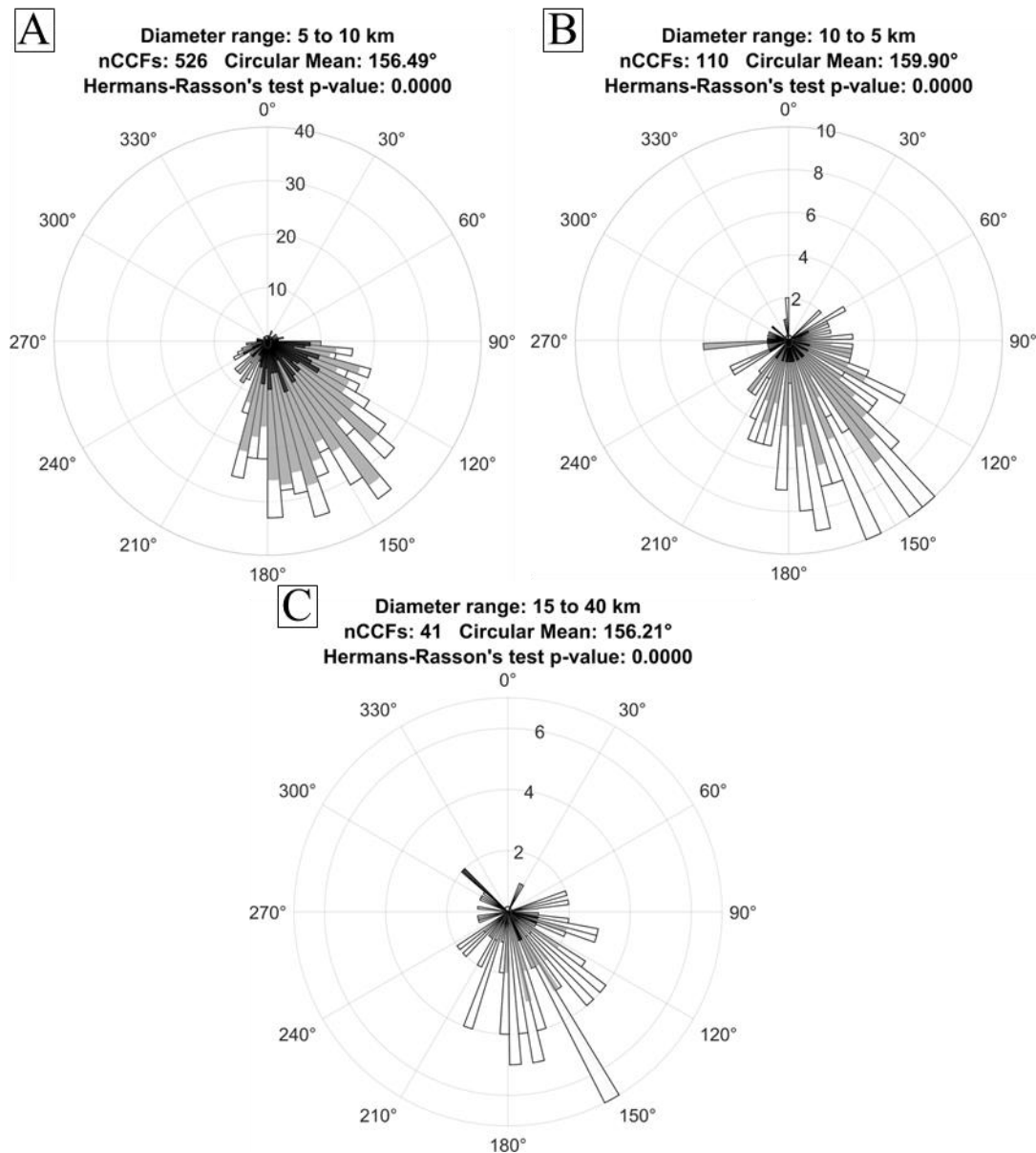


Figure 7.25. Compass-Rose Plots of Southern Hemisphere CCF Aspect Distribution Across Host Crater Diameter Ranges. Southern Hemisphere compass-rose diagrams depicting CCF aspect distribution across three host crater diameter ranges: (A) 5–10 km, (B) 10–15 km, and (C) >15 km. Aspect orientations are binned at 5° intervals. All three diameter ranges illustrate a consistent South and Southeast orientation bias with minimal equator-facing slopes.

CCF aspect distributions vary weakly with D in the northern hemisphere (Figure 7.24), around the total hemispheric distribution reported earlier in Figure 7.16. For smaller host craters ($D < 10$ km, Figure 7.24A), the distributions are broad with a general east-facing bias but also contain secondary clusters toward the south and west. As crater D increases, the circular mean of the aspect shifts poleward. The medium-sized craters (10–15 km, Figure 7.24B) show statistically significant clustering (*Hermans-Rasson*, $p < 0.001$), but the largest craters (15–40 km, Figure 7.24C) show weaker clustering (*Hermans-Rasson*, $p = 0.001$), indicating a possible transition toward more uniform or concentric distributions.

In contrast, the southern hemisphere distributions are strikingly consistent across all D bins (Figure 7.25A–C). The circular means of the bins reflect the persistent south-southeast bias and differ by only 3.7° . All bins exhibit statistically significant clustering (*Hermans-Rasson*, $p < 0.001$), with a scarcity of equatorward orientations. Thus, the orientation spread and clustering of Southern CCFs are largely independent of D . This finding is surprising, given that D accounts for $\sim 41\%$ of the variation in CCF d_a (Chapter 5); as d_a defines the vertical offset between the mean rim elevation and mean CCF surface, one might expect this significant host crater control to influence the preferred aspect through changes in topographic shadowing and radiative exchange.

7.3.3. Overall interpretation

Three key findings have emerged so far: CCFs in both hemispheres exhibit very shallow slopes, a poleward aspect bias (with different strengths in the north and south) and an eastward aspect bias (particularly dominant in the north). The northern hemisphere has a broader aspect distribution than the south, where the poleward orientations are tightly clustered. The poleward aspectual biases, together with the observed poleward decay of CCF slopes (Section. 7.3.1.1; Figure 7.9) indicate hemispherical symmetry in some of the processes (or process histories) determining the CCF surfaces. The shared poleward and eastward biases across hemispheres suggest that, whatever distinct mechanism(s) cause these aspect biases, the same underlying mechanisms are present in both hemispheres. The mechanism causing the poleward bias is stronger in the south, producing a more constrained and uniform aspect distribution across latitude, diameter, and regional subsets.

The observed poleward preference in CCF aspect aligns with biases found in previous studies of VFFs (Milliken et al 2003; Souness et al., 2012; Figure 7.26). Souness et al. (2012) found that northern hemisphere GLFs include many south-facing examples that deviate from the expected poleward bias (Figure 7.26B), a pattern not observed in the southern hemisphere GLF population (Figure 7.26E). Although the northern hemisphere GLFs aspect distribution is less pronounced than the broad northern hemisphere CCF aspect distribution reported in this chapter, the similarity suggests that the mechanism influencing CCF aspects in the north may also operate on northern GLFs, despite their seemingly younger age (Hepburn et al., 2020). The southern hemisphere VFF comparison (Figure 7.26 D-F) is broadly consistent with the CCF results, with GLFs showing a predominantly south-facing, poleward aspect distribution spanning roughly southeast to southwest. Southern LDAs and LVF also show mainly pole-facing orientations, although their distributions appear more strongly biased towards the southwest than the southeast, unlike the dominant southeast clustering of southern CCFs.

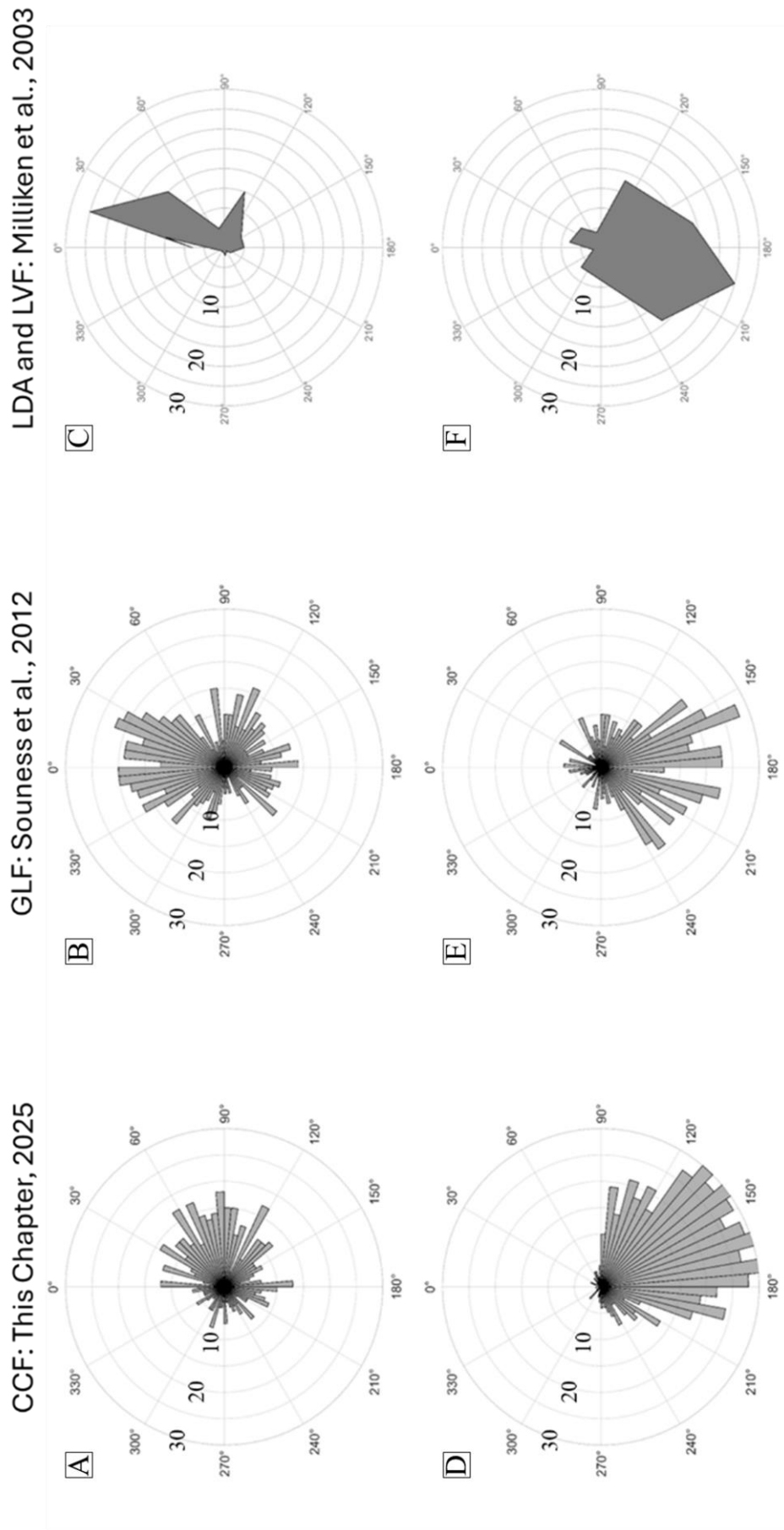


Figure 7.26. Comparison of CCF surface aspect distributions with previous VFF aspect studies. Rose plots showing aspect distributions for VFFs. The top row presents northern hemisphere data; the bottom row shows southern hemisphere data. **Left:** CCFs (this study). **Centre:** GLFs (redrawn from Souness et al., 2012). **Right:** LDA and LVF (redrawn from Milliken et al., 2003). All plots use consistent binning and orientation, with north at 0°.

Separately, it is interesting to consider the findings of this chapter alongside the cratered flow patterns described by Dickson et al. (2012) (Figure 7.5) as, together, they form a broader understanding of the directional properties of crater confined VFFs. However, it is important to recognise that these two sets of findings are not directly comparable, nor are they communicable as measurements of the same phenomenon, because they describe fundamentally different physical properties. While Dickson et al. (2012) mapped the orientation of surface lineations and lobes (flow indicators), this chapter examines the surface normal direction (aspect), which represents the macro-topographic gradient of the CCF surface itself. Furthermore, the features being analysed are fundamentally different. The latitudinal transition of flow orientation at 40–45°N/S (Figure 7.5) identified by Dickson et al. (2012) may primarily reflect the geographic boundary between the prevalence of pole-facing flow features and the onset of CCF dominance. While CCFs (and thus concentric flow orientations) are found equatorward of 40–45°N/S (Levy et al., 2014; Chapter 4), they may not dominate the broader sample of flow features at those lower latitudes. Because this chapter specifically focuses solely on the high-confidence CCF sub-population, the results do not—and should not be expected to—replicate the transition from pole-facing linear flows to concentric patterns. In the southern hemisphere, CCF surface aspects consistently maintain a strong poleward orientation across all latitudes. In the northern hemisphere, an easterly bias dominates below 40°, but clustering weakens beyond this latitude, transitioning into a broader pattern.

Ultimately, considering these two datasets together establishes a distinction in the directional behaviour of ice-related features over time. Crater-confined, pole-facing flow features—which are thought to be geologically recent (~10–100 Myr; Sinha and Vijayan, 2017)—exhibit a strong pole-facing orientation preference in both hemispheres. This hemispheric symmetry is not replicated in CCF surface aspects, which instead display marked hemispheric differences, with a consistent poleward orientation in the south and a more stochastic distribution in the north. This suggests that while recent, small-scale surface flows are driven by symmetric climatic forcing (solar insolation), the macro-topography of the older, bulk CCF deposits may be influenced by hemispheric differences in CCF evolution and longer-term glacial histories.

7.4. Discussion

What mechanisms can cause CCF surfaces to slope (Section 7.4.1)? What explains the prevalence of poleward (Section 7.4.2) and east-facing CCF surfaces (Section 7.4.3)? And what mechanism causes hemispheric differences between their aspectual distributions (Section 7.4.4)? In this section, I discuss potential causes behind these observations, drawing on relevant physical factors on Mars and insights from terrestrial glaciation. I begin first with CCF surface slope development.

7.4.1. CCF slope development

I begin by considering how CCFs can become sloped at all (even though their observed slopes are very gentle). Any persistent slope requires the development of lateral asymmetry in mass balance, either

during palaeo-CCF accumulation or during subsequent retreat. Whether ancestral ice masses (the predecessors of today's CCFs) remained crater-confined throughout their history or overtopped the crater rims for part of that history (e.g. during regional glaciation) also matters and will influence subsequent aspectual discussions. These configurations lead to different interpretations of how slope arose.

Thus, three broad scenarios can be conceived to explain CCF surface slope development:

Scenario 1: Crater-confined slope formation during active glaciation

This scenario assumes that palaeo-CCFs remained confined within their host craters throughout their growth and that present-day slopes were established during the accumulation phase itself.

Under orbital conditions favourable to mid-latitude ice accumulation (e.g., Madeleine et al., 2009), ice is expected to nucleate preferentially on the interior crater walls and to flow downslope as discrete lobes that progressively coalesce to blanket the crater floor (Weitz et al., 2018). In this setting, any lateral asymmetry in mass balance along the crater walls will be directly inherited by the developing CCF geometry. Such asymmetry can arise through several physically motivated mechanisms. Differential insolation on opposing crater walls (Section 7.4.2) can promote preferential accumulation or reduced ablation on pole-facing or otherwise shadowed sectors. Thermal asymmetry within crater interiors (Section 7.4.2) can further bias where ice is most stable on the walls and on the nascent lobes (Section 7.4.3.1). In addition, reflected shortwave radiation from illuminated crater walls onto the ice surface, and diurnally varying cloud cover (Section 7.4.3.2), can introduce systematic lateral differences in net energy balance across the forming glacier surface. In this scenario (Figure 7.27), ice initially accumulates as wall-fed lobes, but asymmetric mass balance causes these lobes to thicken preferentially on one side of the crater. As the lobes merge and the crater floor becomes fully ice-covered, this imbalance is preserved as a gentle but persistent surface tilt. Once accumulation ceases, development of a debris-rich sublimation lag would strongly suppress subsequent modification, allowing the formation-stage surface geometry to be retained through later interglacial periods. Under this model, present-day CCF slopes directly record the spatial pattern of mass balance during the final phase of active glaciation, and the ice mass never overtopped the crater rim.

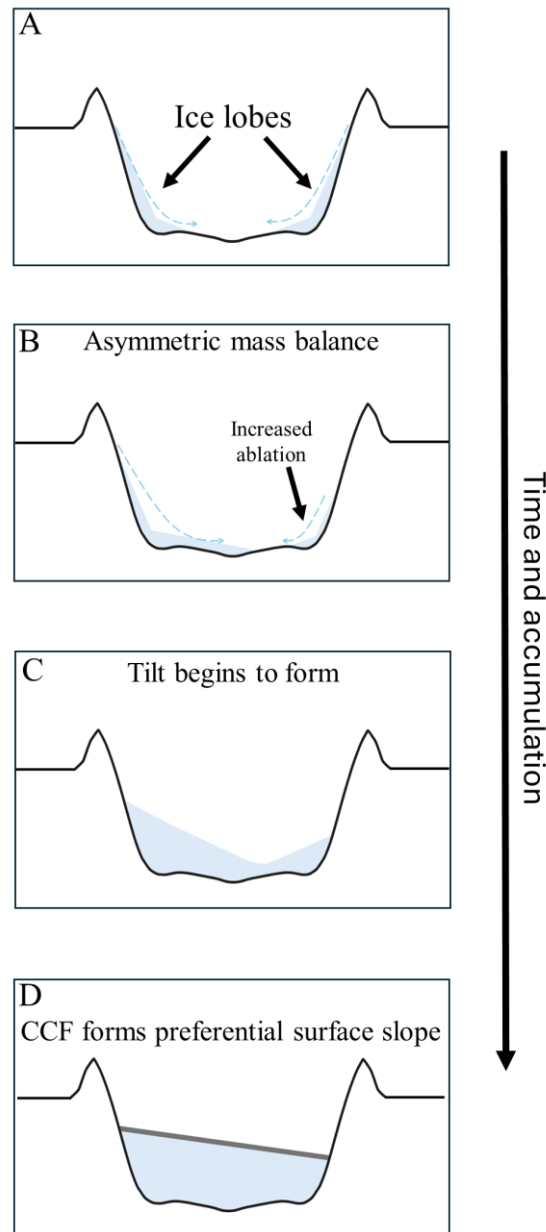


Figure 7.27: Conceptual model illustrating the formation of surface slopes on crater-confined CCFs during active glaciation. In this scenario, present-day slopes reflect formation-stage processes preserved under a sublimation lag during later interglacial periods. (A) During active glaciation, ice accumulates on the interior walls and flows toward the crater floor. (B) Asymmetric accumulation and/or ablation on one side of the interior crater wall promotes uneven ice accumulation. (C) Over time, this imbalance produces a preferential surface slope as the lobes connect and the palaeo-CCF forms, similar in form to the transect in Fig. 7.2B. (D) Over time, as the CCF continues to accumulate and infill the crater, a preferential surface slope develops.

The following two scenarios assume that ancestral CCFs were initially more voluminous than at present and that their modern surface slopes developed primarily during post-glacial mass loss rather than during the main phase of ice accumulation. In these cases, the present-day slope geometry is interpreted as an ablation-generated feature rather than an accumulation one. The same physical

mechanisms capable of producing lateral asymmetry during active glaciation can also operate during retreat, including, reflected shortwave radiation from crater walls (Section 7.4.1), thermal asymmetry within the crater interior (Section 7.4.3.1) and diurnally varying cloud cover (Section 7.4.3.2). Each of these processes can impose persistent lateral gradients in surface energy balance, leading to systematically greater ice loss on one side of the deposit and, over time, the development of a sloping CCF surface.

Scenario 2: Slope formation through post-glacial ablation without overtopping

In this scenario, ancestral CCFs once occupied a larger fraction of their host craters, perhaps even to rimfull states, but never overtopped the crater rim. Their surfaces are assumed to have been approximately horizontal, or at least less inclined than at present, at the end of the main accumulation phase. Subsequent downwasting during post-glacial retreat then increased d_a and, because ablation was laterally asymmetric, generated the present-day surface slope (Figure 7.28). In this interpretation, CCF slope is a purely degradational feature that records the geometry of ice loss rather than the geometry of ice accumulation. This scenario implies that the modern slope direction reflects the long-term pattern of preferential retreat within the crater interior, while the palaeo-CCF surface has been largely overprinted. However, because the ice remained confined to the crater throughout its history, the total amount of surface lowering required is moderate, and the CCF never experienced a phase of large-scale evacuation from beyond the rim.

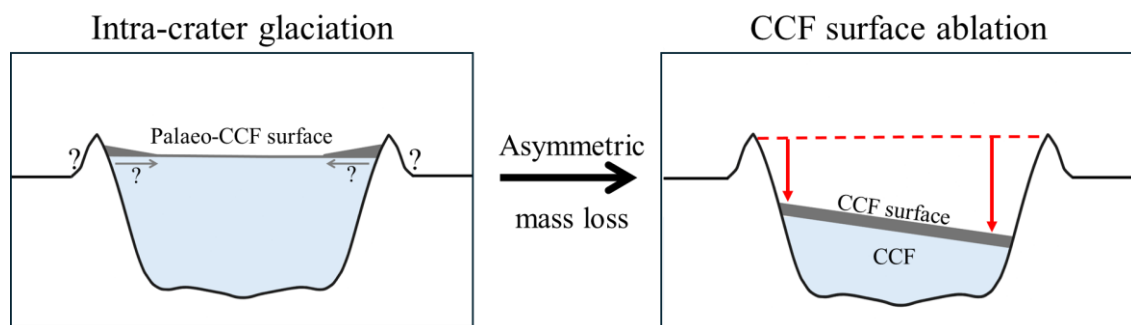


Figure 7.28. Conceptual model for slope formation through post-glacial ablation without overtopping. Slope exaggerated in panel two for visualisation.

Scenario 3: Slope formation following overtopping and retreat

In this scenario, ancestral ice masses overtopped their host-crater rims, either as part of a regional ice sheets or through localised ice domes. The present crater-confined geometry therefore requires substantial post-glacial ice loss to remove the extra-crater ice and to lower the surface back into the crater interior (Figure 7.29). During this large-magnitude retreat phase, laterally asymmetric ablation generated the observed surface slope. Because the magnitude of ice loss is much greater than in Scenario 2, this scenario implies that any original palaeo-CCF slope formed during accumulation is unlikely to

have been preserved. Instead, the modern CCF surface slope is interpreted to record only the recession-phase geometry, with its orientation and magnitude governed primarily by the pattern of asymmetric retreat rather than by the original mode of emplacement. CCF surface slope and aspectual bias can form during ice cap retreat and paleo-CCF crater confinement or during crater confined ablation. In this sense, Scenario 3 mirrors Scenario 2 in its motivation for CCF slope formation, differing only in the magnitude of initial ice cover and the scale of subsequent evacuation.

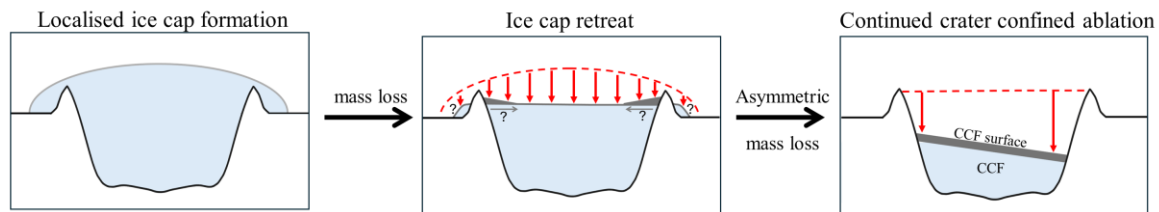


Figure 7.29. Conceptual model of slope formation following CCF overtopping and subsequent retreat. Slope exaggerated in panel three for visualisation.

These scenarios are conjectural. In reality, a complicated history is likely to have played out in Mars's mid-latitudes that, at various times, and/or in various places involved all three scenarios, or hybrids of them. Then, the relevant scenario is the one that pertains to the *last major change/reconfiguration* of the ancestral mid-latitude ice masses leading to their present state, as CCFs. The scenarios highlight that CCF slope and aspect could result from processes active either during glaciation or during ice retreat. They give a foundation for considering the mechanisms or factors behind the needed asymmetry (at various stages of CCF evolution) and specific features of my slope and aspect results, as done next.

Numerical modelling of CCF formation and spatially limited geomorphic evidence suggests that CCF's ancestral ice volumes may have been significantly greater than at present, implying that surface slopes in at least a subset of cases may have developed during post-glacial retreat rather than solely during the primary accumulation phase (i.e., Scenarios 2 and 3, Figure 7.28 & 7.29). Numerical modelling by Weitz et al. (2018), using orbital configurations favourable to mid-latitude glaciation and accumulation rates of $\sim 10 \text{ mm yr}^{-1}$, shows that even large craters ($D \leq 40 \text{ km}$) can be substantially infilled on timescales of $< 5 \text{ Myr}$. Spatially limited geomorphological evidence supports this inference, with interior trimlines documented by Levy et al. (2010) interpreted as recording former CCF surface extent far above present ones, demonstrating major post-glacial thinning within craters and favouring scenarios involving retreat from a more voluminous crater-confined ice body (Scenario 2, Figure 7.28). Exterior moraines and associated landforms described by Dickson et al. (2010) require CCF ice to have extended beyond crater rims, providing direct evidence for rim overtopping followed by large-magnitude retreat (Scenario 3, Figure 7.29). Although these examples are regionally restricted and cannot be assumed to represent the global CCF population, they demonstrate that both substantially thicker crater-confined states and overtopping states were achieved by at least some systems. While these examples are spatially

restricted to two examples in the northern hemisphere, they confirm that both substantial thinning and large-magnitude retreat are physically grounded pathways for generating present-day CCF slopes. Scenario 1 (Figure 7.27), however, may still apply in cases where ice volumes remained modest and crater-confined throughout their history.

7.4.1.1. Radiative Feedback on CCF Slopes

Framed around the scenarios previously discussed, the upcoming discussion sections focus on mechanisms that address how the interaction of slope with solar radiation drives the development of preferential aspects during both glacial and interglacial periods.

A critical consideration, regardless of aspect or proposed mechanism, is that slope–radiation interactions can generate a positive feedback once a CCF surface begins to tilt. The development of even a small initial slope increases the proportion of the adjacent, interior crater wall exposed to direct solar radiation. This increases the wall’s total heated surface area and, consequently, the magnitude of thermal re-radiation incident upon the neighbouring CCF surface. The resulting enhancement of ablation on the slope aligned margin reinforces the tilt, lowering d_a , which in turn exposes more wall area to heating. Over successive cycles, this feedback can magnify slope asymmetry from relatively minor initial differences in wall–ice radiative exchange.

And yet, despite this feedback, CCFs possess very low slopes (Section 7.3.1). Either (i) slopes formed primarily during active accumulation (Scenario 1) and were subsequently stabilised and preserved during retreat, or (ii) if slopes developed during post-glacial downwasting (Scenarios 2 or 3), then a debris-rich sublimation lag must have formed sufficiently early to suppress further feedback-driven slope amplification. In this context, the critical control is not simply the presence of debris cover, but the relative rate at which insulating debris accumulated compared to the rate of ice loss by sublimation. Or (iii) CCFs have not reached mass balance equilibrium under present conditions and are still undergoing mass loss but at an incredibly low undetectable rate due to their overburden.

These perspectives also offer a coherent explanation for why many mid-latitude craters do not host preserved CCFs today. If ancestral crater-confined ice was once more widespread during mid-latitude glaciation, long-term survival would have depended on whether a given crater could develop and sustain an effective sublimation lag/overburden layer. Craters that received sufficient debris input from wall collapse, mass wasting, or atmospheric dust deposition could suppress runaway sublimation and preserve a recognisable CCF into the present. Craters with insufficient debris/dust supply, by contrast, would have been unable to stabilise their ice, leading either to complete loss of the deposit or to the formation of highly degraded, low-definition CCF remnants (Section 3.2.4).

7.4.2. Mechanism for poleward bias

The poleward bias observed in CCF surface slopes is most plausibly explained by asymmetry in surface insolation, with increased insolation on equator facing slopes influencing sublimation rates during active glaciation and/or subsequent retreat. This interpretation is grounded in terrestrial glaciology (E.g., Evans, 1977; Evans, 2006) and well established in previous Martian ice stability studies dependant on obliquity and latitude (e.g., Kreslavsky et al., 2008), which identify pole-facing slopes in mid-latitude craters as preferred sites for cold trapping (Hecht, 2002) and thus long-term ice preservation (Sinha and Vijayan, 2017).

Previous work by Kreslavsky et al. (2008) has shown that the relationship between slope orientation and insolation is not monotonic and depends strongly on obliquity, latitude, and season. Their insolation calculations across varying obliquity, latitude and slope find that at high obliquity, pole-facing slopes at mid-latitudes can in fact receive higher summer insolation than equator-facing slopes, while at low obliquity the orientation dependence is weak except at extreme latitudes. Their schematic (Figure 7.6) shows that insolation patterns on crater walls vary in a complex, non-intuitive manner across orbital states, and that simple expectations of systematically colder pole-facing slopes do not always hold.

As a result, although slope–insolation coupling is well established in principle, it remains unclear whether the observed poleward bias in CCF interior slopes should persist across the range of orbital conditions relevant to CCF emplacement and retreat, or whether it should reverse or weaken under certain climate states. Moreover, existing studies focus on general crater-wall thermal regimes and do not specifically evaluate the magnitude of aspect-dependent insolation contrasts for the very shallow slopes characteristic of CCF surfaces.

To address this, I explicitly quantify how daily-mean and annual-mean insolation vary with latitude, orbital configuration, slope, and aspect for crater-interior slopes ($\beta = 30^\circ$). I then evaluate whether these variations are sufficient to generate and reinforce the observed poleward CCF slope bias, and whether extremely small initial surface tilts ($\beta \approx 0.01^\circ$) are capable of producing a self-reinforcing asymmetry through differential ablation.

To evaluate daily mean insolation under various orbital conditions, I used the equations of Ward (1974) (Equation 5.1), which express insolation (I/S) as a function of solar longitude and latitude in either hemisphere. To incorporate slope and aspect effects, the concept of equivalent latitude (Lee, 1964; Dingman, 2002) was applied following the approach of Williams et al. (2008) for Martian conditions. This method accounts for how slope gradient and aspect shift the effective angle of solar incidence, altering daily insolation. For example, a due-east slope shares a sunrise time with a flat surface but experiences an earlier sunset, reducing total energy input.

Here, δ is latitude, θ is obliquity, ω is the solar longitude of perihelion, v is the solar anomaly, η is the half-daylight angle, representing the angular duration (in radians) over which the Sun remains above the horizon at a given latitude and season. If β is the slope angle and α is the slope aspect, then the latitude of the parallel horizontal plane is the equivalent latitude δ_{eq} , which is given by:

$$\delta_{eq} = \text{asin}[\sin(\beta) \cdot \cos(\alpha) \cdot \cos(\delta) + \cos(\beta) \cdot \sin(\delta)]. \quad (7.6)$$

And the difference in longitude between the original slope and the parallel horizontal plane $\Delta\Omega$, is given by

$$\Delta\Omega = \text{atan} \left[\frac{\sin(\beta) \cdot \sin(\alpha)}{\cos \delta \cdot \cos \beta - \sin \delta \cdot \sin \beta \cdot \cos \alpha} \right]. \quad (7.7)$$

Solar noon then occurs at $-\Delta\Omega/\rho$, where ρ is the angular velocity of Mars' rotation and the times of sunrise (T_{RS}) and sunset (T_{SS}) are:

$$T_{RS} = - \frac{a \cos[-\tan(\delta_{eq}) \cdot \tan(\sigma) - \Delta\Omega]}{\rho}, \quad (7.8)$$

$$T_{SS} = + \frac{a \cos[-\tan(\delta_{eq}) \cdot \tan(\sigma) - \Delta\Omega]}{\rho}, \quad (7.9)$$

and

$$solar_t = local_time - (12.33 + solar_noon). \quad (7.10)$$

in the above expressions, σ is the declination, and 12.33 is the time in (terrestrial) hours of local noon for Mars (Williams et al., 2008). The hour angle is then

$$t = \rho * solar_t. \quad (7.11)$$

The extraterrestrial radiation flux on a sloping plane k_{ET} is given by:

$$k_{ET} = S \cdot (a/r)^2 \cdot (\cos(\delta_{eq}) \cdot \cos(\sigma) \cdot \cos(t + \Delta\Omega) + \sin(\sigma) \cdot \cos(\delta_{eq})), \quad (7.12)$$

where a is Mars' semi-major axis, ϵ is the eccentricity, and S is the solar constant at 1.52 AU, and S^* is the solar flux as a function of latitude then

$$S^* = S(a/r)^2, \quad (7.13)$$

$$r = a(1 - \epsilon^2)/(1 + \epsilon \cos(v)). \quad (7.14)$$

The daily total extraterrestrial radiation is given by integrating Eqn. 7.11 between sunrise (Eqn. 7.8) and sunset (Eqn. 7.9):

$$K_{ET} = \int_{T_{RS}}^{T_{SS}} k_{ET} \cdot dt. \quad (7.15)$$

Note that K_{ET} represents only the hypothetical direct-beam solar radiation reaching inclined surfaces and does not incorporate the effects of reflected shortwave radiation scattering, shadowing, or surface albedo. While these factors can affect glacier energy balance, they introduce complexities and a generally unconstrained range of combinations beyond the scope of my present analysis.

Three slope angles were selected for the insolation analysis: $\beta = 30^\circ$, 0° , and 0.01° , representing crater-wall slopes during active glaciation, a horizontal CCF surface, and the mean present-day CCF surface gradient, respectively. $\beta = 30^\circ$ is used to evaluate Scenario 1, in which CCF surface slopes are initiated during active accumulation as ice lobes develop along interior crater walls prior to fully blanketing the crater floor. Fresh and minimally modified impact craters retain interior wall slopes commonly exceeding 10° and approach $\sim 30^\circ$ (Kreslavsky and Head, 2018), and Amazonian degradation of mid-latitude crater walls is limited relative to earlier epochs. It is therefore reasonable to suggest that palaeo-CCFs formed within steep-sided craters, where insolation asymmetry between opposing crater walls could impose asymmetric mass balance during accumulation and initiate a persistent surface tilt.

The remaining values, $\beta = 0^\circ$ and $\beta = 0.01^\circ$, are used to assess Scenarios 2 and 3, in which CCF surface slopes develop during post-glacial retreat rather than during accumulation. A horizontal surface ($\beta = 0^\circ$) provides a reference case with no aspect-dependent insolation bias, while $\beta = 0.01^\circ$ reflects the typical present-day CCF surface gradient measured across the global dataset. Although the absolute insolation differences between these cases are necessarily small, the purpose is to evaluate the *direction* of energy-balance perturbations associated with poleward versus equatorward tilts, which—once a slope is initiated—could act as a feedback mechanism that progressively amplifies preferential ablation over prolonged post-glacial timescales.

The orbital parameters were selected to be consistent with those used by Madeleine et al. (2009), representing modelled conditions associated with mid-latitude glaciation thought to be necessary for emplacement of mid-latitude CCF deposits (e.g., Fastook and Head, 2014; Levy et al., 2014; Berman et al., 2015), with obliquity taken to be 35° and eccentricity set at 0.1, and perihelion coinciding with $L_s = 270^\circ$ for the northern hemisphere and $L_s = 90^\circ$ for the southern hemisphere. For post-glacial scenarios, present-day orbital configurations were used; I did not try to determine other specific orbital states (out of a potentially infinite number of such states) that put CCFs under interglacial conditions. Figure 7.30 presents modelled daily mean insolation integrated over one Martian year for different aspects and latitudes under $\beta = 30^\circ$ and glacial orbital conditions, with annual totals summarised in Table 7.1. Figure 7.31 and Table 7.2 show the equivalent results under present-day orbital conditions. In both orbital states, pole-facing slopes receive systematically lower annual-mean insolation than equator-facing slopes at all tested latitudes. This establishes that, for steep crater-wall slopes representative of the accumulation phase, aspect-dependent radiative forcing is strong and persistent, and is capable of driving lateral asymmetry in surface energy balance during active glaciation.

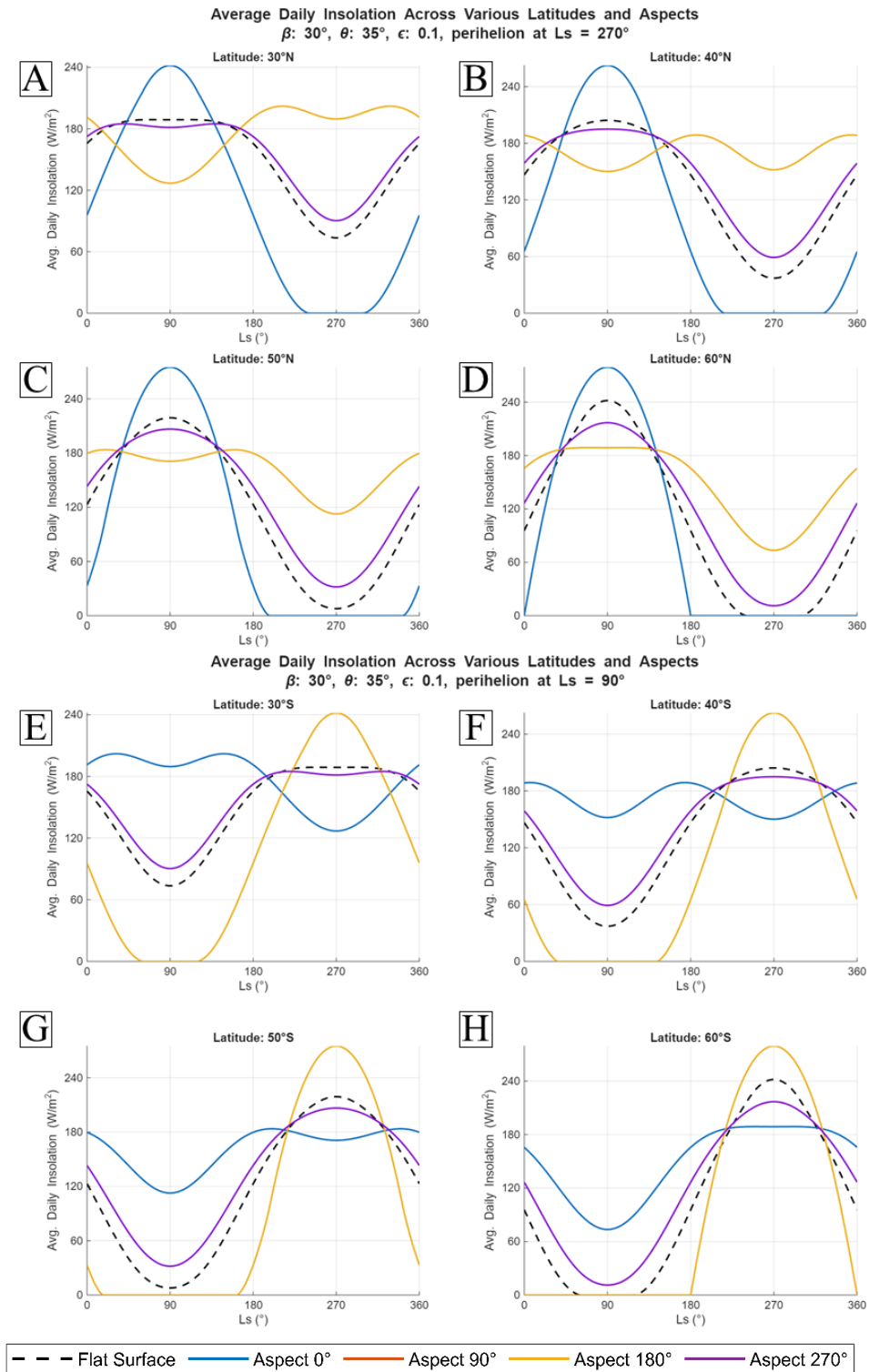


Figure 7.30. Average daily insolation over one L_s for varying aspects on 30° sloping crater walls under an intermediate-obliquity orbital model. Average daily insolation (W m^{-2}), integrated over a full Martian year ($L_s = 0\text{--}360^\circ$), shown for $\beta = 30^\circ$ at four cardinal aspects ($0^\circ, 90^\circ, 180^\circ, 270^\circ$) and for a range of latitudes in both hemispheres. Panels A–D show northern hemisphere latitudes of $30^\circ, 40^\circ, 50^\circ$, and 60° N, respectively, while panels E–H show southern hemisphere latitudes of $-30^\circ, -40^\circ, -50^\circ$, and -60° , respectively. The black dashed line in each panel shows the corresponding flat-surface insolation. The calculations use $\theta = 35^\circ$ and $\epsilon = 0.1$, with perihelion at $L_s = 270^\circ$ for A–D and $L_s = 90^\circ$ for E–H, following the configuration adopted by Madeleine et al. (2009). Annual totals are summarised in Table 7.1.

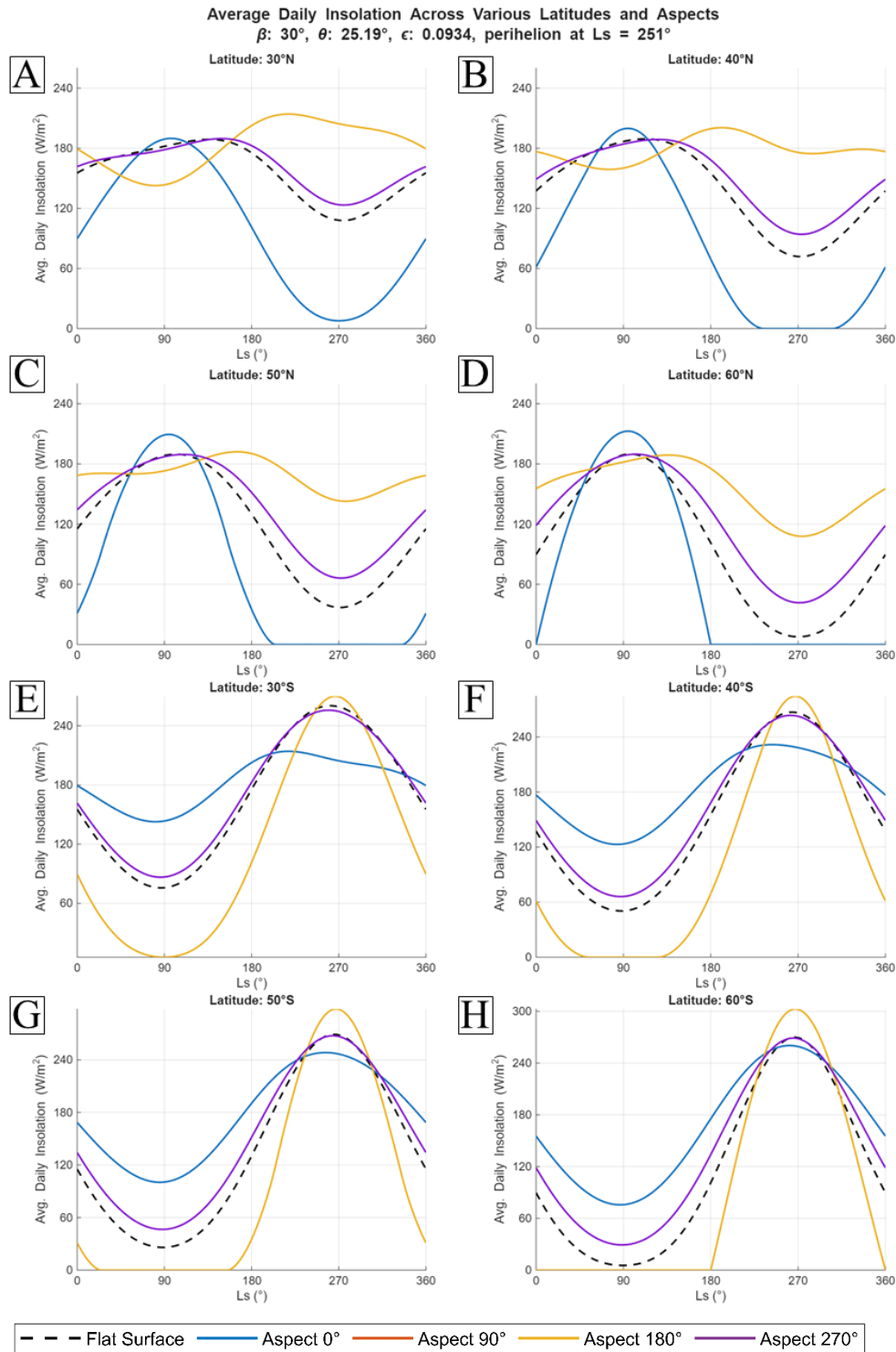


Figure 7.31. Average daily insolation over one L_s for varying aspects on 30° sloping crater walls under present-day orbital conditions. Average daily insolation ($W\ m^{-2}$), integrated over a full Martian year ($L_s = 0\text{--}360^\circ$), shown for $\beta = 30^\circ$ at four cardinal aspects (0° , 90° , 180° , 270°) and for the same set of northern (30° , 40° , 50° , 60° N; panels A–D) and southern (-30° , -40° , -50° , -60° ; panels E–H) hemisphere latitudes as in Figure 7.30. The calculations use the present-day orbital configuration ($\theta: 25.19^\circ$, $\epsilon: 0.09$, perihelion at $L_s = 251$). The black dashed line in each panel shows the corresponding flat-surface insolation. The calculations use the present-day orbital configuration ($\theta = 25.19^\circ$, $\epsilon = 0.09$, perihelion at $L_s = 251^\circ$). Annual totals are summarised in Table 7.2.

Table 7.1. Average daily insolation by latitude and aspect under intermediate obliquity

North Hemisphere					
Latitude (°)	0° (pole facing)	90° (east facing)	180° (equator facing)	270° (west facing)	Flat surface
(δ)					
30	173.25	152.60	103.33	152.60	146.82
40	168.22	141.31	96.69	141.31	131.77
50	159.25	129.27	93.10	129.27	115.70
60	146.82	117.93	91.90	117.93	103.33
South Hemisphere					
Latitude (°)	0° (equator facing)	90° (east facing)	180° (pole facing)	270° (west facing)	Flat surface
(δ)					
30	103.33	152.60	173.25	152.60	146.82
40	96.69	141.31	168.22	141.31	131.77
50	93.10	129.27	159.25	129.27	115.70
60	91.90	117.93	146.82	117.93	103.33

Average daily insolation over one L_s (W/m²) for different latitudes and aspects for the north and south hemisphere for an intermediate-obliquity orbital configuration used to represent mid-latitude glaciation

Table. 7.2. Average daily insolation by latitude and aspect under present day conditions

North Hemisphere					
Latitude (°)	0° (pole facing)	90° (east facing)	180° (equator facing)	270° (west facing)	Flat surface
(δ)					
30	104.10	152.97	173.23	152.97	147.26
40	97.52	141.80	168.36	141.80	132.34
50	93.97	129.86	159.55	129.86	116.39
60	91.90	117.93	146.82	117.93	103.33
South Hemisphere					
Latitude (°)	0° (equator facing)	90° (east facing)	180° (pole facing)	270° (west facing)	Flat surface
(δ)					
30	180.48	169.41	114.38	169.41	164.53
40	179.95	159.57	100.25	159.57	150.42
50	174.54	147.90	93.68	147.90	133.11
60	164.53	135.70	91.61	135.70	114.38

Average daily insolation over one L_s (W/m²) for different latitudes and aspects for the north and south hemisphere for Mars' current orbital configuration

To assess whether similar asymmetries could operate on the extremely gentle present-day CCF surfaces, the same calculations were repeated for $\beta = 0^\circ$ and $\beta = 0.01^\circ$. As expected, the absolute annual-mean insolation values for these two cases are nearly identical. The diagnostic quantity is therefore the difference between them, which isolates the first-order effect of introducing a very small surface tilt. Figure 7.32 presents the modelled difference in annual-mean insolation between $\beta = 0.01^\circ$ and $\beta = 0^\circ$ under glacial orbital conditions, with annual differences summarised in Table 7.3. Figure 7.33 and Table 7.4 show the equivalent results under present-day orbital conditions.

Under both orbital states, the largest insolation differences occur on pole-facing slopes, with equator-facing slopes showing the opposite sign. For glacial orbital parameters, the magnitude of this difference varies with latitude between ~ 0.01 and $\sim 0.02 \text{ W m}^{-2}$. Under present-day conditions, the difference is $\sim 0.02 \text{ W m}^{-2}$ in the northern hemisphere and varies between ~ 0.01 and $\sim 0.02 \text{ W m}^{-2}$ in the southern hemisphere. Although these absolute values are small, they represent a *systematic and persistent* energy-balance perturbation that acts in the same direction over each annual cycle. These results demonstrate that even an extremely gentle surface slope is sufficient to introduce a measurable and consistently signed insolation asymmetry. Once any small pole facing tilt exists, aspect-dependent radiative forcing will therefore act as positive feedback on the tilted surface (Table 7.3–7.4), promoting preferential ablation on the more strongly illuminated side and progressively amplifying the slope over long post-glacial timescales.

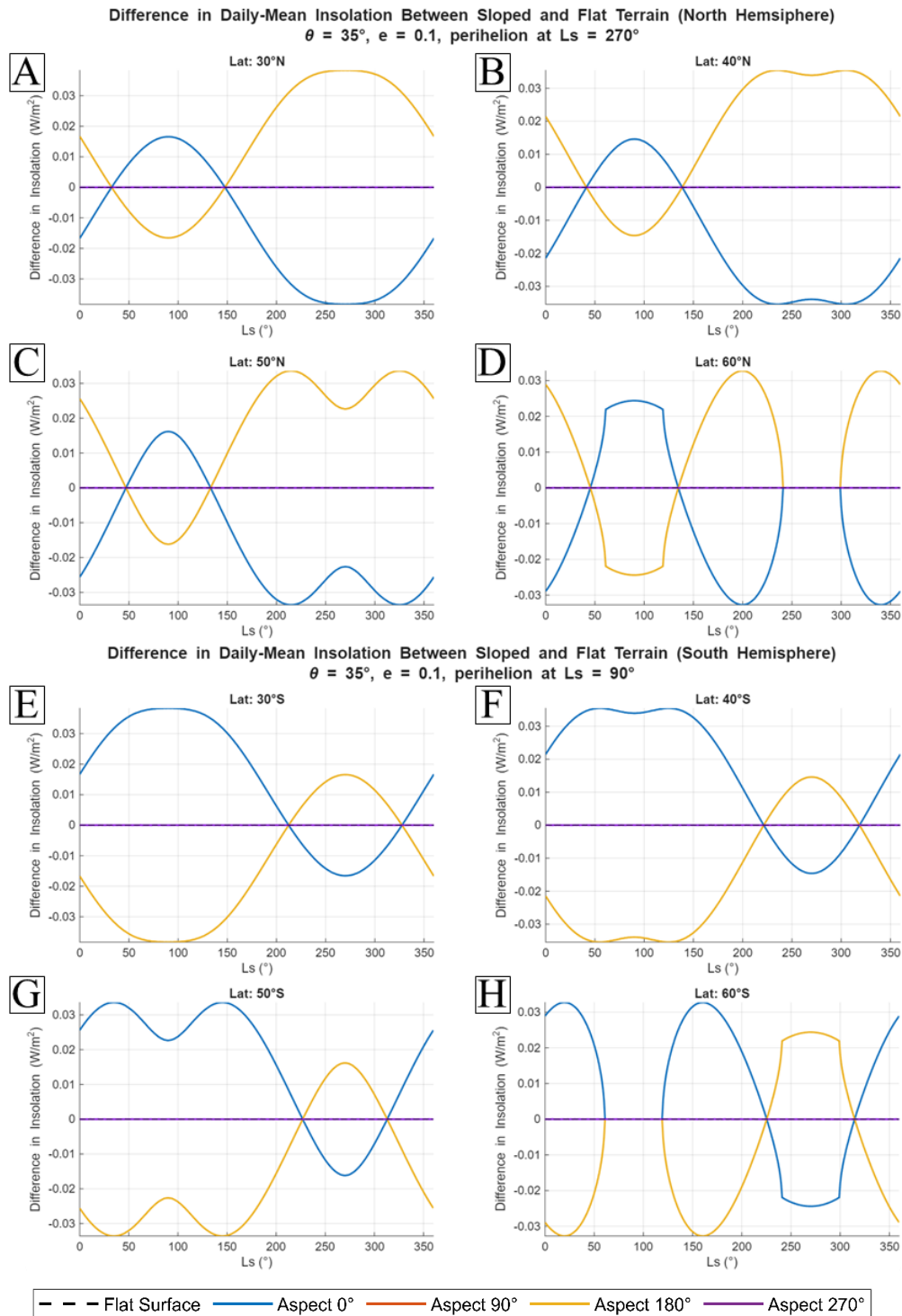


Figure 7.32. Difference in average daily insolation between nearly flat and very gently sloping CCF surfaces ($\beta = 0.01^\circ - \beta = 0^\circ$) for varying aspects over one L_s for an intermediate-obliquity model. Difference in average daily insolation ($\Delta W \text{ m}^{-2}$), integrated over a full Martian year ($L_s = 0-360^\circ$), shown for four cardinal aspects (0° , 90° , 180° , 270°) and for a range of latitudes in both hemispheres. Panels A–D show northern hemisphere latitudes of 30° , 40° , 50° , and 60° N, respectively, while panels E–H show southern hemisphere latitudes of -30° , -40° , -50° , and -60° , respectively. Positive values indicate greater insolation for $\beta = 0.01^\circ$ relative to a flat surface ($\beta = 0^\circ$). The calculations use $\theta = 35^\circ$ and $\epsilon = 0.1$, with perihelion at $L_s = 270^\circ$ for A–D and $L_s = 90^\circ$ for E–H, following the configuration adopted by Madeleine et al. (2009). Annual totals are summarised in Table 7.3

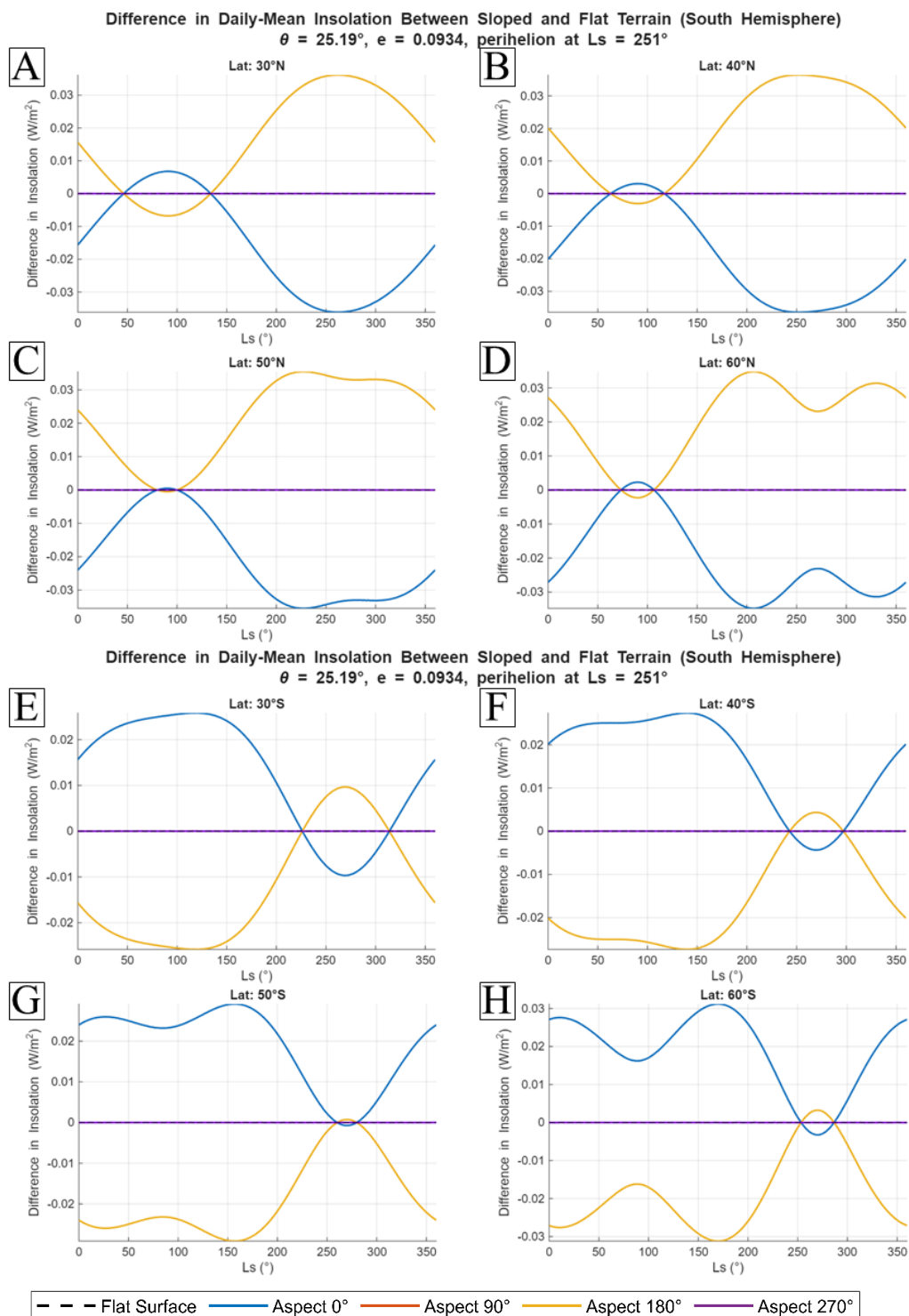


Figure 7.33. Difference in average daily insolation between nearly flat and very gently sloping CCF surfaces ($\beta = 0.01^\circ - \beta = 0^\circ$) for varying aspects over one L_s under present-day orbital configuration. Difference in average daily insolation ($\Delta W \text{ m}^{-2}$), integrated over a full Martian year ($L_s = 0-360^\circ$), shown for four cardinal aspects (0° , 90° , 180° , 270°) and for a range of latitudes in both hemispheres. Panels A–D show northern hemisphere latitudes of 30° , 40° , 50° , and 60° N, respectively, while panels E–H show southern hemisphere latitudes of -30° , -40° , -50° , and -60° , respectively. Positive values indicate greater insolation for $\beta = 0.01^\circ$ relative to a flat surface ($\beta = 0^\circ$). The calculations use the present-day orbital configuration ($\theta = 25.19^\circ$, $\varepsilon = 0.09$, perihelion at $L_s = 251^\circ$). Annual totals are summarised in Table 7.4

Table. 7.3. Difference in slope-induced insolation under intermediate obliquity

North Hemisphere				
Latitude (°)	0° (pole facing)	90° (east facing)	180° (equator facing)	270° (west facing)
(δ)				
30	-0.01	0.00	0.01	0.00
40	-0.02	0.00	0.02	0.00
50	-0.02	0.00	0.02	0.00
60	-0.01	0.00	0.01	0.00
South Hemisphere				
Latitude (°)	0° (equator facing)	90° (east facing)	180° (pole facing)	270° (west facing)
(δ)				
30	0.01	0.00	-0.01	0.00
40	0.02	0.00	-0.02	0.00
50	0.02	0.00	-0.02	0.00
60	0.01	0.00	-0.01	0.00
Difference in average daily insolation (W/m^2) for Different Latitudes and Aspects for the North and South Hemisphere for an intermediate-obliquity orbital configuration used to represent mid-latitude glaciation				

Table. 7.4. Average daily slope-induced insolation under present-day conditions

North Hemisphere				
Latitude (°)	0° (pole facing)	90° (east facing)	180° (equator facing)	270° (west facing)
(δ)				
30	-0.02	0.00	0.02	0.00
40	-0.02	0.00	0.02	0.00
50	-0.02	0.00	0.02	0.00
60	-0.02	0.00	0.02	0.00
South Hemisphere				
Latitude (°)	0° (equator facing)	90° (east facing)	180° (pole facing)	270° (west facing)
(δ)				
30	0.01	0.00	-0.01	0.00
40	0.02	0.00	-0.02	0.00
50	0.02	0.00	-0.02	0.00
60	0.02	0.00	-0.02	0.00
Difference in average daily insolation (W/m^2) for Different Latitudes and Aspects for the North and South Hemisphere under present-day conditions				

Two physically plausible mechanisms follow from these results. First, if slopes formed primarily through asymmetric mass balance during active glaciation (Scenario 1), higher insolation on equator-facing interior crater walls (Tables 7.1–7.2) would have enhanced ablation of equator-facing ice lobes, favouring preferential thickening on pole-facing sectors and promoting the development of a poleward surface tilt. In this case, both the cumulative annual insolation contrast (Tables 7.1–7.2) and the seasonal enhancement of equator-facing insolation (Figure 7.34A) would contribute to establishing the asymmetry during accumulation. Second, reflected shortwave radiation from crater walls onto the CCF surface provides an additional source of lateral energy asymmetry (Figure 7.34B). Equator-facing crater walls receive higher insolation under tested conditions (Tables 7.1–7.2) and may reflect additional solar energy onto adjacent parts of the CCF surface, promoting localised ablation. This process can operate both during active glaciation (Scenario 1, Figure 7.26) and during subsequent retreat phases (Scenarios 2 and 3, Figure 7.27 and 7.28) and would reinforce any pre-existing surface tilt or help generate one during downwasting.

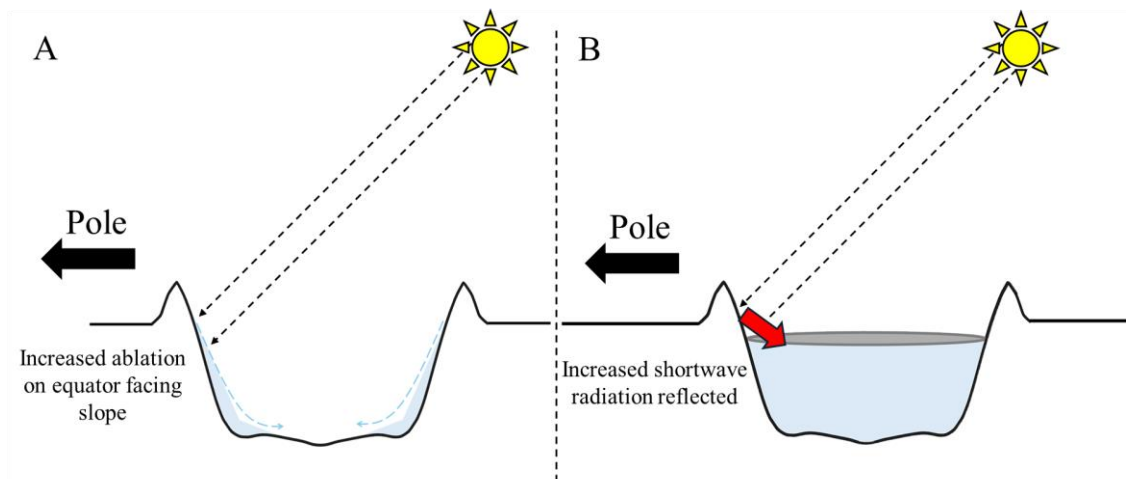


Figure 7.34. Conceptual schematic for insolation-driven asymmetry in CCF surface slope formation.

(A) During periods of enhanced obliquity, increased insolation on equator-facing interior crater walls promotes greater ablation on equator-facing CCF ice lobes, driving the development of pole-facing surface slopes. (B) Reflected shortwave radiation from equator-facing crater walls may enhance localised ablation of adjacent CCF surfaces. This process may operate during both glacial accumulation and interglacial

Tested insolation results for varying obliquities, latitudes and slopes (Figures 7.30–7.33; Tables 7.1–7.4) provide consistent support for both mechanisms (Figure 7.34). Equator-facing interior crater-wall slopes receive greater total annual insolation than pole-facing slopes across all tested orbital configurations, latitudes, and hemispheres within the CCF latitude range. Although seasonal insolation distributions vary with obliquity and latitude, cumulative annual energy input remains systematically higher on equator-facing slopes under all modelled conditions.

During active glaciation, this asymmetry is consistent with the development of poleward surface slopes through preferential ablation of equator-facing ice lobes and/or enhanced accumulation in pole-facing, more weakly irradiated sectors of the crater interior (Mechanism 1). During retreat, even extremely small surface tilts introduce a persistent energy-balance asymmetry ($\beta = 0.01^\circ$ experiments), implying a positive feedback in which the more strongly illuminated side ablates slightly faster, progressively amplifying the slope. Once CCF surfaces fall below the crater rim (i.e., $d_a > 0$), partial exposure of interior crater walls would further enhance this feedback through reflected shortwave radiation onto the CCF surface, promoting continued equator-facing ablation until mass balance stabilises under an insulating debris lag, allowing a poleward surface slope to be preserved (Mechanism 2)

7.4.3. Possible drivers of the east-facing bias

In addition to the dominant poleward bias in surface aspect, CCF surfaces in both hemispheres exhibit a secondary but systematic preference for east-facing orientations, here defined as aspects within $\pm 45^\circ$ of due east (i.e., spanning northeast to southeast). This eastward bias is strongest in the northern hemisphere, where it is expressed across a broad range of orientations, and is present but subordinate to the poleward bias in the southern hemisphere. The biases indicate an east-west asymmetry in CCF evolution. Such asymmetry is not unique to CCFs, as eastward skew has been identified in GLFs aspects (Milliken et al., 2003; Souness et al., 2012). As reviewed in Section. 7.1.2, no definitive mechanism has yet been proposed to explain these biases (Souness et al., 2012).

Figure 7.35 illustrates the spatial distribution of east-facing ($\pm 45^\circ$ of east) CCFs identified in this chapter together with the east-facing GLFs reported by Souness et al. (2012). In both datasets, east-facing VFFs are distributed broadly across the mid-latitudes in both hemispheres, with no evidence for regional clustering or confinement to specific physiographic provinces. This spatial dispersion argues against explanations tied to local or regional controls, such as specific topographic settings, wind corridors, or geographically restricted climatic anomalies. Instead, the widespread and hemispherically recurrent nature of the signal indicates that the east-facing bias most likely reflects a globally acting control on surface energy balance or atmospheric forcing.

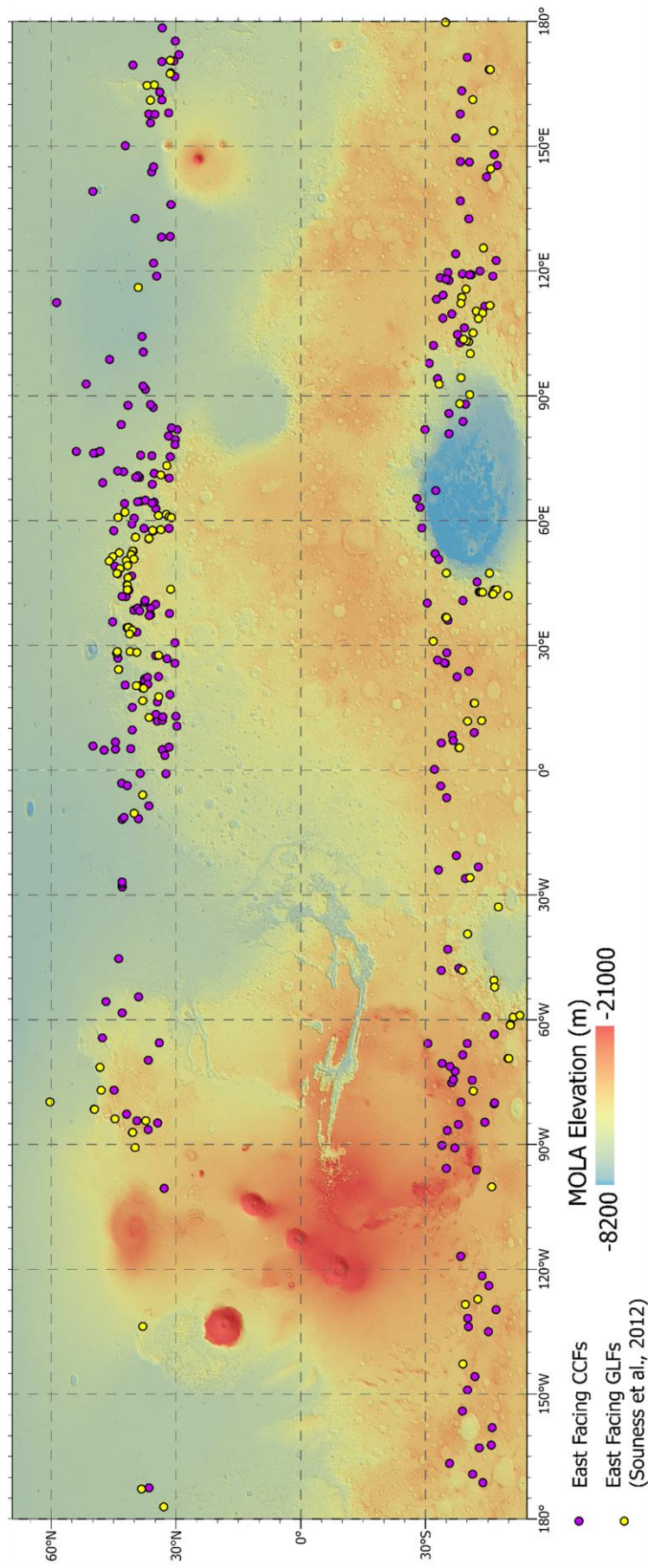


Figure 7.35. Spatial distribution of east-facing CCFs and GLFs. Map showing east-facing CCFs (this study) and east-facing GLFs from Souness et al. (2012), overlaid on MOLA gridded topography and MOLA-derived hillshade. East-facing features are within aspect bins of 67.5° and 112.5° . Both CCFs and GLFs are broadly scattered across the mid-latitudes in both hemispheres, with no apparent regional clustering.

A plausible initial question is whether the mechanism behind asymmetry involves diurnal insolation. In the simplest, ideal situation with a circular host crater and horizontal CCF, due to the symmetrical nature of the Sun's diurnal arc, east- and west-facing slopes, asymmetry cannot arise in the east-west direction in the spatial pattern of total daily insolation (Dingman, 2002; Kreslavsky et al., 2008; Williams et al., 2008; Figures 7.30–7.33; Tables 7.1–7.4); thus, it seems that eastward or westward slope cannot develop from insolation effects on mass balance or preservation.

However, in terrestrial glaciology, east-facing biases in glacier surface aspect have been linked to diurnal cycles in temperature (Garg et al., 2017) and cloud cover (e.g., Evans, 1977, 2006, 2020). This motivates me to query whether/how analogous mechanisms could operate on Mars, by considering diurnal radiative/thermal effects in more detail.

7.4.3.1. Asymmetry in East-West facing slope temperatures

Here, I argue that thermal asymmetry within crater interiors can arise from the interplay between diurnal solar geometry, topographic shading, and thermal inertia. While east- and west-facing slopes with the same topographic gradient and aspectual deviation from north receive the same total longwave radiation over the course of a Martian sol, the timing and angle at which this radiation is delivered differ, creating different thermal responses in time (during the sol). This affects both ice-free crater walls and the ice lobes during CCF formation (Scenario 1) or CCF surface retreat (Scenario 2 and 3). In the following example, I assume a symmetrical crater (with equal wall slopes everywhere on the circumference) and a horizontal CCF/ice surface; however, these principles apply equally to the individual ice lobes during the Scenario 1.

First, consider the illumination and thermal history of the crater walls. In the morning, the sun rising in the east casts semi-direct insolation onto east-facing crater walls (i.e. western inner flank of the crater) (Figure 7.36A). These slopes become illuminated when the sun rises above the opposing (eastern) rim; before this, they are shaded. In contrast, west-facing crater walls (on the eastern flank) remain in shadow for a longer time in the morning. They begin to receive direct insolation later, once solar elevation exceeds an angle similar to their slope angles (Figure 7.36B). In the early part of the morning, a high angle of incidence leads to low radiation receipt on these slopes (Figure 7.36B).

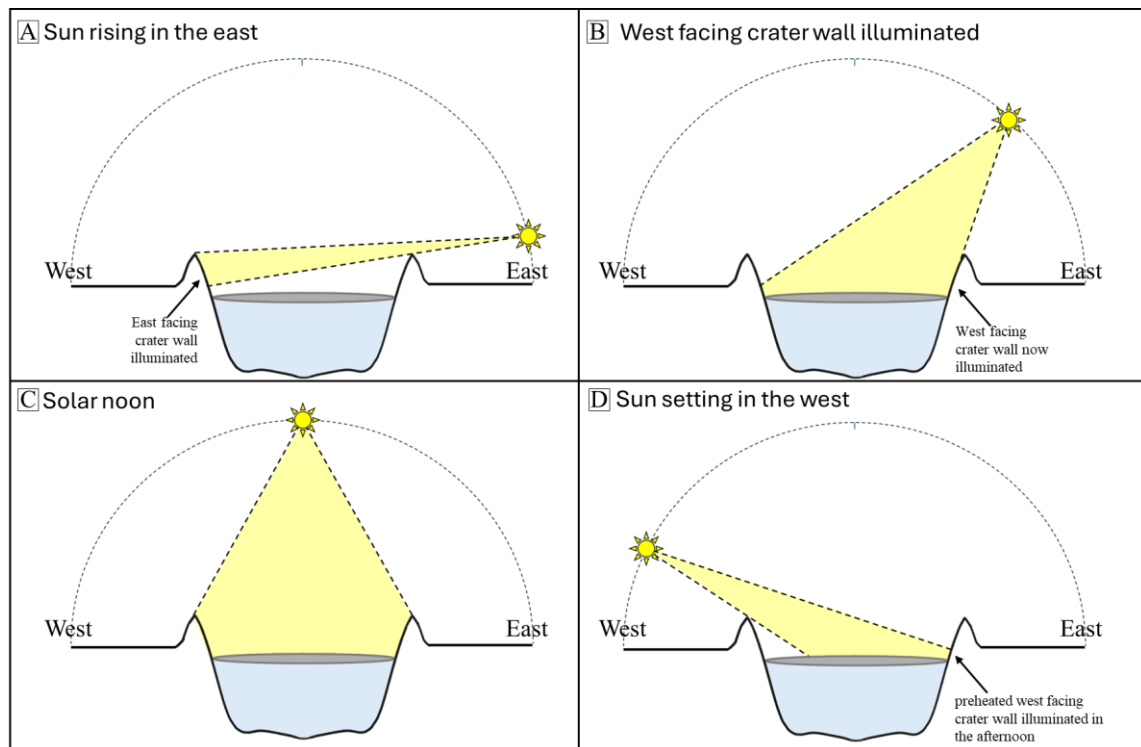


Figure 7.36. Conceptual model of diurnal illumination of east- and west-facing crater interior walls. Schematic diagrams illustrate how the geometry of a simple bowl-shaped crater produces systematic east–west asymmetry in solar illumination over one sol. (A) In the early morning, the rising Sun illuminates the east-facing interior wall and the west-facing wall remains in shadow behind the eastern rim. (B) As solar elevation increases, the west-facing wall begins to receive direct sunlight only once the Sun rises sufficiently above the opposing rim. (C) Solar noon, both crater walls are illuminated. (D) In the afternoon, the west-facing wall receives increasingly direct, high-incidence-angle radiation, while the east-facing wall progressively loses illumination and eventually falls into shadow cast by the western rim. The sequence illustrates the intrinsic diurnal asymmetry in energy receipt between east- and west-facing crater slopes.

During the late morning and across midday (7.36C) both slopes are illuminated, then as the solar elevation continues to increase and then decrease past solar noon, the west-facing slopes, which are already mildly warmed, receive more direct, high-angle radiation leading to intense, focused heating. Heating on them is the most intense in the mid-afternoon when the angle of incidence reaches 90° (Figure 7.36D). Meanwhile, east-facing slopes, now past the point of perpendicular irradiation, receive less and less sunlight; eventually they fall into shadow from the western crater rim in the mid/late afternoon, when the west-facing slopes are still lit (Figure 7.36D).

Now consider the specific geometry – and the corresponding time of day – when west or east facing crater walls are most strongly heated by solar radiation at an angle of incidence of 90° . At this point, the walls re-radiate most strongly onto their adjacent CCF/ice surfaces, to enhance ablation on those surfaces. For east-facing walls, this point occurs in the mid-morning (Figure 7.37A), by which time they have been illuminated and heated for a relatively short time (from early morning to mid-morning); for west-facing walls, the point occurs in the mid-afternoon (Figure 7.36C, 7.37), by which time they have been illuminated and heated for a relatively long time (from mid-morning to mid-afternoon; Figure

7.36). Because the west-facing walls are warmer than the east-facing walls at the respective points, an asymmetry develops in the *ablation enhancement* on the CCF surface due to thermal re-radiation from adjacent crater walls: it is higher on eastern edge/part of the CCF than on the western edge/part of the CCF (Figure 7.37B). Over time, this causes the CCF surface to tilt and curve to become eastward-facing. In principle, this mechanism can operate during both mid-latitude glaciation and post glacial periods. Note that, at the two respective time points, the incidence angle of *direct* solar radiation on the (initially horizontal) CCF surface is equal, so this radiation component does not cause an asymmetry in ablation. While this diurnal heating asymmetry can produce eastward bias in CCF aspects, it does not explain why the bias is more strongly expressed in the northern hemisphere. Other factors must modulate the bias to cause the observed hemispheric differences.

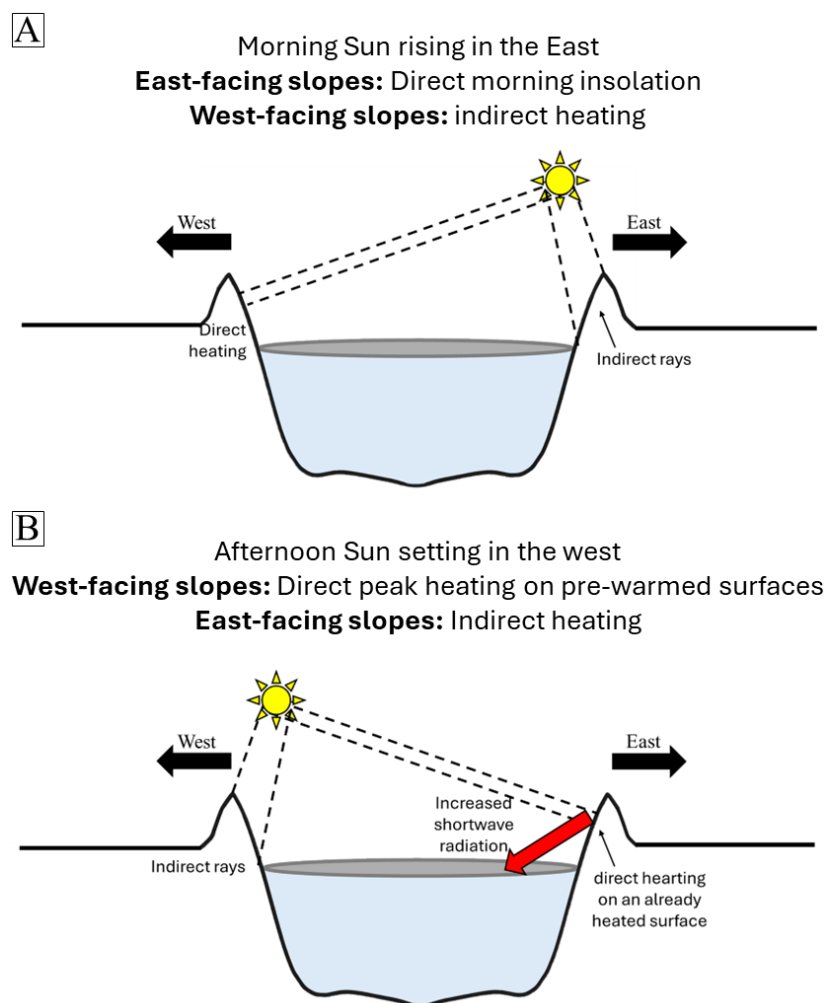


Figure 7.37. Schematic of diurnal heating asymmetry and its potential influence on east-facing slope formation. West-facing crater walls receive indirect heating in the morning (A) and direct insolation during the afternoon solar peak (B), reaching higher peak temperatures than east-facing slopes due to cumulative surface and atmospheric pre-heating throughout the morning (A). This thermal imbalance can enhance local ablation directly via thermal radiation, or indirectly via reflected shortwave radiation onto opposing east-facing CCF surfaces (red arrow). While this mechanism operates globally, the more pronounced east-facing slope signal observed in the Northern Hemisphere suggests that this diurnal process may be modulated by additional hemispheric-scale climatic or topographic controls.

7.4.3.2. *Diurnal cloud cover*

In terrestrial glaciology, subtle east–west asymmetries in glacier orientation have been linked to diurnal variations in cloud cover, which modulate surface energy balance (Evans, 1977, 2006, 2011; Evans and Cox, 2017). In dry mid-latitude environments, morning cloudiness can reduce solar input on east-facing slopes, whereas in tropical or highly continental settings, convective afternoon cloud cover can suppress radiation on west-facing slopes. Although these effects are typically secondary to dominant north–south insolation gradients, they have been shown to influence glacier orientation under specific environmental regimes. These terrestrial observations provide a conceptual analogue for evaluating whether diurnal atmospheric cycles *could*, in principle, produce comparable aspect biases in CCF and VFF surfaces on Mars.

On Mars, the existence of Earth-like diurnal convection cycles is not well established (both spatially or temporally). However, recent observations confirm that the present-day aphelion cloud belt exhibits a consistent diurnal cycle, with cloud cover thickest during the early to mid-morning and thinning through the afternoon (Szantai et al., 2021; Wolff et al., 2022; Fernando et al., 2024; Figure 7.38). Although the timing of this morning maximum is robust, the spatial extent and optical thickness of the clouds vary seasonally and interannually, often intensifying during dusty periods when circulation enhances cloud formation and persistence.

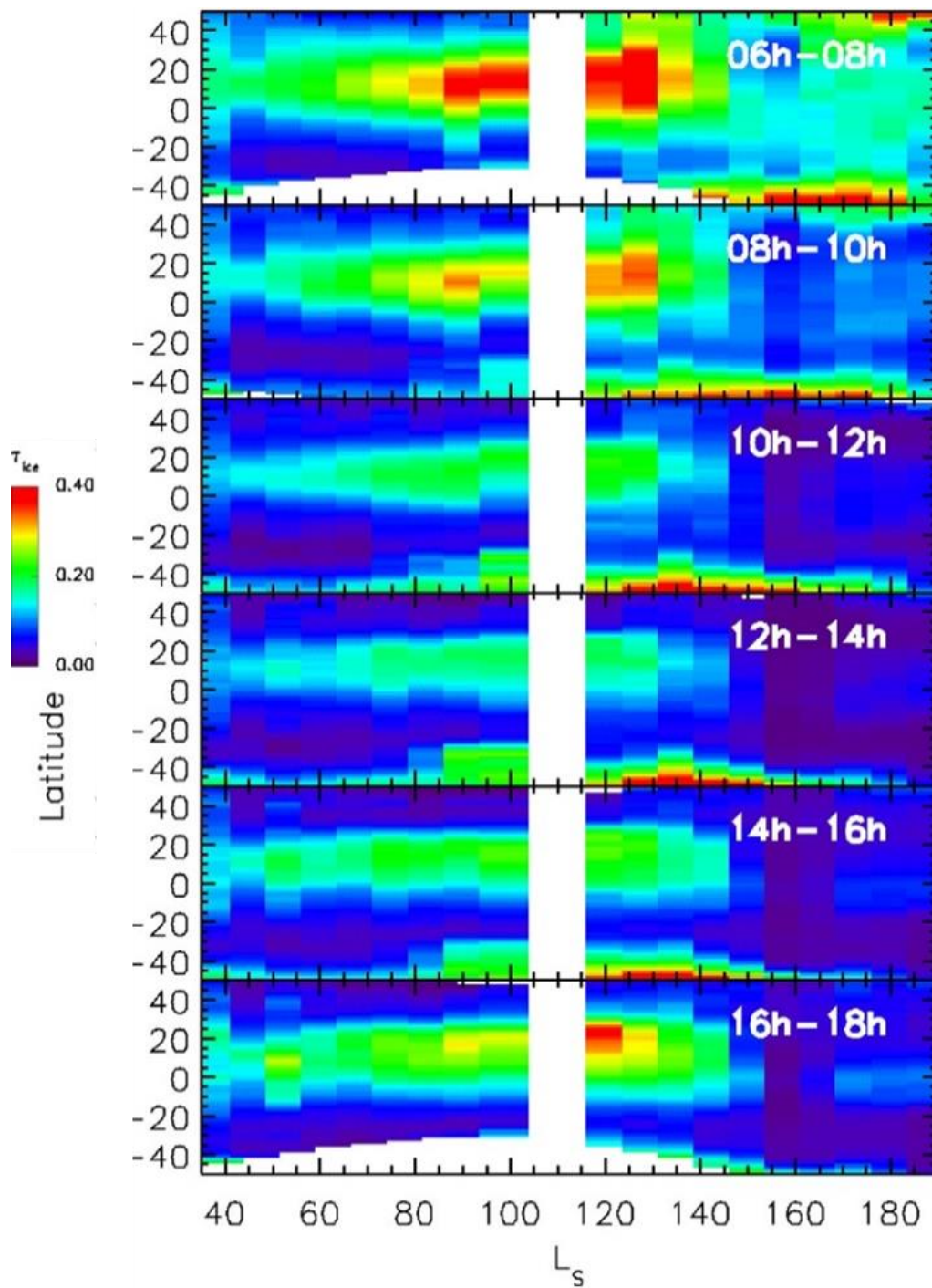


Figure 7.38. Diurnal variability of the aphelion cloud belt in Mars Year 36. Zonal-mean water-ice cloud optical depth (τ_{ice}) derived from Emirates eXploration Imager (EXI) observations for $L_s \approx 35^\circ$ – 190° , shown for morning (left) and afternoon (right) local times. Each panel displays τ_{ice} as a function of latitude (-40° to $+40^\circ$) and season, with colour indicating extinction optical depth. The data have been regridded and reformatted here to emphasise seasonal progression (time decreasing downward within each panel). The cloud belt remains largely confined to low and subtropical latitudes, with only limited penetration into the mid-latitudes in either hemisphere, and τ_{ice} is systematically higher in the morning than in the afternoon across the same latitude bands, demonstrating a pronounced diurnal cycle in cloud opacity. Redrawn and modified from Wolff et al. (2021).

These recent observations align with long-standing theoretical frameworks established by early GCM simulations by Richardson et al. (2002) using a Mars Global Climate Model to predict water-ice cloud distributions across local times (00:00, 06:00, 12:00, and 18:00), finding that cloud optical thickness was typically lowest before dawn and increased steadily through the day. Building on this, Madeleine et al. (2012) demonstrated using a fully radiatively active GCM that cloud formation over regions such as the Tharsis plateau peaks during the early morning hours. Their simulations showed substantial low-altitude cloud concentrations between local times ~00:00 to ~08:00, accompanied by significant thermal influences: cooling of the lower atmosphere, warming aloft, and the strengthening of diurnal thermal tides. These results underscored the capacity of even tenuous ice clouds to modulate near-surface temperatures through both direct radiative forcing and indirect dynamical feedback.

For past climates, GCM simulations of intermediate-obliquity conditions predict the formation of a dense mid-latitude cloud belt over the northern hemisphere during glacial intervals (Madeleine et al., 2009; Figure 7.39), with optical thicknesses up to two orders of magnitude greater than those of the present-day aphelion cloud belt. Although these models do not resolve the diurnal behaviour of this palaeo-cloud belt, it is plausible that such a system would also have exhibited systematic daily variations in cloud thickness and radiative forcing, broadly analogous to those observed today.

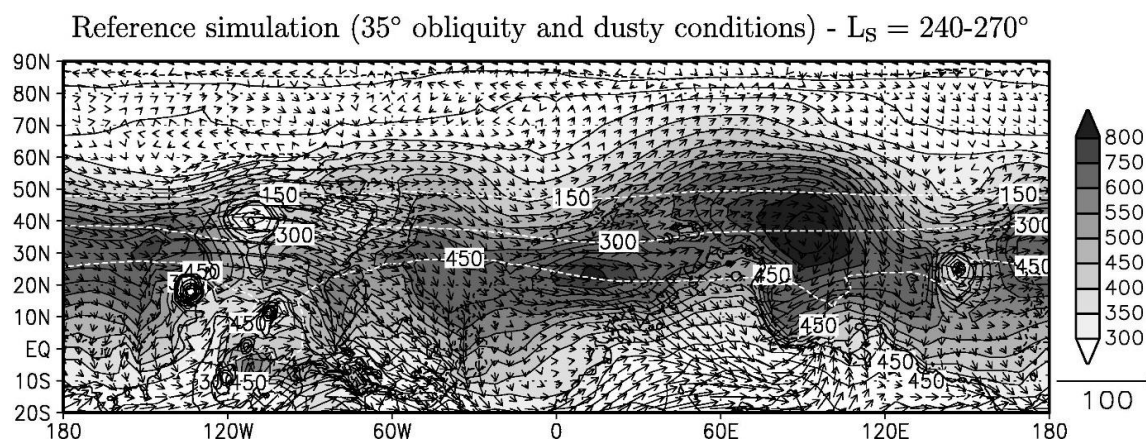


Figure 7.39. Predicted mid-latitude cloud belt in the northern hemisphere under intermediate obliquity. Average cloud ice content (shaded; pr. μm) and horizontal wind field at the 5.6 km level (m s^{-1}) predicted by the Global Climate Model of Madeleine et al. (2009) for an obliquity of 35° during northern summer ($L_s = 240\text{--}270^\circ$). White contours denote the column-integrated water vapour abundance (pr. μm). The map spans ~20°S to 90°N and illustrates the development of a mid-latitude cloud belt and associated circulation patterns under this orbital configuration. Redrawn from Madeleine et al. (2009).

Any application of diurnal cloud-modulated energy balance to CCF evolution must therefore remain speculative. If diurnally varying cloudiness or atmospheric opacity did occur under certain past Martian climate states, such effects could, in principle, operate during both active glacial phases and interglacial periods. During active accumulation, reduced morning insolation on east-facing crater walls could suppress ablation of accumulating ice, while enhanced afternoon insolation on west-facing walls could promote greater ablation, potentially favouring net preservation on east-facing crater interiors.

Alternatively, during interglacial periods or late-stage glaciation when CCF surfaces lie below the crater rim, morning cloudiness could limit both longwave emission and shortwave reflections from eastern crater walls onto the CCF surface, whereas clearer afternoon conditions could increase reflected energy from western walls (Figure 7.40). In all cases, such effects would act only as second-order modifiers superimposed on the dominant latitudinal insolation control, and their geomorphic significance for CCFs remains to be tested by future climate and ice-evolution modelling.

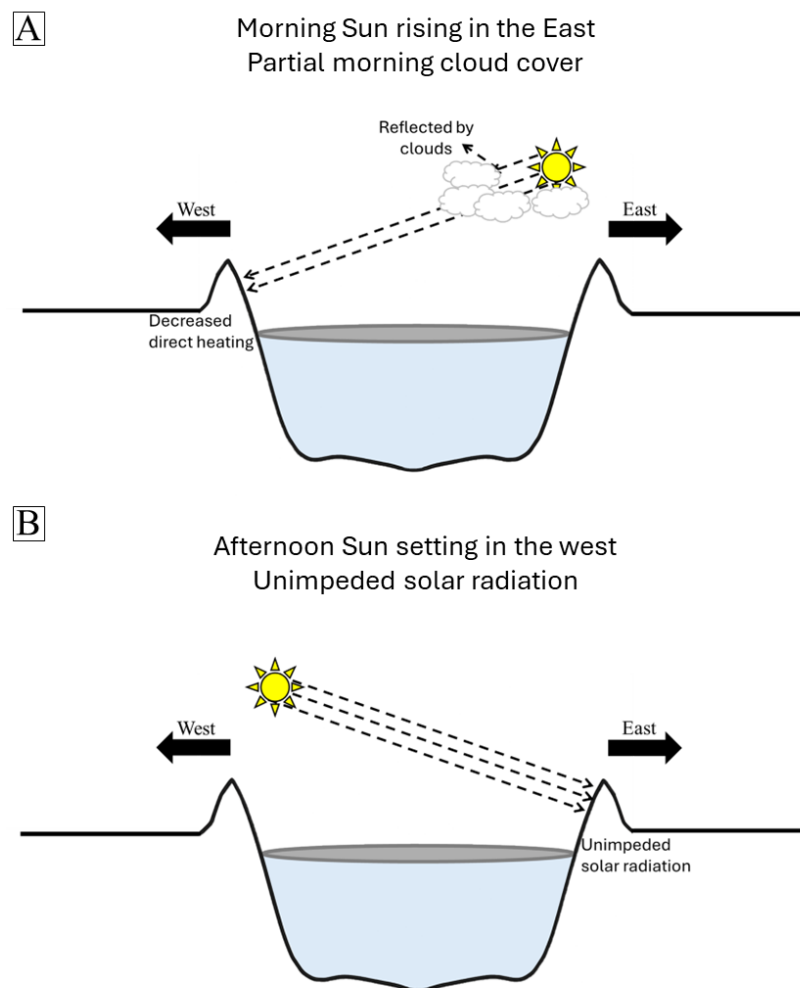


Figure 7.40. Hypothesised influence of diurnal cloud cycles on the formation of CCF east-facing surface slopes. (A) Morning: Increased cloud cover (characteristic of the aphelion cloud belt) attenuates incoming insolation, sheltering east-facing slopes from early-day heating. (B) Afternoon: Dissipation of cloud cover allows for unimpeded solar radiation on west-facing slopes. This diurnal asymmetry—perhaps exacerbated by afternoon heating of pre-warmed surfaces (Figure 7.39)—drives increased ablation on west-facing walls. This mechanism may influence CCF development either through direct ablation of ice lobes on crater walls (Scenario 1) or via enhanced shortwave radiation reflected onto opposing east-facing surfaces (Scenarios 2 and 3), contributing to east facing aspectual bias.

7.4.4. Differences in north–south aspect distribution

CCF aspect distributions between the Northern and southern hemispheres are markedly different, despite both regions exhibiting similar low slope magnitudes and a shared poleward and eastward bias. In the northern hemisphere, CCF aspects display a nearly concentric distribution, with no statistically significant clustering (Figure 7.16A, 7.18). By contrast, southern hemisphere CCF surface slopes exhibit a concisely pole-facing distribution (Figure 7.16B, 7.19). A similar hemispheric difference in aspect distribution is observed for GLFs (Souness et al., 2012, Figure 7.26B, E), suggesting this contrast may reflect broader differences in glacial processes or environmental conditions rather than CCF specific.

Three factors may be responsible for this hemispheric difference. The first is asymmetric accumulation (scenario 1) where higher accumulation rates in the northern mid-latitudes allowed CCF slopes to form without a strong aspect preference, whereas the clustered southern pattern required preferential accumulation on pole-facing slopes under more climatically constrained conditions. The second is asymmetric ablation in scenario 2 and 3, where the dispersed northern aspects imply that local factors and/or crater-scale microclimates exerted greater influence on slope evolution than large-scale climatic controls, necessitating an as-yet unidentified hemispheric mechanism that moderated ablation in the north relative to the south, potentially during active glaciation, retreat, or under present-day conditions. The third is the pronounced hemispheric difference in elevation, with the northern mid-latitudes characterised by low-relief terrain and the south by rougher, higher-relief landscapes (Milliken et al., 2003; Tanka et al., 2014).

Terrestrial studies show that lower altitudes typically constrain glaciers to preferred aspects, whereas higher altitudes produce more diverse orientations (Evans, 1977; Evans, 2006; Evans, 2020). However, this terrestrial analogy does not transfer straightforwardly to Mars, as Mars' thin atmosphere lacks such a strong near-surface lapse rate compared to Earth, and thus, surface energy balance is instead dominated by radiative rather than atmospheric controls. Elevation therefore is not expected to impose the same first-order climatic constraint on ice stability. Consequently, the pronounced elevation contrast between the northern lowlands and southern highlands cannot, by itself, explain why the southern hemisphere exhibits more tightly clustered, pole-facing CCF slopes while the north shows a more dispersed pattern. In the following, I explain why hemispheric differences in accumulation are the most likely explanation and briefly discuss the possible role of hemispheric asymmetries in ablation.

7.4.4.1. Evidence for increased northern hemisphere accumulation

The observed difference in CCF hemispheric aspect patterns (Figure 7.16) suggests that additional climatic or glaciological factors have shaped hemispheric differences in CCF slope orientation. The broad aspect distribution in the northern hemisphere, contrasted with the consistently poleward-oriented

slopes in the south, may indicate fundamentally different accumulation regimes occurring across the two hemispheres.

Evidence from previous studies supports the idea that widespread, slope-independent accumulation can explain concentric surface slope patterns. Dickson et al. (2012) proposed that such concentric orientations in high mid-latitudes result from regional-scale ice accumulation, occurring irrespective of slope aspect. However, the findings presented in this chapter show that this broad distribution is present only in the northern hemisphere.

Climate models further indicate that hemispheric asymmetries in ice accumulation are an expected outcome of orbital forcing. GCM simulations (e.g., Madeleine et al., 2009) have modelled that under intermediate-obliquity conditions with increased dust opacity, ice can accumulate extensively across the northern mid-latitudes (Figure 7.41A). However, when identical conditions are modelled with a different solar longitude of perihelion, accumulation occurs in the southern hemisphere but is markedly less widespread (Figure 7.41B). These simulations suggest that the northern hemisphere—under these specific conditions—is prone to more extensive accumulation in comparison to the south. However, as discussed in Chapter 4, these simulations represent a snapshot of orbital conditions. It is therefore likely that certain orbital configurations result in the south hemisphere receiving greater or more extensive accumulation than the north.

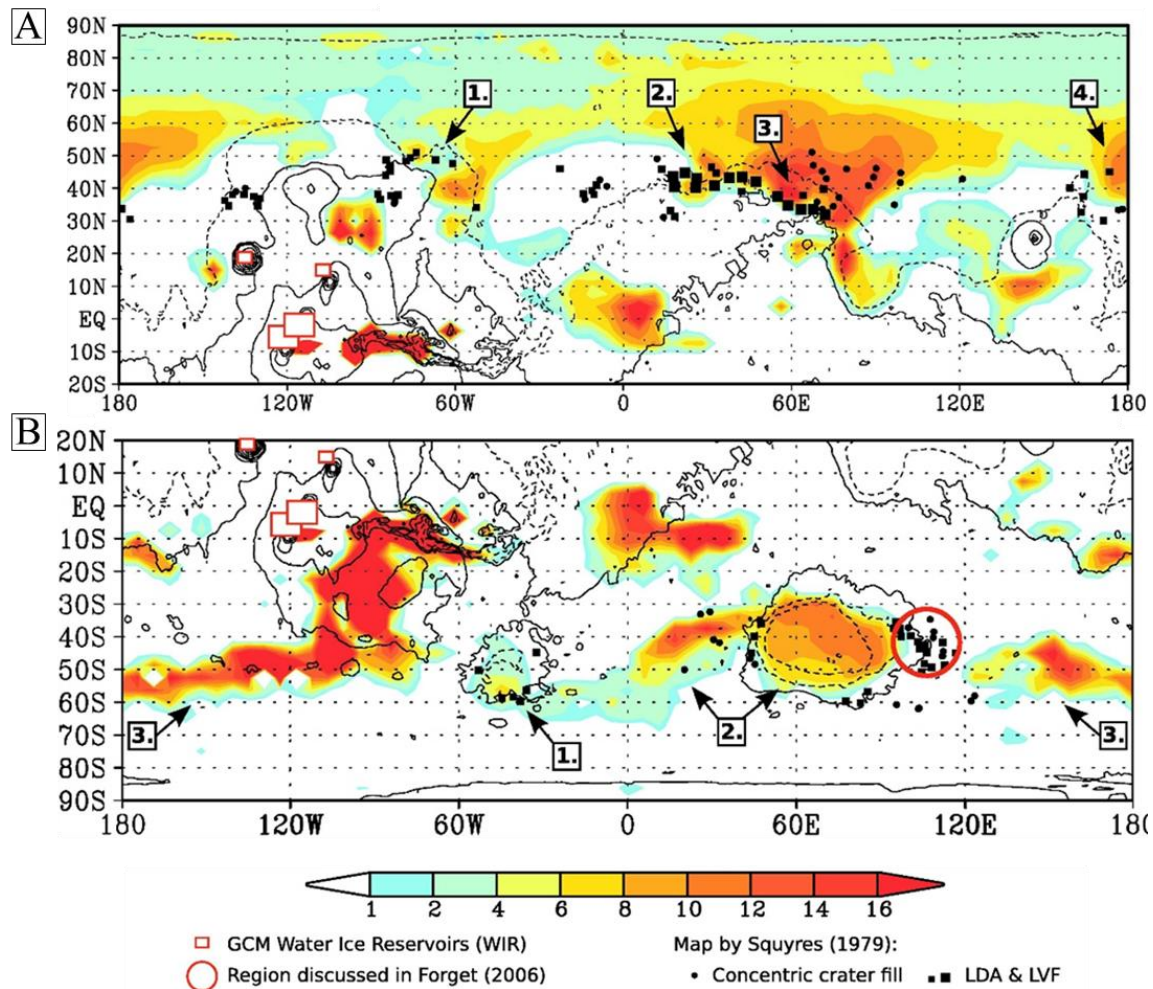


Figure 7.41. Mid-latitude net ice accumulation predicted by GCM simulations. Net ice accumulation rate (mm yr⁻¹) predicted by the Global Climate Model simulations of Madeleine et al. (2009) for an θ of 35°, eccentricity of 0.1, and dust opacity of 2.5, with perihelion at $L_s = 270^\circ$ for the Northern Hemisphere (A) and $L_s = 90^\circ$ for the Southern Hemisphere (B). Both panels are superposed on the mapped distributions of VFFs compiled by Squyres (1979). In (A), labelled regions are: (1) Tempe Terra, (2) Deuteronilus Mensae, (3) Nilosyrtris Mensae, and (4) Phlegra Montes. In (B), labelled regions are: (1) Argyre Planitia, (2) western Hellas, (3) Terra Sirenum, and (4) Terra Cimmeria. Redrawn from Madeleine et al. (2009).

Morphological evidence provides tentative support for greater ice accumulation in the northern hemisphere. At present, the only documented examples indicating higher palaeo-CCF surface elevations and/or greater ice mass are confined to the northern hemisphere, notably in Deuteronilus Mensae (Levy et al., 2010) and Phlegra Montes (Dickson et al., 2010), although this inference is necessarily based on a very small sample ($n = 2$) and may partly reflect preferential observational or reporting bias. Independent support comes from the results in this Thesis: northern hemisphere CCFs exhibit smaller d_a values (Chapter 5) and higher fill fractions (Chapter 6) than their southern counterparts, implying that they remain, on average, closer to rim-full states today. This morphological evidence, together with the d_a and volume results of this Thesis and the broader concentration of GLFs

in the north (Souness et al., 2012), is consistent with a history of more extensive, more sustained or perhaps more recent ice accumulation in the northern mid-latitudes.

If northern hemisphere CCFs experienced greater ice accumulation, perhaps they once filled their craters to a greater extent (Scenario 2) and may have overflowed their rims (Scenario 3). Then as ice sublimated and CCF apparent depth increased, local-scale (or microclimate) factors played a larger role in influencing slope orientation during CCF evolution to their current extent. In contrast, the comparatively reduced accumulation in the southern hemisphere likely left poleward and eastward climatic controls as the dominant influences on slope development, as discussed above. In this sense the southern hemisphere is more likely restricted to Scenario 1 or Scenario 2. If Scenario 2 operated in both hemispheres, then northern CCFs likely began retreat from near-rim-full ($d_a \approx 0$), quasi-horizontal surfaces, whereas many southern CCFs may already have been partially inset ($d_a > 0$). In the latter case, reflected shortwave radiation from equator-facing crater walls would preferentially enhance equator-facing ablation (Figure 7.34B), systematically driving poleward-facing slopes, whereas in the north retreat from a near-level surface would allow local microclimatic factors to exert greater control on the eventual slope orientation.

It is important to emphasise that the apparent absence of southern hemisphere examples of rim-overflowing CCFs does not constitute evidence that such cases never existed, but more likely reflects the limited scope of detailed geomorphological investigations in the south. Likewise, most GCM studies have focused on reproducing and explaining the well-documented mid-latitude ice deposits of the northern hemisphere (e.g., Madeleine et al., 2009) and therefore explore parameter spaces that best match northern observations. As a result, the potential for equivalent accumulation and overflow scenarios in the southern hemisphere remains comparatively poorly constrained.

Nonetheless, the combined morphological, modelling, and aspect data presented here suggest that the northern hemisphere has experienced more extensive or more sustained mid-latitude glaciation. This provides a coherent explanation for the hemispheric differences in slope orientation, with northern CCFs shaped by more variable, and possibly more microclimate-sensitive, evolutionary pathways, and southern CCFs more strongly reflecting dominant large-scale climatic forcing.

7.4.4.2. Interglacial Hemispheric Asymmetry

Alternatively, if CCFs in the southern hemisphere experienced accumulation rates comparable to those in the North—whereby sustained accumulation led to the occupation of significantly greater volumes (Scenario 2) or the overtopping of host crater rims (Scenario 3)—then the modern hemispheric disparity in aspect distribution likely reflects interglacial ablation processes rather than initial glacial accumulation patterns. In this scenario, an unidentified mechanism active in the northern hemisphere during interglacial conditions may have reduced (or may still be reducing, should CCFs remain out of

mass balance equilibrium) ablation on equator-facing slopes. This preservation mechanism would effectively explain the broader range of slope aspects currently observed in the north.

Such a process must be restricted to the northern hemisphere, yet its precise nature remains unknown. If the mechanism is linked to insolation, one possibility is that it operates near L_s 90°, during northern winter. At this solar longitude, equator-facing interior crater walls receive significantly reduced insolation over the course of a Martian year, potentially matching the lower insolation levels of other slope aspects.

Ultimately, this specific mechanism, or combination of mechanisms, responsible remains unresolved. Whether these differences reflect past glacial dynamics, present-day climate influences, or a combination thereof, further investigation is required to fully constrain their origin.

7.5. Conclusions

This chapter presents the first global assessment focused exclusively on the surface slope and aspect properties of CCFs, offering new insights into their evolution and the climatic conditions shaping mid-latitude glaciation on Mars. By analysing hemispheric biases in slope magnitude and orientation, it distinguishes between globally consistent processes and those influenced by regional glacial regimes, insolation geometry, and diurnal effects.

CCFs across either hemisphere are characterised by very gentle surface slopes, with typical gradient magnitudes below 0.1. This consistency suggests that slope formation or modification is driven by a common mechanism operating across Mars, likely linked to asymmetries in solar insolation during either mid-latitude glaciation or interglacial periods.

In contrast, CCF surface aspects show distinct hemispheric differences. While both hemispheres share a general poleward and eastward bias, the southern hemisphere shows a tightly clustered pole-facing orientation, whilst the northern hemisphere exhibits a broad aspect distribution. These patterns reflect findings from other VFF populations (e.g. GLFs; Souness et al., 2012) and align with insolation models favouring pole-facing slopes for ice retention (e.g., Kreslavsky et al., 2008).

While the hemispheric consistency in slope magnitude supports a shared formative mechanism, the specific origin of CCF surface slopes remains uncertain. Crucially, it is not yet possible to determine whether these gentle gradients formed during the initial phase of glacial accumulation or during subsequent interglacial modification. To address this uncertainty, three conceptual scenarios are proposed, each reflecting different stages of the CCF life cycle and varying degrees of ice volume change. These include: (1) slope formation during active, crater-confined accumulation; (2) slope development through asymmetric ablation during interglacial retreat without overtopping; and (3) slope development following overtopping and significant ice loss. Each scenario carries distinct implications

for the timing, processes, and environmental conditions involved in CCF evolution, and provides a framework for interpreting how slope may encode information about past glacial activity on Mars.

I draw the following conclusions regarding CCF slope orientation:

1. Poleward bias supports preservation-driven slope development

The consistent pole-facing preference amongst CCFs and other VFFs (Figure 7.26) suggests that insolation-driven preservation governs surface orientation. Pole-facing slopes receive less direct solar input, favouring either reduced ablation or preferential accumulation. This aligns with both previous Martian isolation modelling studies (e.g., Kreslavsky et al. (2006; Williams et al., 2008) and terrestrial analogues (e.g., Evans and Cox, 2005; Evans, 2006; Section 7.11), where north- and south-facing slopes serve as cold traps under asymmetric solar forcing.

Given the uniformly low slope magnitudes across all aspect directions in either hemisphere (Figure 7.15), it is likely that surface tilt in both poleward and non-poleward orientations is still governed by interactions with solar insolation. However, for non-poleward aspects to emerge despite the climatic preference for pole-facing slopes, some factor must locally diminish the strength of this poleward bias. This is most plausibly explained by crater-specific microclimatic conditions, such as rim asymmetry, or background topographic shadowing, or variations in background slope geometry, which reduces insolation on equator-facing interior walls, thereby enabling alternative slope aspects to develop and persist.

2. Widespread east-facing bias points to a diurnal control

A consistent eastward-facing bias is observed in both hemispheres (Figure 7.26), and crucially, shows no spatial clustering for both CCFs and GLFs (Figure 7.35). This suggests that the mechanism responsible for the eastward bias operates globally, rather than being tied to specific regions. I propose that this signal reflects diurnal-scale asymmetries in insolation delivery. Two mechanisms are hypothesised: (i) differential heating between east- and west-facing slopes due to solar geometry (Section 7.4.3.1, Figure 7.36, 7.37) and (ii) variations in morning versus afternoon cloud cover (Section 7.4.3.2, Figure 7.40). Both mechanisms are grounded in terrestrial glaciological studies, where diurnal temperature and/or cloud regimes can drive aspect asymmetries. While these influences are secondary to the dominant poleward gradient, they offer a plausible explanation for the modest but persistent east-facing signal.

As this east-facing bias is not uniform across all craters, its presence likely depends on local microclimatic conditions previously mentioned. In this way the diurnal signal, despite theoretically being present across the entire hemisphere acts as a localised modifier, emerging only where specific conditions allow.

3. The broader aspect spread in the northern hemisphere may reflect greater glacial accumulation

Terrestrial glaciology would predict that the higher-relief terrain of the southern hemisphere should produce a more variable range of slope aspects. However, the opposite is observed. This inversion supports the interpretation that northern hemisphere CCFs once experienced more extensive or prolonged glaciation. Under these conditions, CCFs may have accumulated with little dependence on slope orientation, or to such a degree that it overtopped crater rims. As CCF extent retreated during subsequent interglacial phases, localised microclimatic effects, such as rim asymmetry, background slope, or shading, may have exerted stronger controls on the final orientation of preserved surface slopes, resulting in the broader range of aspects observed today in the north hemisphere.

This interpretation aligns with GCM simulations predicting widespread northern hemisphere accumulation under past orbital configurations (e.g., Madeleine et al., 2009), as well as morphological evidence from Deuteronilus Mensae and Phlegra Montes, where CCFs appear to have previous greater extent and overtopped their host craters (Levy et al., 2010; Dickson et al., 2010). Chapters 5 and 6 further support this view, showing that northern CCFs exhibit lower apparent depths and greater volumes, consistent with more substantial glacial input.

4. Alternatively, interglacial ablation may selectively shape northern aspect distributions

If future studies find that southern hemisphere CCFs also once hosted large ice volumes, then the observed hemispheric difference in aspect spread must result from differential ablation hemispheric histories. In this scenario, a mechanism active only in the northern hemisphere may reduce ablation on equator-facing slopes, allowing a broader distribution of aspects to be preserved. One possible control is reduced insolation at specific orbital intervals (e.g., $L_s 90^\circ$), which disproportionately limits solar input on equator-facing crater walls during northern winter.

Though the precise mechanism driving the hemispheric differences in aspect remains uncertain, the contrast in aspect patterns likely reflects a combination of global slope-forming mechanisms and hemisphere-specific climatic influences, either from enhanced accumulation or modified ablation regimes in the northern hemisphere. The observed uniformity in slope magnitude across CCFs supports interpretations of globally consistent formation processes, while aspect differences record spatial variability in mass balance dynamics.

Together, the slope and aspect properties of CCFs preserve subtle signatures of Mars' mid-latitude glacial history. These findings and their interpretations offer a testable framework for assessing whether similar slope- and aspect-forming mechanisms apply beyond this CCF sample and across other VFF

types globally. This chapter highlights the potential influence of diurnal energy balance on VFF development, offering a plausible explanation for the unresolved east-facing bias observed across multiple VFF populations. They also contribute to our understanding of hemispheric-scale glaciation variability and provide additional support for scenarios involving more extensive northern hemisphere accumulation. As such, these results offer valuable context for future mid-latitude glaciation models and help constrain the climatic and geomorphological conditions that have governed the long-term evolution of CCFs and wider mid-latitude glaciation.

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Chapter 8: Synthesis and Future Work

8.1. Synthesis

Here I integrate the Thesis findings to interpret the history of mid-latitude glaciation recorded by high-confidence CCFs. The distributional, morphometric, and volumetric results and interpretations (Chapters 3–7) are used together to interpret CCF glacial history. In doing so, this chapter consolidates the findings and interpretations of those chapters, interprets the origin of the dominant environmental signals, and sets out an updated evolutionary framework for CCFs.

8.1.1. CCFs record environmental gradients

A fundamental question in Martian glaciology is whether the present-day distribution of VFFs reflects past atmospheric ice deposition primarily or a secondary signal imposed by post-depositional preservation (Souness et al., 2012). CCFs are near-ubiquitous across the mid-latitudes (Levy et al., 2014), and the high-confidence subset isolated in this Thesis (Chapter 3) is similarly widespread (Chapters 3–4), making them the most spatially extensive surviving archive of mid-latitude glaciation on Mars. Crucially, because CCFs are confined within impact craters whose size and shape scale predictably with diameter, their evolution—or at least their post glacial ablation if they once overtopped their rims—takes place within a fixed and well-defined topographic container. This makes CCFs essentially an experimental laboratory for assessing whether observed spatial variability primarily reflects environmental gradients or is instead dominated by geometric accommodation and preservation filtering. Chapter 4 shows that CCF-hosting craters form a statistically distinct subset of the mid-latitude crater population: although CCF abundance and crater abundance are positively correlated, crater availability explains only a moderate fraction of the spatial variability in CCF occurrence. Large departures from proportionality occur in both hemispheres, including a strong north–south asymmetry in which the northern hemisphere hosts more CCFs despite having fewer suitable craters. This demonstrates that today’s pattern of CCF occurrence reflects at least partial environmental control acting on formation and/or long-term survival, as well as topographic opportunity.

When host-crater influence is accounted for—through binning by D —environmental gradients emerge consistently across multiple independent surface-geometry metrics. Within individual D bins, CCF surfaces lie progressively closer to their crater rims toward the poles (decreasing d_a), and the dispersion of d_a narrows poleward in both hemispheres (Chapter 5). Elevation exerts a weaker but statistically detectable influence at the global scale. A similar latitudinal pattern is expressed in CCF surface slopes and aspects. CCF surface slopes decrease poleward in both hemispheres (Chapter 7), as expected with progressively more rim-proximal, lower d_a surfaces at higher latitudes. Aspect distributions show a hemispheric contrast, with persistent poleward bias in the south—regardless of latitude—and a more diffuse, latitude-dependent and hemispheric-dependent structure in the north (Chapter 7). Thus, CCF

surface-geometry metrics exhibit coherent, latitude-dependent organisation that cannot be attributed to or explained by host-crater geometry alone.

In contrast, no comparably strong or consistent latitudinal or elevation dependence is expressed in V_{CCF} (Chapter 6). This decoupling between surface geometry and V_{CCF} establishes a central observational tension: while d_a , slope, and aspect show a clear climatic imprint, V_{CCF} is governed by host-crater geometry and lacks a coherent environmental signal.

8.1.2. Post-glacial ablation as the dominant control on CCF surface geometry

8.1.2.1. *Synthesis: CCFs primarily record post-glacial ablation*

This Thesis finds that, although CCF surface geometry could in principle retain signals of initial accumulation, the dominant environmental gradients recorded by CCF surfaces are more consistently and most simply explained as those associated with post-glacial ablation. The poleward decrease in d_a , and the simultaneous narrowing of its dispersion, align closely with the poleward decrease in mean annual insolation, which persists through current and intermediate obliquity states, only reversing at extreme obliquities ($\sim 50\text{--}55^\circ$; Figure 5.24).

Reduced insolation at higher latitudes suppresses long-term sublimation losses and therefore favours preservation of more rim-proximal surfaces, while enhanced insolation at lower latitudes promotes greater downwasting. A weaker but statistically significant global elevation effect—shown by slightly higher d_a values at higher elevations—is likewise consistent with elevation-dependent sublimation efficiency driven by lower surface pressures and enhanced wind-driven ablation. Together, these relationships indicate that CCF surface geometry is systematically controlled by spatial variations in long-term ablation potential rather than by initial accumulation history. The surface slope results reinforce this interpretation: the most plausible mechanisms for slope development are those involving asymmetric ablation during downwasting (Scenarios 2 and 3 in Chapter 7), and the uniformly low slope magnitudes across all aspects indicate that insolation-mediated energy balance remains the governing control.

If present-day d_a (and thus surface slope and aspect) were primary signals of initial accumulation, CCFs would have had to experience negligible net mass loss during post-glacial retreat, for generating their observed fill levels today. This scenario is incompatible with the volume analysis in Chapter 6: V_{CCF} increases systematically with D at the hemispheric scale, across all tested regions, and within localised clusters, yet shows no coherent relationship with environmental gradients even when binned by D .

The decoupling of volume from surface geometry is most evident at the local scale. If d_a represented a preserved accumulation state, craters of similar size within the same vicinity—sharing the same regional glacial history—would require strong, highly localised controls on accumulation to account for the variation in d_a at similar latitudes observed in Chapter 4. More importantly, the volume results

demonstrate that CCFs are not supply-limited: V_{CCF} increases systematically with crater diameter, and large craters commonly contain ice volumes sufficient to fill neighbouring smaller craters to rim level ($d_a \approx 0$) or to overspill in places $\leq$. The fact that those smaller craters nonetheless retain a range of $d_a > 0$ therefore cannot be explained by limitation in the initial ice supply. Under an accumulation-controlled interpretation, latitude-dependent insolation would establish a minimum bound on attainable d_a , with local/crater-specific factors determining where within that bound individual CCFs accumulated to and stabilised. However, such an interpretation requires additional, unidentified mechanism(s?) to explain the observed hemispheric asymmetry in surface aspect distribution. Differential post-glacial mass loss offers a simpler explanation for these data. It simultaneously accounts for local-scale variability in d_a , latitudinal patterns in surface slope, and the hemispheric divergence in aspect distributions. Crucially, this model explains the lack of environmental gradients in V_{CCF} as a consequence of ablation-driven retreat from a formerly more voluminous state.

This interpretation is further supported by geomorphological evidence for formerly thicker or rim-overflowing states (e.g., Levy et al., 2010; Dickson et al., 2010), which, although regionally restricted, is fully consistent with the results presented here. The simplest synthesis is therefore that CCF d_a , surface slope and aspect predominantly record signals of spatially variable post-glacial ablation, with initial accumulation geometry acting only as a secondary or partially overprinted control.

8.1.2.2. Why do some craters host CCFs and not others?

The interpretation that present-day CCF surface geometry primarily reflects differential post-glacial ablation and preservation, rather than original accumulation patterns, has direct implications for why some craters host well-preserved CCFs, others contain only degraded or ‘low-definition’ fills (Levy et al., 2010), and many host no identifiable ice-related deposits at all. If present-day CCF surface geometry primarily reflects post-glacial ablation rather than accumulation, then the current high-confidence CCF distribution maps not where ice once accumulated, but where crater-confined ice has preferentially survived. As shown in Chapter 6 and synthesised in Section 8.1.2.1, V_{CCF} is only weakly related to latitude or elevation and exhibits large scatter, implying that differential preservation rather than location alone governs which palaeo-CCFs remain recognisable today. The parsimonious inference is therefore that a much larger fraction of mid-latitude craters once hosted palaeo-CCFs, but only a subset of these has persisted to the present in identifiable form.

In this framework, the observed diversity of CCF types—from high-confidence CCFs, through low-definition CCFs, to complete absence of preserved ice—is interpreted to represent a degradation spectrum governed by differential post-glacial ablation. Craters that developed and maintained an effective sublimation lag would preferentially retain thick, high-confidence CCFs. Within this subset, the distinction between “classic” examples (Chapter 3.3.1), which preserve near-pristine surface morphologies, and morphologically variable “non-classic” examples (Chapter 3.3.2), in which partial

degradation has modified surface textures without erasing the underlying traceline structure, can be interpreted as reflecting different degrees of post-glacial modification. Palaeo-CCFs that failed to achieve or maintain mass-balance equilibrium—perhaps due to thin, discontinuous, or repeatedly disrupted overburden—would evolve toward progressively more degraded, low-definition morphologies (Levy et al., 2010), while those lacking sufficient dust or debris supply, or favourable microclimatic conditions, would have ablated entirely. This interpretation is consistent with the findings in Chapter 3, which show that the high-confidence CCF population represents a morphologically filtered subset of a much broader class of crater fills (Levy et al., 2014).

Local and crater-specific factors are therefore likely to exert a critical secondary control on long-term CCF survival. Plausible influences include the efficiency of debris or dust supply to the CCF surface, crater rim geometry and degradation state, and surrounding topographic setting. Some of these influences are hinted at by the slope results in Chapter 7, where the steepest and most anomalous surface tilts occur preferentially in complex topographic settings, but they are not yet systematically quantified. While initial accumulation history may still modulate preservation outcomes in some local settings, I argue that differential preservation under post-glacial ablation is the dominant control on CCFs present state.

8.1.3. Hemispheric differences in CCF glacial history

My results show that the high-confidence CCF population exhibits important hemispheric contrasts. It is expressed consistently across multiple independent metrics. Differences between the northern and southern hemispheres are evident in crater occupation frequency and latitudinal extent (Chapter 4; Levy et al., 2014), in spatial density and mean occupancy (Chapter 4), in apparent depth through a systematic north–south offset (Chapter 5), in volume and host-crater fill fraction (Chapter 6), and in surface aspect distributions (Chapter 7). Comparable hemispheric contrasts are documented in other VFF classes, including differences in GLF aspect (Souness et al., 2012; Brough et al., 2019) and the large-scale volume imbalance reported by Levy et al. (2014), in which cumulative northern LDA and LVF ice volumes ($\sim 2.1 \times 10^5 \text{ km}^3$ and $\sim 5.4 \times 10^4 \text{ km}^3$) greatly exceed their southern counterparts ($\sim 5.5 \times 10^4 \text{ km}^3$ and $\sim 1.2 \times 10^4 \text{ km}^3$). These population-scale differences (including the asymmetry I find for CCFs) reflects hemispheric difference in mid-latitude glaciation generally rather than a peculiarity of individual landform class. The poorer correspondence between southern CCFs and GCM-predicted accumulation zones (Madeleine et al., 2009; Chapter 4) further indicates hemispheric differences in both present-day preservation and the extent of past ice accumulation, although these mismatches may partly reflect limitations in current GCM simulations.

A simple explanation for this asymmetry would be preferential modification by the LDM, particularly in the north. However, LDM alone cannot account for either the magnitude or coherence of the observed differences. Typical LDM thicknesses ($\sim 30 \text{ m}$; Conway and Balme, 2014) are less than the $\sim 40 \text{ m}$ north–

south offset in d_a (Chapter 5), and LDM preferentially affects the outer parts of CCF surfaces rather than their full extents, particularly within the non-classic subset (Chapter 3.3.2). Moreover, the mapping protocol adopted here avoided retracing LDM where underlying CCF morphology remained identifiable (Chapter 5). LDM may therefore locally modify surface slopes and elevations, but it cannot plausibly explain the systematic hemispheric differences in volume, surface elevation, and aspect distributions.

Instead, the coherence of the hemispheric asymmetry across multiple VFF classes (e.g. Souness et al., 2012; Levy et al., 2014; Brough et al., 2019; this Thesis) points to a deeper control linked to long-term asymmetry in the Martian water cycle. Orbital precession, modulated by eccentricity, imposes hemispheric asymmetry on volatile transport by controlling which hemisphere experiences more intense summers and therefore more effective sublimation and redistribution (e.g., Montmessin et al., 2007). Both observations and models show that inter-hemispheric volatile exchange is sensitive to the timing of perihelion and the resulting asymmetry in seasonal insolation, atmospheric humidity, and Hadley circulation strength (e.g., Mischna et al., 2003; Smith, 2004; Bapst et al., 2015). The GCM simulations of Madeleine et al. (2009) exploit this effect by placing perihelion in the hemisphere of interest to reproduce mid-latitude ice distributions, and long-term orbital solutions indicate that the sense of this asymmetry has varied through time (Laskar et al., 2004). Such forcing could therefore have operated both during mid-latitude glaciation and the subsequent retreat, providing a plausible mechanism for generating systematic hemispheric asymmetry in post-glacial modification even if initial accumulation were broadly similar. In this framework, the northern mid-latitudes may have experienced either more extensive or more recent accumulation, while the southern mid-latitudes may have undergone more effective or more persistent net ice loss, consistent with their lower volumes, smaller d_a , and stronger pole-facing aspect bias.

At the same time, terrestrial glaciology cautions against expecting simple or globally synchronous hemispheric behaviour. Glacier response to climate forcing varies strongly with local setting, topography, and moisture supply, and inter-hemispheric or even inter-regional synchrony is often weak or absent (e.g., Winkler and Matthews, 2010). The Martian hemispheric asymmetry documented here therefore need not imply fundamentally different climate states, but may instead reflect differences in how similar orbital forcing was expressed through accumulation, flow, and ablation under different boundary conditions.

Despite these differences, the CCF population also exhibits robust hemispheric similarities. In both hemispheres, CCF occurrence is biased toward small and intermediate crater diameters (Chapter 4), d_a increases with D , decreases with latitude and increases with elevation (Chapter 5), volume increases with D (Chapter 6), and surface slopes are uniformly shallow regardless of aspect (Chapter 7). Both hemispheres also show an overall tendency toward pole-facing slopes, albeit much more strongly in the

south (Chapter 7). These shared properties indicate that CCFs in both hemispheres formed through broadly similar mechanisms under comparable climatic regimes, with hemispheric asymmetry arising primarily from differences in the magnitude and/or timing of post-glacial modification rather than from fundamentally different modes of formation.

8.1.4. Implications for CCF glacial evolution and history

8.1.4.1. An updated evolutionary model for CCFs

This section places the interpretations developed above within the broader evolutionary framework for CCFs outlined in Section 2.4.3 (Figure 2.25). Figure 8.1 presents an updated version of that schematic, retaining the three core phases of glacial accumulation, post-glacial retreat, and long-term modification. I have annotated this schematic to indicate (i) the stages at which the morphometric signals documented in Chapters 5–7 emerge, and (ii) how hemispheric asymmetry is introduced and amplified throughout the CCF lifespan.

Peak accumulation states and the origin of hemispheric differences (Fig. 8.1 C–E)

Panels C–E represent palaeo-CCF states in which ice volumes and surface elevations exceeded those observed today. This is a necessary precondition for the interpretation developed in Sections 8.1.2–8.1.3; if present-day surface geometry primarily records post-glacial ablation, then earlier CCFs must have been thicker and more extensive. Hemispheric asymmetry could have already been expressed at this stage, either through differences in accumulation rate, accumulation duration, or the timing of the most recent major accumulation phase.

The similarity in slope magnitudes between hemispheres (Chapter 7) suggests that the same physical processes governed surface evolution in both hemispheres. But the broader and more diffuse aspect distributions in the north may reflect higher or more rim-proximal palaeo-surfaces prior to retreat, allowing non-poleward slopes to develop from initially flatter surfaces during downwasting. In contrast, southern CCFs, which appear to have already been systematically lower relative to their rims, would have exposed a greater fraction of crater walls during retreat, enhancing wall-controlled, pole-facing ablation effects (Section 7.4.2).

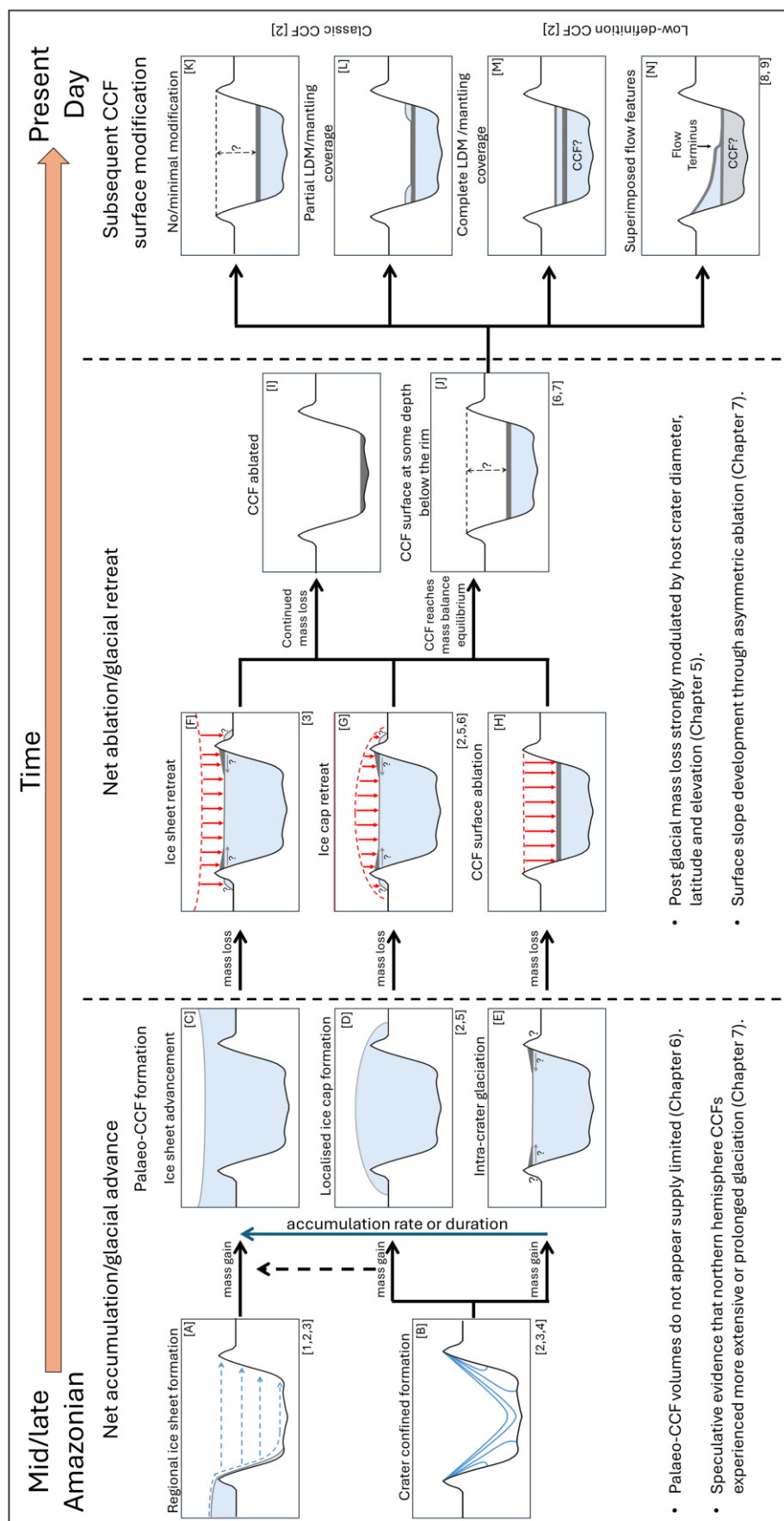


Figure 8.1. Updated schematic of CCF evolution. Conceptual diagram summarising the sequence of processes proposed in the literature for the evolution of concentric crater fills, from initial ice accumulation, through phases of growth and subsequent ablation, to progressive surface modification and degradation leading to present-day classic and low-definition morphologies. The schematic is annotated to indicate (i) the stages at which the morphometric signals documented in Chapters 5–7 are expected to emerge, and (ii) how hemispheric asymmetry is introduced and progressively amplified through the CCF lifespan. Numbered stages correspond to published conceptual and modelling frameworks: [1] Head et al. (2010); [2] Levy et al. (2010); [3] Fastook and Head (2014); [4] Weitz et al. (2018); [5] Dickinson et al. (2010); [6] Levy et al. (2014); [7] Pedersen and Head (2010); [8] Pedersen et al. (2012); [9] Dickinson et al. (2017).

Post-glacial retreat and establishment of modern surface geometry (Fig. 8.1F–H)

Panels F–H illustrate the retreat and downwasting phase during which CCFs transition to their modern, crater-confined configurations. This is the phase when the dominant morphometric signals identified in this Thesis are generated. Chapter 5 shows that surface depth below the rim varies systematically with diameter, latitude, and elevation, with the poleward decrease best explained by latitude-dependent insolation forcing (Panel J). Chapter 7 shows that surface slopes and aspects are likewise consistent with insolation-controlled, asymmetric ablation within the host crater.

During this phase, CCF surfaces either retreat until a new mass-balance configuration is achieved beneath a protective overburden or sublimation lag or continue to ablate at extremely low rates. The weak but detectable elevation dependence of d_a may reflect either hemispheric-scale differences in accumulation history or enhanced sublimation efficiency at higher elevations due to lower atmospheric pressure. In either case, the synthesis indicates that modern global pattern of CCF surface geometries is fundamentally ablation-dominated rather than recording the construction phase (earlier accumulation).

Long-term preservation and the degradation spectrum (Fig. 8.1K–L)

Panels K and L illustrate the final divergence into well-preserved, high-confidence CCFs and degraded or low-definition examples (Panels I, L and M). As argued in Section 8.1.2.2, this spectrum is most parsimoniously interpreted as the outcome of differential post-glacial preservation. CCFs that developed and maintained an effective insulating cover stabilised and remain recognisable today, whereas others underwent continued degradation or complete loss.

8.2.1.2. CCF accumulation timescales and the duration of glacial emplacement

A key question of the history of CCFs, and VFFs more generally, is whether they formed during a single episode of mid-latitude glaciation or experienced cumulative growth across multiple orbital cycles (Section 2.3.2). While smaller glacier-like forms (GLFs), with mean thicknesses of ~ 130 m, could plausibly accumulate within ~ 13 ka under net accumulation rates of ~ 10 mm yr⁻¹ (Brough et al., 2019)—well within the ~ 20 – 40 ka duration of a high-obliquity interval (Laskar et al., 2004)—the substantially larger volumes of most CCFs imply fundamentally longer and more complex growth histories. Numerical modelling by Weitz et al. (2018) indicates that craters up to ~ 40 km in diameter require ~ 5 Myr to reach rim-full conditions under intermediate-obliquity forcing, implying that CCF growth—with assumed plausible accumulation rates—must span multiple orbital excursions.

Simple scaling calculations provide a complementary and independent test of this conclusion. Using the calculated V_{CCF} and d_c derived in Chapter 6, together with hemispheric crater depth–diameter scaling relationships (Chapter 5), first-order estimates of the time required to reach both present-day surface elevations and rim-full states can be made for various (assumed) net accumulation rates. This approach is not intended to replace numerical ice-flow modelling by Weitz et al. (2018), but rather to test whether

single-cycle emplacement is even physically plausible. If we adopt a net accumulation rate of 10 mm yr^{-1} —the mean value predicted by GCMs for intermediate obliquity (Madeleine et al., 2009), then the present-day CCF surface depths would require $\sim 150 \text{ ka}$ of accumulation in the north and $\sim 124 \text{ ka}$ in the south, while achieving rim-full states would require $\sim 187 \text{ ka}$ and $\sim 167 \text{ ka}$, respectively (Figure 8.2). These durations exceed the length of a single 20–40 ka high-obliquity interval. Using a more conservative accumulation rate of 1 mm yr^{-1} , consistent with the lower bounds explored by Weitz et al. (2018), lengthens these timescales by an order of magnitude, yielding Myr-scale growth durations entirely consistent with their numerical results.

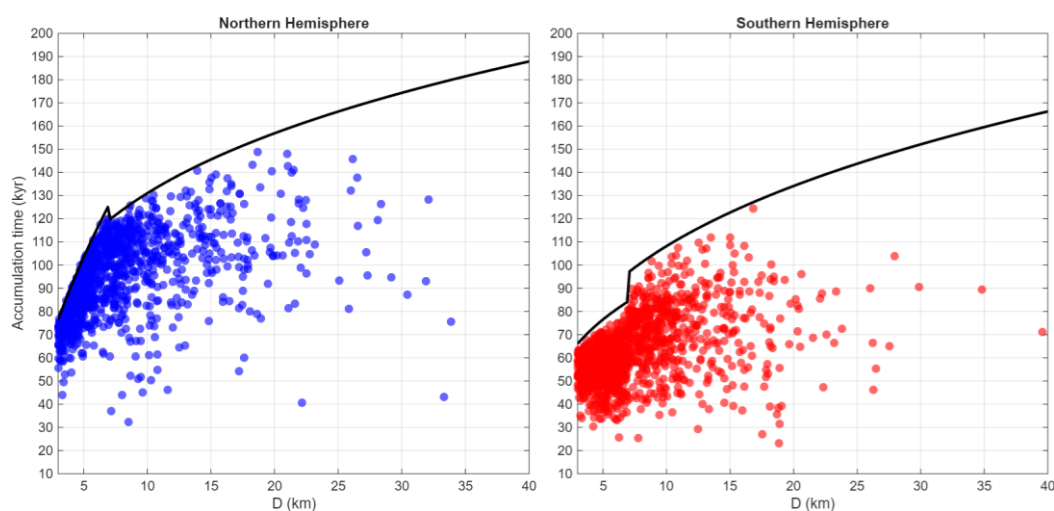


Figure 8.2. Estimated accumulation times for CCFs. Panels show accumulation time (ka) as a function of host crater diameter (km) for (A) northern and (B) southern CCFs. Times correspond to the duration required for each CCF to achieve its present thickness assuming a constant net accumulation rate of 10 mm yr^{-1} consistent with peak GCM-derived values for mid-latitude glaciation (e.g., Madeleine et al., 2009). The black line denotes the time required for CCFs to reach the rim-full threshold ($d_a = 0$) under the same assumed rate, assuming the theoretical d/D scaling.

Importantly, GCM-derived accumulation rates in these scenarios assume an equatorial source for mid-latitude ice. Even under highly optimistic conditions in which precipitation rates are increased by an order of magnitude, Parsons and Holt (2016) show that formation times for large VFFs remain on the order of $\sim 20\text{--}50 \text{ ka}$. While such rates might permit rapid growth of small systems, they remain difficult to reconcile with the volumes required to fill most CCF-hosting craters unless unrealistically efficient and sustained accumulation is assumed. As they note, kyr-scale growth times for large VFFs are only achievable if accumulation rates reach $\sim 30\text{--}70 \text{ mm yr}^{-1}$ (Forget et al., 2006), values that are not currently supported by GCMs.

Consequently, if mid-latitude ice supply is realistically limited to $\lesssim 10\text{--}16 \text{ mm yr}^{-1}$, the observed volumes of CCFs cannot be explained by a single glacial episode. Instead, CCFs are most plausibly interpreted as the cumulative products of repeated accumulation phases operating across multiple

intermediate-to-high obliquity cycles, likely punctuated by intervals of stagnation or net ablation. CCFs therefore most probably record integrated glacial histories spanning multiple orbital excursions. This conclusion does not exclude the possibility that smaller or atypical CCFs formed rapidly during particularly favourable intervals, but at the population scale, both numerical modelling and first-order scaling arguments indicate that multi-cycle growth is the norm unless accumulation rates were dramatically higher than currently thought.

8.2. Future work

This thesis has established a quantitative and morphometric framework for interpreting the distribution, geometry, and preservation state of concentric crater fills, but it also exposes a set of fundamental unresolved questions concerning their initial volumes, emplacement pathways, and long-term evolution. In this final section, I therefore shift from synthesis to prospectus and briefly outline several priority directions for future research that I believe are required to test, refine, and extend the interpretations developed here. These avenues span improved constraints on palaeo-ice thickness and timing, tighter coupling between observations and climate–ice modelling, and targeted investigation of key transitional or ambiguous landforms. The purpose is not to be exhaustive, but to identify the most consequential gaps that now limit further progress and to indicate how they might be addressed.

8.2.1. Evolutionary pathways and initial ice volumes of paleo-CCFs

The glacial evolution of CCFs to their current extent remains unresolved. Uncertainties remain on whether their present surface elevations record maximum accumulation or represent the ablated remnants of formerly thicker, potentially rim-full ice bodies. Throughout the empirical chapters (Chapters 5–7), global variations in CCF d_a , V_{CCF} , and surface aspects have been evaluated against both accumulation-dominated and ablation-dominated scenarios. As discussed above, while an accumulation-controlled interpretation would require present CCF surfaces to approximate long-term mass-balance equilibrium, the systematic hemispheric and elevation-linked trends identified here are interpreted to indicate post-glacial ablation. The magnitude of this mass loss—and thus the former maximum volume of paleo-CCFs—remains unknown.

Direct geomorphic evidence for more extensive CCF configurations in the past is currently limited. Outward-facing lobate ridges interpreted as rim-overtopping moraines have been reported at Deuteronilus Mensae (Dickson et al., 2010), and interior trimlines indicating higher former paleo-CCF surface stands have been described in Phlegra Montes (Levy et al., 2010). The restriction of such evidence to only two sites, both in the northern hemisphere, leaves it unclear whether rim-overtopping was a rare, localised phenomenon or a widespread phase of CCF evolution that is now poorly preserved. Independent motivation to revisit this question comes from both numerical modelling and the empirical results of this thesis. Ice-flow simulations indicate that host craters 80 km in diameter can be filled to the rim within ~ 8 Myr at accumulation rates of only ~ 1 mm yr⁻¹ (Weitz et al., 2018), which are lower

than those predicted for favourable orbital configurations by GCMs (e.g., Madeleine et al., 2009). Furthermore, Chapter 6 demonstrates that V_{CCF} increases with D across both hemispheres; notably, the V_{CCF} contained within larger host craters frequently exceeds the total V_{host} capacity of smaller host craters. Together, these observations make it plausible that some CCFs reached rim-full or overtopping states. While it is tempting to adopt a rim-full initial state as a baseline to interpret the volume equivalent of present d_a as a measure of post-glacial mass loss—yielding an estimated loss of $5.3 \times 10^4 \text{ km}^3$, or a GEL of 0.36 m—this assumption is arbitrary and difficult to test or support. Paleo-volumes may well have once exceeded crater capacity.

It may be possible to tackle the issue by undertaking systematic assessment of the high-confidence CCF population for geomorphic indicators of greater former thickness and extent, including exterior moraines and interior trimlines. While the volume–diameter relationship constrains the ice volume required to reach the rim, only the identification of overtopping candidates can determine whether paleo-volumes commonly exceeded crater capacity. Conversely, a confirmed absence of such features at the population scale would place equally strong limits on the maximum thickness of CCF ice. Establishing whether rim-overtopping was rare or widespread is therefore essential for discriminating between accumulation-limited and ablation-dominated evolutionary pathways. Even a modest increase in the number of confirmed overtopping or high-stand cases would support the interpretation that some CCFs formed locally extensive ice masses, potentially analogous to small ice caps. More generally, quantifying the spatial distribution and magnitude of former high-stand indicators—if found—would permit first-order reconstruction of paleo-CCF volumes and more robust estimates of the total post glacial mass loss. Whether present CCF surfaces record equilibrium accumulation levels or represent the skeletal remnants of once-thicker ice bodies remains the central unresolved question in reconstructing the Amazonian mid-latitude ice inventory.

8.2.2. CCF morphological subtypes

The morphological diversity of CCFs represents a primary, yet largely untapped, archive for reconstructing the glacial history of the Martian mid-latitudes. High-confidence CCFs exhibit a broad spectrum of surface morphologies (Chapter 3), including both ‘classic’ ($n = 390$) and ‘non-classic’ ($n = 2,668$) forms, which may reflect differences in accumulation pathways, internal flow behaviour, ice deposition and/or the degree of post-glacial modification. While Chapter 3 established the first global framework for subdividing CCF morphologies, this scheme remains primarily descriptive. A refined classification grounded in surface process interpretation would permit direct testing of whether different morphological subtypes systematically differ in d_a , surface slope and aspect, geographic distribution, or host-crater D range, and whether such differences account for a portion of the ~50% of variance in d_a that remains unexplained by the MLR (Chapter 5). If specific morphotypes cluster in particular hemispheres, elevation ranges, or D bins, this may help constrain CCF formation pathways

and evaluate the role of post-glacial modification. Falling outside the scope of the present Thesis, achieving this will require detailed surface mapping supported by high-resolution images, DTMs, and systematic subdivision by D and geomorphic context.

8.2.3. Crater confined VFF within LVF and LDA systems

In certain geomorphic settings, impact craters form integral components of regional LDA and LVF systems, where breached rims facilitate the inflow or outflow of debris-covered ice that either terminates within or flows through the crater interior (e.g., Levy et al., 2014; Wueller et al., 2025; Figure 2.17F, 4.13). These configurations represent observable analogues for the ice-sheet inflow model CCF formation pathway (e.g., Head et al., 2010; Levy et al., 2010; Fastook and Head, 2014; Figure 8.1A). Initial classification work suggests that such crater-filling units typically exhibit linear ridge-and-trough surface textures, with ridges oriented convex downslope, consistent with LDA-style flow morphology (Wueller et al., 2025) rather than concentric traceline patterns. However, it remains unclear whether these features represent a fundamentally distinct population of crater-filling ice bodies or whether they form part of a broader continuum that includes some CCF subtypes. Comparison of their morphology, surface textures, and host-crater characteristics with those of CCF subtypes (Section 8.2.2) would test whether some CCFs may represent relict or modified remnants of regional ice-sheet incursions, or whether crater-filling ice associated with LDA/LVF systems constitutes a separate genetic class whose present lack of concentric fabric reflects a different flow regime rather than a different preservation history.

8.2.4. CCF–gully associations as indicators of CCF post-glacial evolution

This Thesis has demonstrated that CCF surface elevation relative to the host-crater rim varies systematically with crater size, latitude, and elevation. A substantial fraction of the CCF population is not rim-full; southern hemisphere CCFs show an average increase in d_a of ~ 40 m, with $\sim 50\%$ of the observed variance in d_a explained by these environmental and geometric controls (Chapter 5). This implies that many CCF-hosting craters—particularly in the south—preserve large areas of exposed crater wall despite being morphologically classified as crater-filling.

Gullies are interpreted as transitional landforms linked to the re-equilibration of surface and near-surface ice under recent climatic variability (e.g., Head et al., 2003; Conway et al., 2019). Given their preference for steep slopes—with $\sim 63\%$ of all Martian gullies occurring within impact craters (Noblet et al., 2024)—they might be expected to occur preferentially in under-filled CCF-hosting craters. However, gullies occur in only $\sim 12\%$ of craters hosting ice-rich VFFs (Conway et al., 2019), and more generally a strong anti-correlation is observed between gullies and VFFs across the mid-latitudes (e.g., Conway et al., 2017; Noblet et al., 2024). This anti-correlation is commonly attributed to the thermal and mechanical stability of intact subsurface ice, which suppresses incision, and the tendency for debris-mantled surfaces to bury or overprint earlier gully systems (Conway et al., 2017). For crater-confined

VFFs specifically, this suppression should be effective as the ice reduces exposed wall area and limits potential catchment zones (Christensen, 2003; Conway et al., 2018). The systematic departure of many CCFs from rim-full conditions therefore creates a clear, testable tension between expected gully suppression and the availability of exposed crater-wall slopes.

Preliminary analysis using the high-confidence CCF subset and the global gully inventory of Noblet et al. (2024) reveals a pronounced hemispheric asymmetry: only ~4% ($n=64$) of northern CCF-hosting craters contain gullies, compared to ~25% ($n=360$) in the south. This disparity implies more extensive post-glacial modification of southern CCFs and mirrors the hemispheric contrasts documented throughout this thesis, including higher d_a values (Chapter 5), lower volumes (Chapter 6), and a stronger poleward-facing slope bias (Chapter 7). The coherence of these independent metrics indicates more effective or prolonged post glacial ablation in the southern hemisphere. While this thesis has explored both accumulation and ablation drivers, the spatial relationship between gullies and CCFs has not been examined systematically. Resolving this relationship offers a direct geomorphic constraint on whether present-day CCF surfaces record maximum accumulation states or represent the ablated remnants of once-thicker ice bodies.

These observations suggest several avenues for targeted investigation. First, does a threshold in d_a govern gully initiation? Because northern hemisphere CCFs are more rim-full (lower d_a ; Chapter 5), such a threshold could explain the pronounced hemispheric asymmetry in CCF–gully co-occurrence. Second, the pole-facing bias of southern CCF surfaces (Chapter 7) implies greater exposure of equator-facing crater walls. If gully activity is linked to ice degradation, incision should preferentially develop on these high-insolation slopes. A systematic test of gully occurrence against both d_a and wall-slope aspect would confirm whether incision is directly coupled to CCF surface drawdown.

A further unresolved issue is why specific CCF-hosting craters develop gullies while others with near identical geometries do not. Local-scale contrasts (Figure 8.3A) illustrate this clearly: two adjacent craters host high-confidence CCFs, yet only the classic feature possesses extensive gully systems (Figure 8.3B & C), while its non-classic neighbour exhibits disorganised tracelines and no gully presence. Such disparities suggest that CCF–gully interactions are not uniform and may reflect crater scale localised differences in subsurface ice content, emplacement history, post glacial modification history or complex feedback between surface deformation and incision. These contrasts motivate a broader assessment of gully co-occurrence across the full CCF inventory (e.g., Levy et al., 2014). Specifically, research should compare gully incidence across three distinct morphological subsets: high-confidence classic, high-confidence non-classic, and low-definition CCFs. If low-definition or non-classic features represent more advanced degradation states, they might be expected to show higher gully frequencies due to increased wall exposure. Conversely, if gully formation requires a specific

threshold of ice purity or volatile availability, they may preferentially associate with high-confidence CCFs.

Resolving these relationships would provide a direct means of evaluating whether gullies serve as a proxy for active CCF degradation. Determining whether gullies track the "drawdown" of the ice surface will clarify whether current CCF morphologies record the maximum extent of Amazonian accumulation or represent the ablated, skeletal remnants of once-thicker ice bodies.

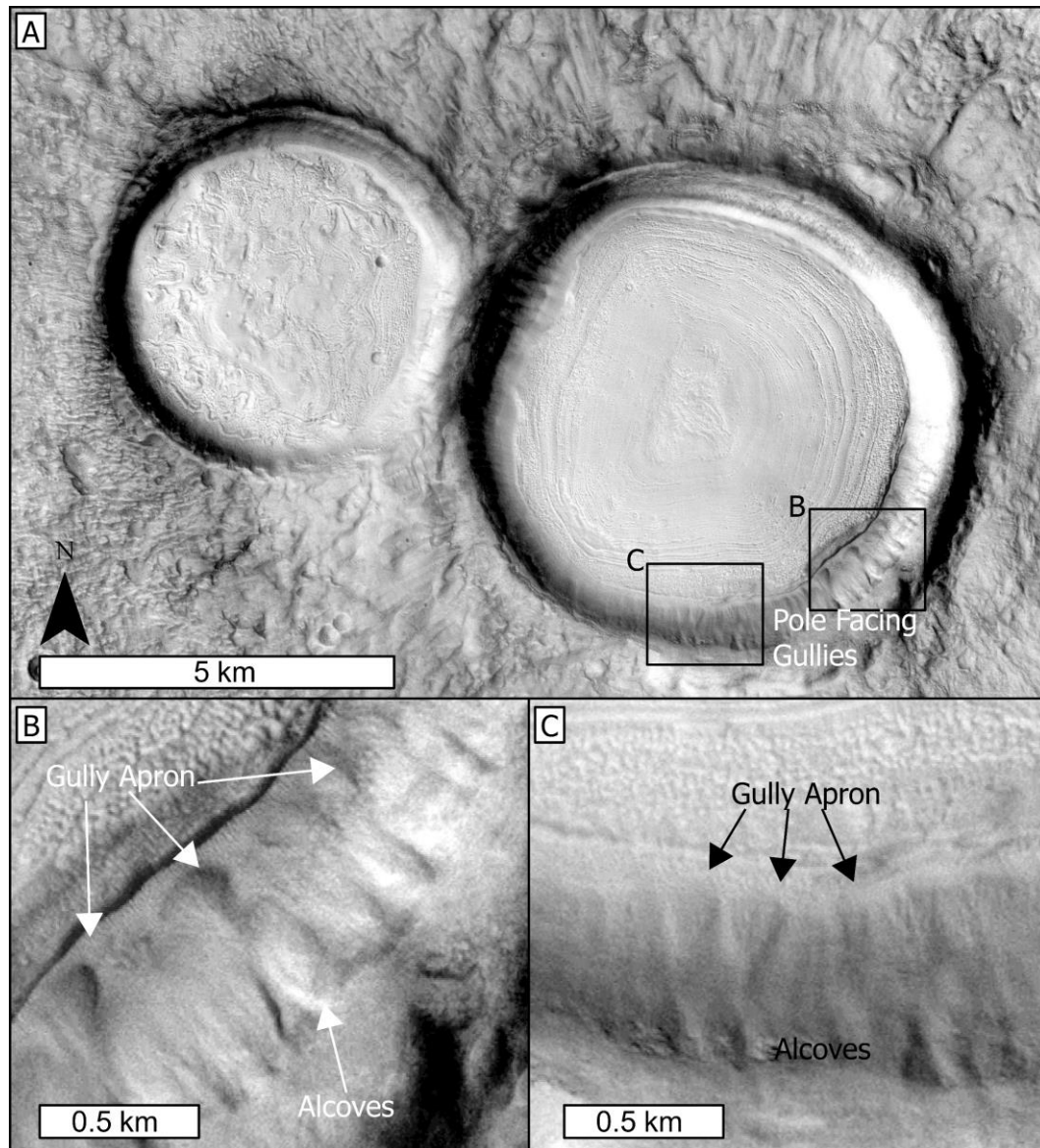


Figure 8.3. Local-scale contrast in gully development on adjacent CCF-hosting craters. (A) CTX image P15_007099_2242 showing two neighbouring craters at 43.88° N, 51.03° W, both hosting high-confidence CCFs. The right-hand crater hosts a classic CCF and displays well-developed, pole-facing gully systems, whereas the left-hand crater hosts a more degraded, non-classic CCF and shows no gully development. (B–C) Close-ups of the gully systems on the classic CCF.

8.3 References

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Appendix A

Solving for d_c :

$$V = \int_0^{d_c} \pi \left(\frac{z}{k} \right)^{\left(\frac{2}{n} \right)} dy$$

$$V = \int_0^{d_c} \pi \left(\frac{z^{\left(\frac{2}{n} \right)}}{k^{\left(\frac{2}{n} \right)}} \right) dy$$

$$V = \frac{\pi}{k^{\left(\frac{2}{n} \right)}} \int_0^{d_c} \left(z^{\left(\frac{2}{n} \right)} \right) dy$$

Using the power law for integration $\int y^a dy = \frac{y^{a+1}}{a+1}$ and where $a = \frac{2}{n}$

$$V = \frac{\pi}{k^{\left(\frac{2}{n} \right)}} \cdot \left[\frac{y^{\frac{2}{n}+1}}{\frac{2}{n}+1} \right]_0^{d_c}$$

Simplifying the exponent and denominator $\frac{2}{n} + 1 = \frac{2}{n} + \frac{n}{n} = \frac{2+n}{n}$ gives:

$$V = \frac{\pi}{k^{\left(\frac{2}{n} \right)}} \cdot \left[\frac{y^{\frac{2+n}{n}}}{\frac{2+n}{n}} \right]_0^{d_c}$$

Plugging in the upper bounds (d_c) and lower bounds (0):

$$V = \frac{\pi}{k^{\left(\frac{2}{n} \right)}} \cdot \left(\frac{(d_c)^{\frac{2+n}{n}}}{\frac{2+n}{n}} - \frac{(0)^{\frac{2+n}{n}}}{\frac{2+n}{n}} \right)$$

$$V = \frac{\pi}{k^{\left(\frac{2}{n} \right)}} \cdot \left(\frac{(d_c)^{\frac{2+n}{n}}}{\frac{2+n}{n}} \right)$$

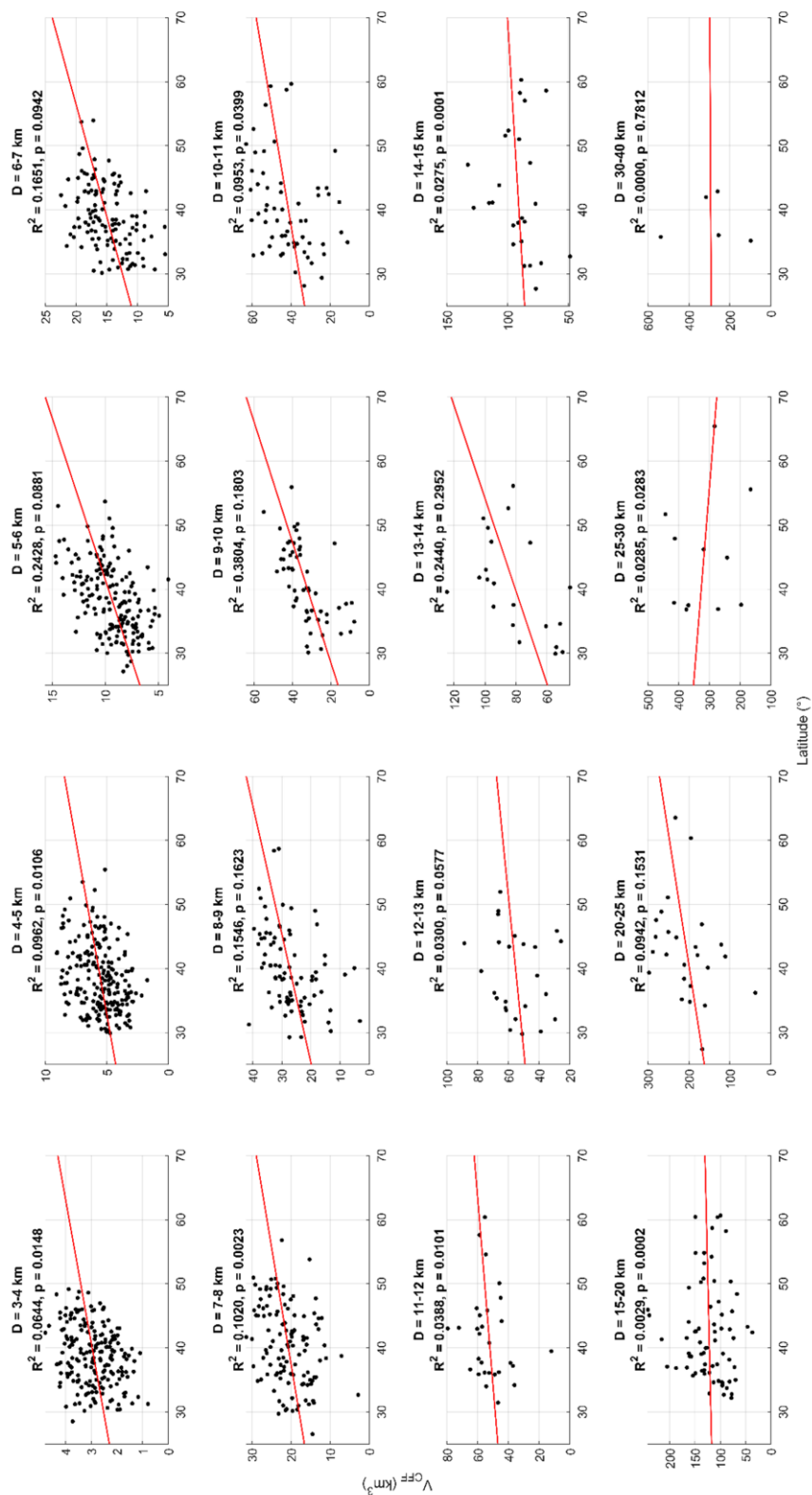
$$V = \frac{\pi n}{k^{\left(\frac{2}{n} \right)} \cdot (2+n)} \cdot (d_c)^{\frac{2+n}{n}}$$

$$(d_c)^{\frac{2+n}{n}} = \frac{V \cdot k^{\left(\frac{2}{n} \right)} \cdot (2+n)}{\pi n}$$

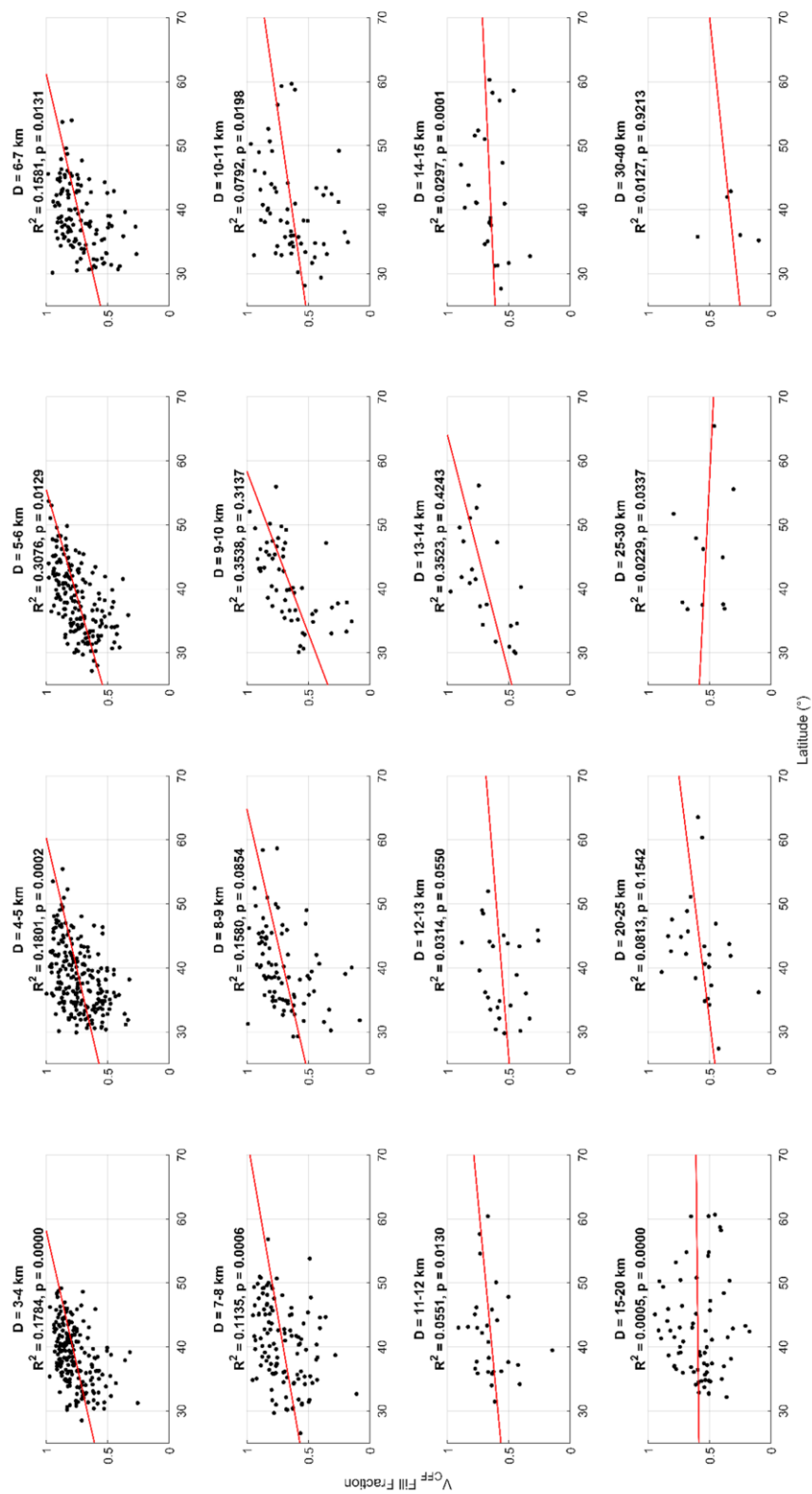
Raising both sides of the equation to its reciprocal power $\frac{n}{2+n}$ gives:

$$\left((d_c)^{\frac{2+n}{n}}\right)^{\frac{n}{2+n}} = \left(\frac{V \cdot k\left(\frac{2}{n}\right) \cdot (2+n)}{\pi n}\right)^{\frac{n}{2+n}}$$

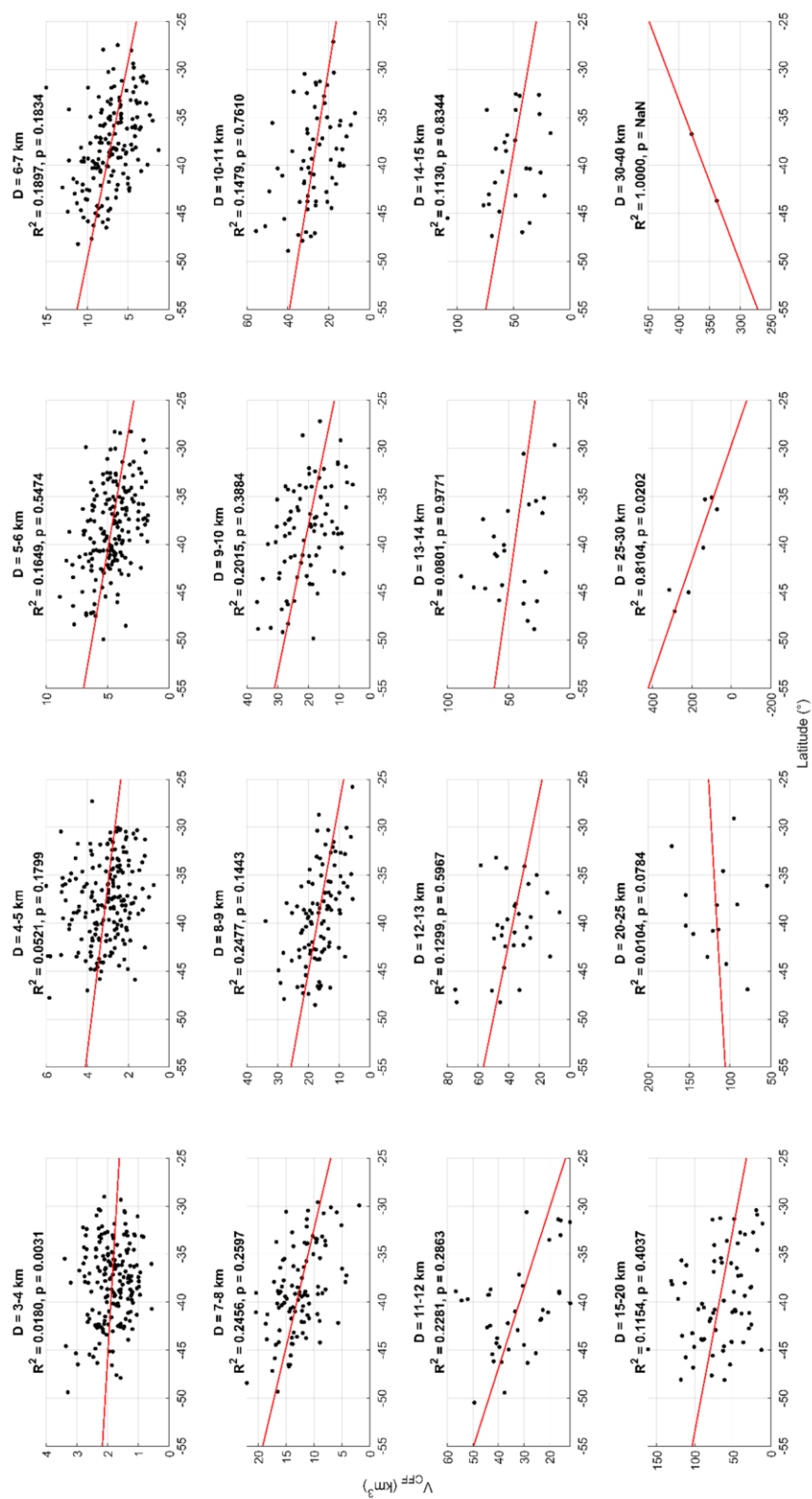
$$d_c = \left(\frac{V \cdot k\left(\frac{2}{n}\right) \cdot (2+n)}{\pi n}\right)^{\frac{n}{2+n}}$$



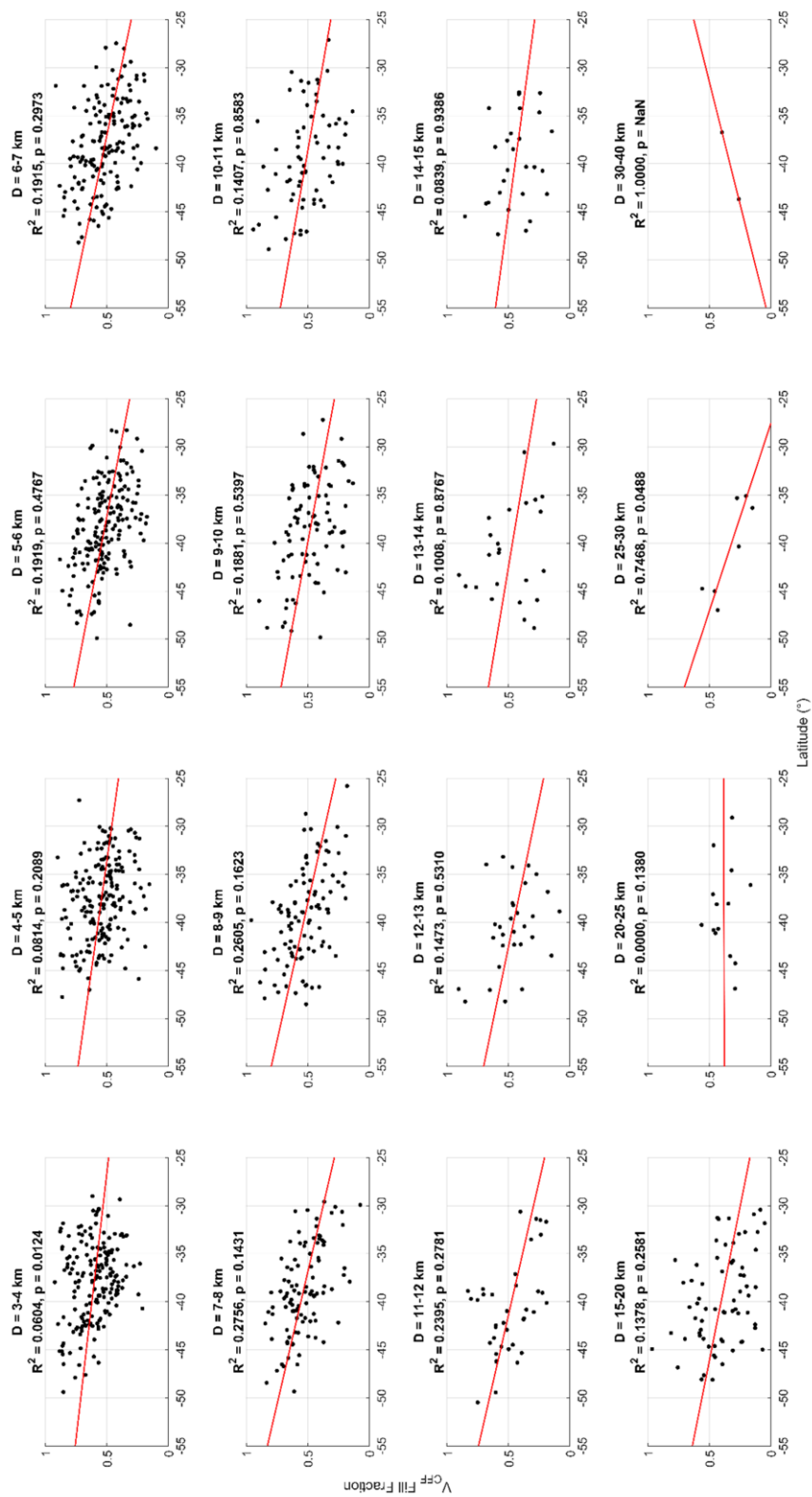
A.1. Scatter plot of V_{CCF} against latitude for CCFs in the northern hemisphere, in different crater-diameter bins. Red lines denote LR, with corresponding R^2 and p -values, are labelled for each plot.



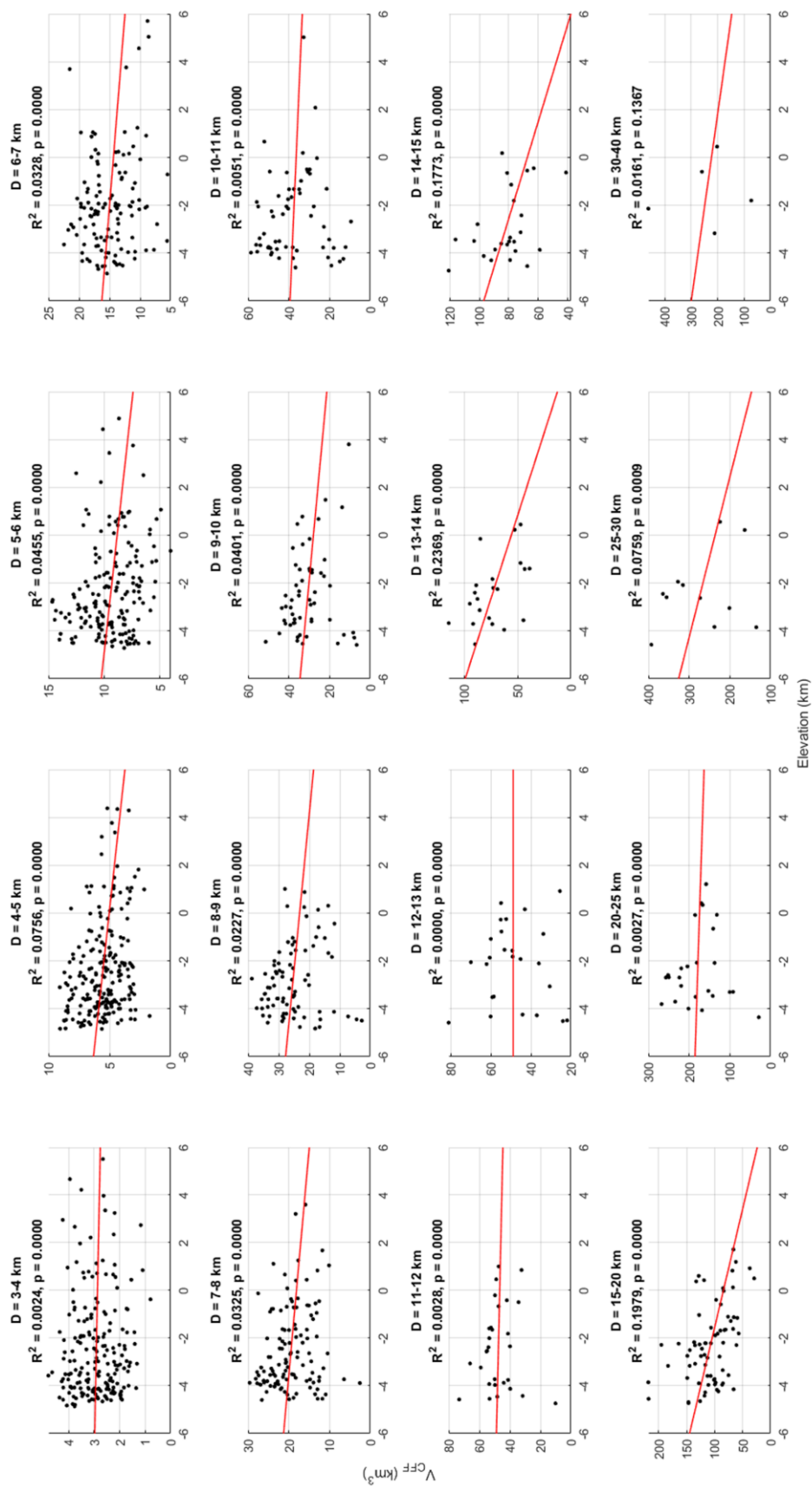
A.2. Scatter plot of fill fraction against latitude for CCFs in the northern hemisphere, in different crater-diameter bins. Red lines denote LR, with corresponding R^2 and p -values, are labelled for each plot.



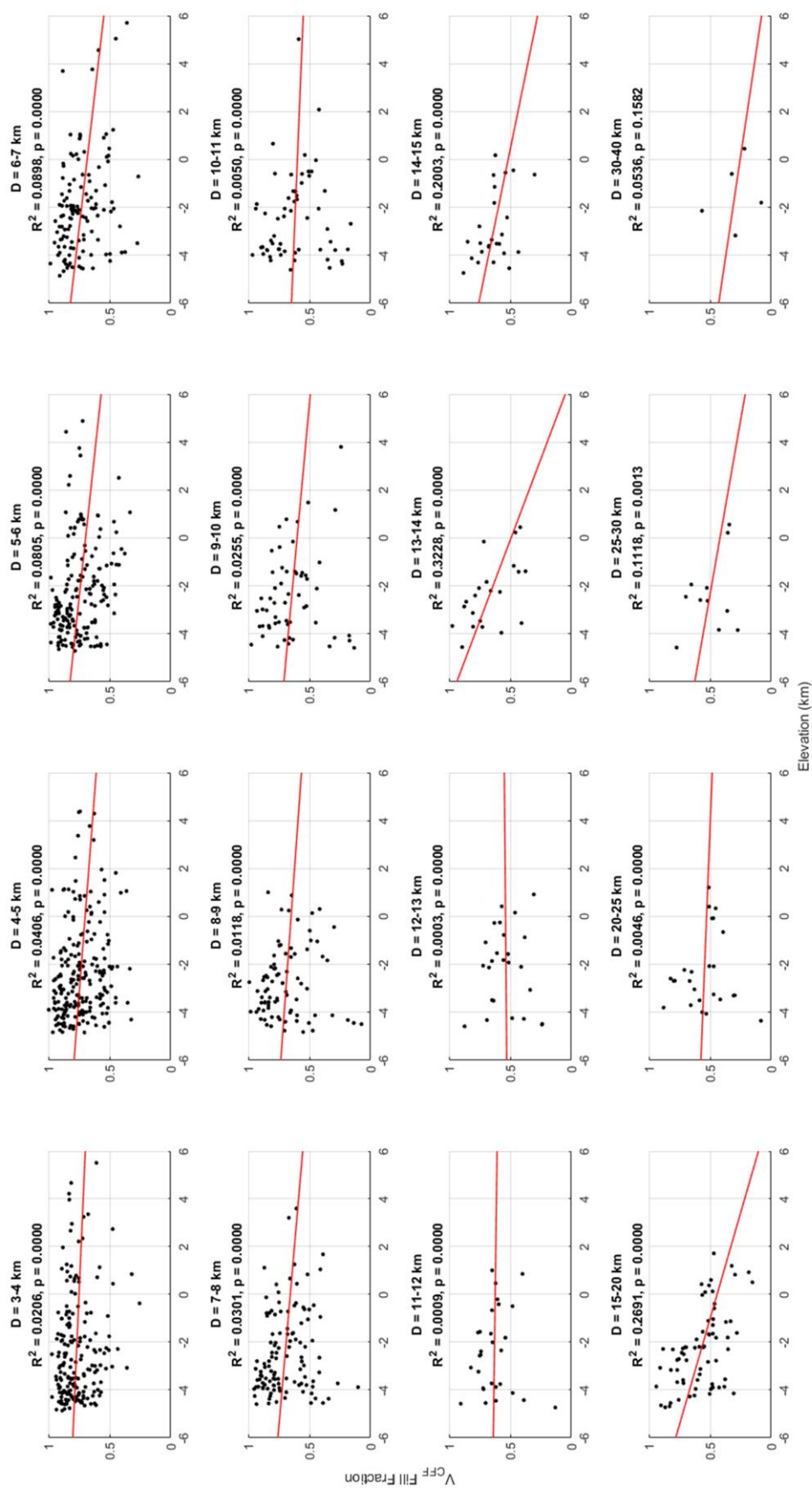
A.3. Scatter plot of V_{CCF} against latitude for CCFs in the southern hemisphere, in different crater-diameter bins. Red lines denote LR, with corresponding R^2 and p -values, are labelled for each plot.



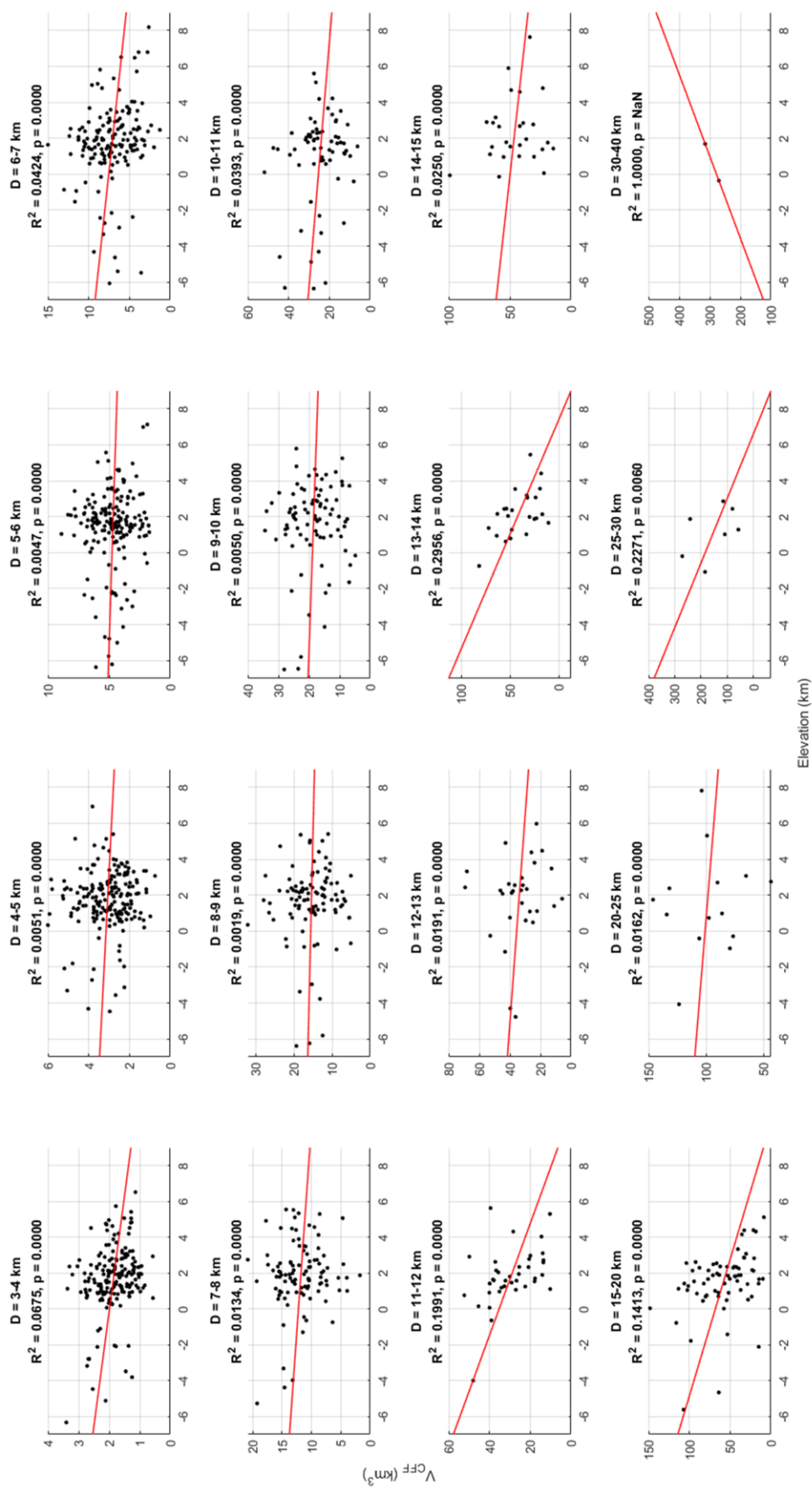
A.4. Scatter plot of fill fraction against latitude for CCFs in the southern hemisphere, in different crater-diameter bins. Red lines denote LR, with corresponding R^2 and p -values, are labelled for each plot.



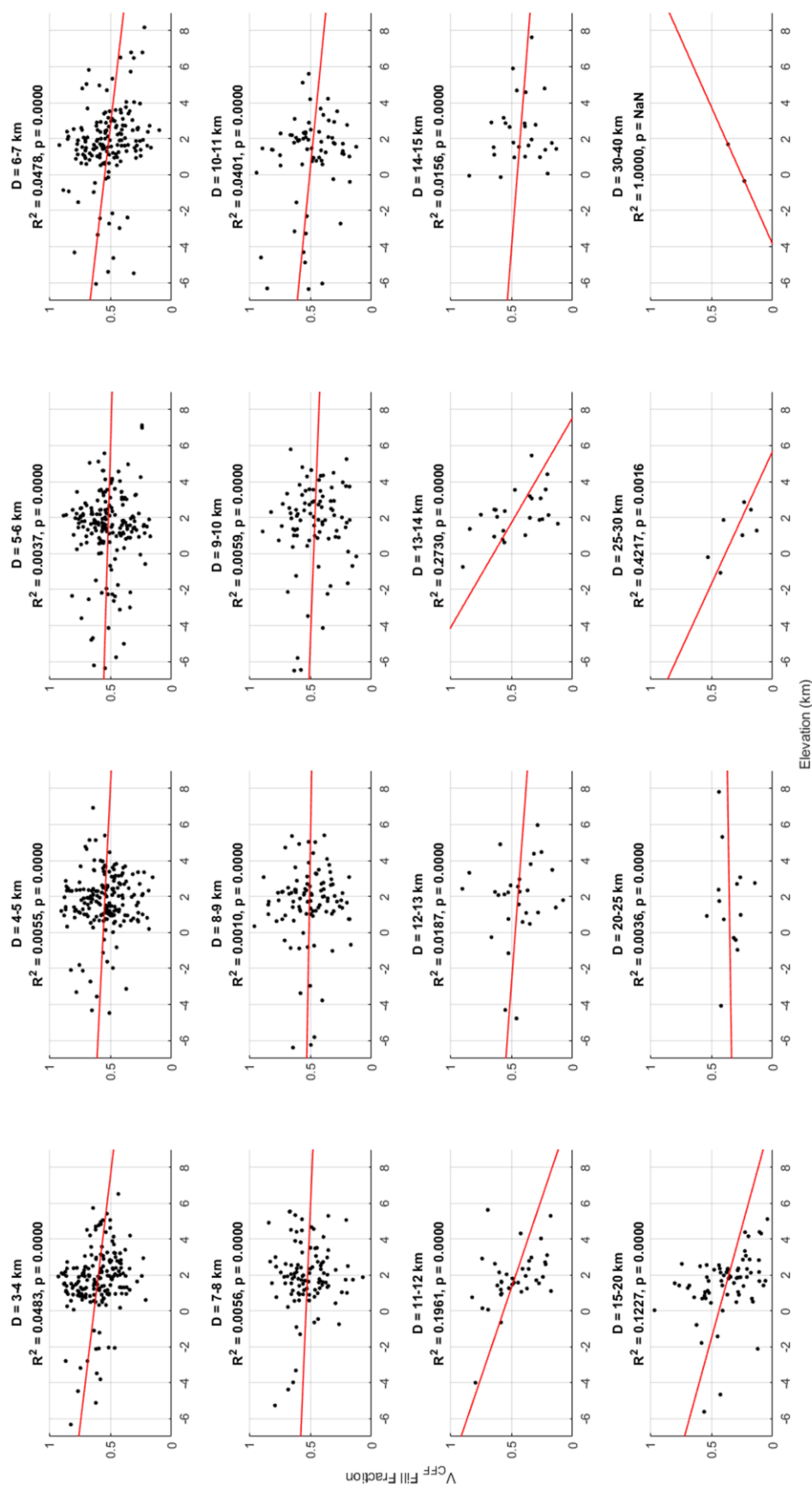
A.5. Scatter plot of V_{CCF} against elevation for CCFs in the northern hemisphere, in different crater-diameter bins. Red lines denote LR, with corresponding R^2 and p -values, are labelled for each plot.



A.6. Scatter plot of fill fraction against elevation for CCFs in the northern hemisphere, in different crater-diameter bins. Red lines denote LR, with corresponding R^2 and p -values, are labelled for each plot.



A.7. Scatter plot of V_{CCF} against elevation for CCFs in the southern hemisphere, in different crater-diameter bins. Red lines denote LR, with corresponding R^2 and p -values, are labelled for each plot.



A.8. Scatter plot of fill fraction against elevation for CCFs in the southern hemisphere, in different crater-diameter bins. Red lines denote LR, with corresponding R^2 and p -values, are labelled for each plot.