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The Martian Crustal Magnetic Field

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Mars' crustal magnetic field holds information on the planet's interior evolution and exterior processes that have modified the crust. Crustal magnetization records an ancient dynamo field that indicates very different interior conditions in the past, possibly linked to the presence of a thicker early atmosphere. Current data sets have provided a wealth of information on the ancient magnetic field, and on the acquisition and modification of magnetization in the crust. However, many puzzles remain regarding the nature and origin of crustal magnetization, and the timing and characteristics of the past dynamo. Here we use recent advances in understanding martian magnetism to highlight open questions, and ways in which they can be addressed through laboratory analysis, modeling and new data sets. Many of the outstanding key issues require data sets that close the gap in spatial resolution between available global satellite and local surface magnetic field measurements. Future missions such as a helicopter, balloon or airplane can provide areal high resolution coverage of the magnetic field, vital to major advances in understanding planetary crustal magnetic fields.

Keywords: Mars, planetary magnetism, magnetic fields, crustal magnetism, ancient mars

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1 INTRODUCTION

The lack of a present-day global magnetic field on Mars was first indicated by the Mariner IV fly-by (Smith et al., 1965), and confirmed over three decades later by data from the Mars Global Surveyor (MGS) mission (Acuna et al., 1999). The clear identification of crustal magnetic fields in MGS data provided unexpected evidence for a global dynamo field in the past (Acuna et al., 1999). A remarkable aspect of crustal magnetic fields at Mars is that they are about an order of magnitude stronger than their terrestrial counterparts. Whether this reflects differences in dominant magnetic mineralogies, and/or larger volumes of magnetized material, and/or dynamo field strengths at the time of remanence acquisition that exceed Earth's present field, is still unknown (Acuna et al., 1999; McEnroe et al., 2004; Dunlop and Arkani-Hamed, 2005).

Crustal magnetic fields carry information on the interplay of a variety of processes during planetary evolution. First, the magnetization of crustal rocks reflects the composition of the crust (in particular its iron mineralogy), and its modification over the past ~4.5 Gyr, by tectonic, magmatic, impact, fluvial, and hydrothermal processes. Second, crustal magnetic fields record the history of the deep interior, specifically the longevity, morphology, and strength of the core dynamo. This, in turn, constrains core composition and dynamics, as well as the evolution of heat flow and mantle dynamics. Third, crustal magnetic fields are critical sources of information on past and present habitability from two standpoints. Constraints on dynamo history and properties are key pieces of information in the poorly-understood links between atmosphere evolution and the presence of a global magnetic field. Furthermore, the degree to which crustal magnetic fields interact with, or

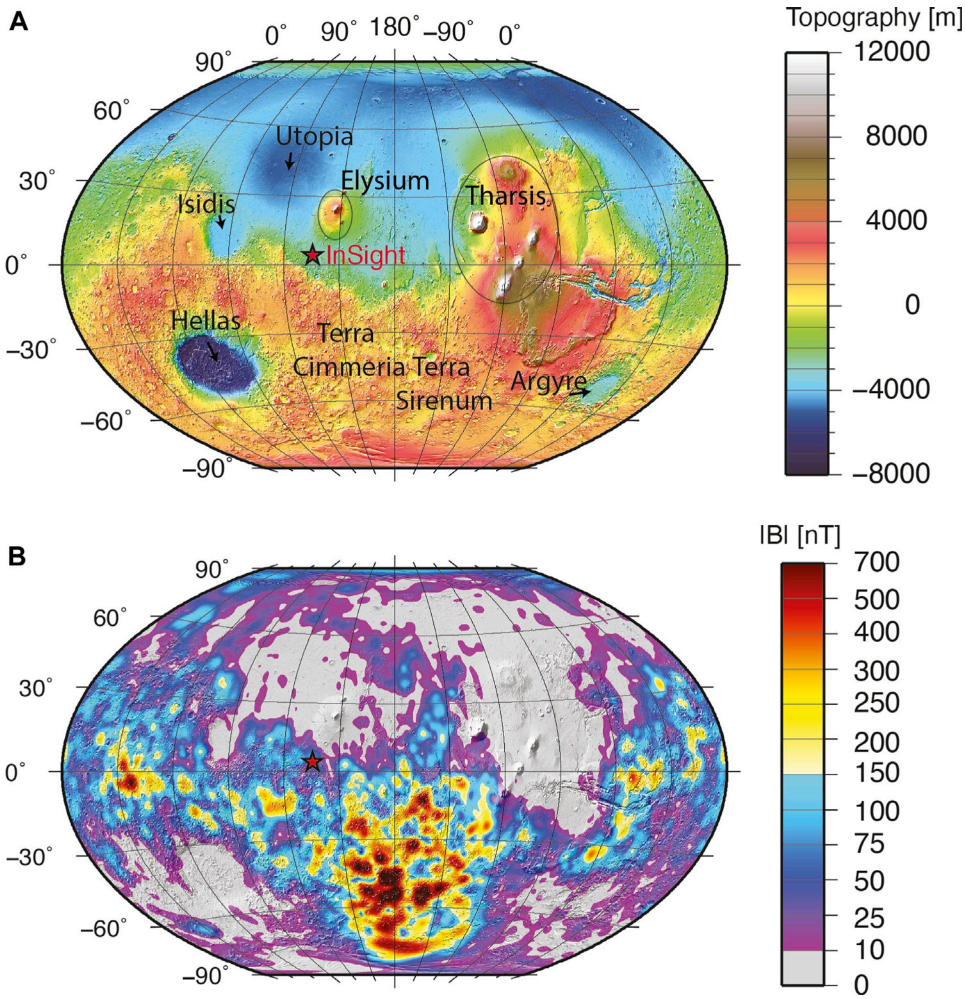

FIGURE 1 | (A) MOLA topography showing the major physiographic features on Mars including the North-South dichotomy, the impact basins, Hellas, Utopia, Isidis and Argyre (black arrows), and two major volcanic regions, Tharsis and Elysium (black circles). The InSight landing site is shown (red star). **(B)** Magnetic field amplitude in nT predicted at 130 km altitude after (Langlais et al., 2019).

deflect, harmful solar energetic particles may affect habitability and is an important consideration in ongoing space exploration efforts.

Crustal magnetization acquired during an active dynamo period can be a thermal, shock or chemical remanent magnetization (TRM, SRM, and CRM). For example, hot lava that cools through a critical temperature (the Curie temperature), can acquire a TRM with direction and amplitude determined by the dynamo field present at the time of cooling. Similarly, magnetized rock will become demagnetized if it is reheated in absence of a global magnetic field. Shock resulting from an impact, or chemical alteration from e.g., rock-water interactions, will lead to SRM and CRM respectively. Investigating the timing of ancient fields and the nature of crustal magnetization thus forms the basis for understanding planetary dynamos, in addition to concurrent and subsequent surface and subsurface conditions. For Mars, this is evidenced even at the global scale by the spatial relationships of crustal fields with major physiographic provinces – the dichotomy, impact basins and major volcanic provinces – summarized below.

The most obvious large-scale geological feature on Mars is the North-South dichotomy, observed in a range of data sets, and clearly seen in the topography difference between the southern highlands and the northern lowlands (**Figure 1A**; Smith et al., 1999). The southern hemisphere is more heavily cratered and thus has an older surface age than the northern hemisphere. The origin of the hemispherical difference is unclear, and exogenic (from a large impactor; Andrews-Hanna et al., 2008; Marinova et al., 2008) and endogenic (degree-1 convection; Zhong and Zuber, 2001; Roberts and Zhong, 2006) processes have been suggested. The dichotomy is hypothesized to be the oldest geological feature on Mars (Bottke and Andrews-Hanna, 2017). Overall, crustal magnetic fields reflect the dichotomy structure; they are primarily associated with the southern hemisphere (Acuna et al., 1999; Acuña et al., 2001) (**Figure 1B**) and have been interpreted as resulting from magnetization acquired either when the southern hemisphere crust formed (Acuna et al., 1999; Acuña et al., 2001; Arkani-Hamed, 2004), or from later intrusions (Schubert et al., 2000).

Four basins, Hellas, Utopia, Isidis and Argyre (HUIA), are the result of large impacts around 4 Ga ago (**Figure 1A**; timeline in **Figure 5A**); two (Hellas and Argyre) are in the southern hemisphere, one (Utopia) is in the northern hemisphere and one (Isidis) straddles the dichotomy boundary. From orbit, only very low amplitude magnetic anomalies, close to the noise level in the data, have been observed above these basins. This has typically been interpreted as evidence for unmagnetized crust and hence the lack of a dynamo field at the time of the impacts (Acuna et al., 1999; Schubert et al., 2000; Lillis et al., 2013; Vervelidou et al., 2017).

Volcanism has persisted throughout Mars' history, most notably recorded in the large Tharsis and Elysium provinces. Although the bulk of the Tharsis volume was emplaced by ~3.9 Ga (Phillips et al., 2001), fresh lava flows provide evidence for geologically young activity. Young volcanic activity is consistent with a general lack of magnetized crust in such regions because of the demagnetizing effect of temperature in absence of a global magnetic field (Arkani-Hamed, 2004; Johnson and Phillips, 2005; Lillis et al., 2009). Crustal fields over the southernmost portion of Tharsis suggest that at least some of the underlying crust is magnetized, but the precise relationship of the dynamo timing to Tharsis' evolution cannot be determined from these observations alone (Johnson and Phillips, 2005). Signatures above other volcanoes and lava flows have been found, and taken as evidence for a dynamo at a time recorded by the surface units, but it is generally challenging to ascertain whether older underlying intrusions or surficial flows are magnetized (Lillis et al., 2006; Langlais and Purucker, 2007; Hood et al., 2010).

Lastly, morphological evidence for fluvial activity is provided by a record of valley networks and erosional features, most of which are dated around 3.7 Ga (Fassett and Head, 2008), and indicate very different climatic conditions on early Mars (e.g., Pollack et al., 1987; Palumbo et al., 2020; Grau Galofre et al., 2020). Valley networks mostly occur in the Southern hemisphere and along the dichotomy boundary (Hynek et al., 2010), and it has been noted that there are some spatial associations with stronger crustal fields (Jakosky and Phillips, 2001; Harrison and Grimm, 2002) (see **Figure 4B**).

In summary, there is, and has been for some time, general consensus that an early global dynamo was active at some point during the first ~1 Ga of martian history (Acuna et al., 1999; Arkani-Hamed, 2004). However, the details of that history, its relationship to other planet-scale processes, and the nature of the crustal record remain elusive, despite their pivotal importance in establishing a full understanding of the planet's evolution. Here we touch on some of the "hot topics" in martian magnetism, motivated by new data sets that highlight knowledge gaps. As such we do not provide a complete review of literature on Mars' magnetic field, but first summarize data sets that have enabled the state of knowledge (**Section 2**) and general limitations of current data and modeling efforts (**Section 3**). We then address outstanding puzzles and recent developments that point to or start to address, one or more of these puzzles (**Section 4**). We separate this discussion into aspects related to the distribution and origin of magnetization (**Section 4.1**) and the ancient dynamo (**Section 4.2**). Finally, we discuss issues for which progress in the next decade is possible through cutting-edge laboratory measurements, new analysis/modeling approaches, numerical simulations of planetary and core dynamo evolution, and dedicated observation programs on future missions (**Section 5**).

2 DATA

Since the discovery of crustal magnetic fields in MGS data, one other orbital mission at Mars, Mars Atmosphere and Volatile Evolution (MAVEN; Jakosky et al., 2015), has provided magnetic field data. In addition, the Interior Exploration using Seismic Investigation, Geodesy and Heat Transport (InSight; Banerdt et al., 2020), mission has provided the first measurements of the surface magnetic field since landing in 2018 (**Figure 2**; Johnson et al., 2020). Although the Chinese Zhurong mission has landed and taken magnetic field measurements on the surface (Du et al., 2020), no reports of these data have been published at the time of writing. Furthermore, martian meteorites provide a unique source of information on crustal magnetism.

2.1 Orbital Data and Modeling

Mars has an ionosphere with which the interplanetary magnetic field interacts especially on the day-side, and so night-time magnetic measurements are preferentially used in identifying fields of crustal origin. We focus here on results from data collected at night in the shadow of the planet.

2.1.1 MGS

MGS operated from 1997–2006 and spent most of its time in orbit at approximately 400 km altitude (**Figure 2B**) and at 87° inclination, crossing the equator at 2 am and 2 pm local time, the so called Mapping Orbit (MO; 1999–2006). Magnetic field data were also obtained at lower altitudes early in the mission during Aerobraking (AB) and the Science Phasing Orbit (SPO). However, other than some night–time measurements over the South polar region, all data below 400 km were obtained during the day-time (see **Figure 1** in Mittelholz et al., 2018a). Thus, MGS provided repeated coverage of the magnetic field at 400 km altitude, yielding a data set that is ideal for robust characterization of crustal fields because outliers can be identified and removed, but that lacks low altitude information.

2.1.2 MAVEN

MAVEN has been in orbit for over 7 years (2014—present) providing a complementary data set to that of MGS because the MAVEN spacecraft orbit is highly eccentric (**Figure 2B**), and the body-fixed location and local time of periapsis evolve with time. Deep–dip campaigns have provided data at approximately 110 km altitude but over only a few locations; however, almost

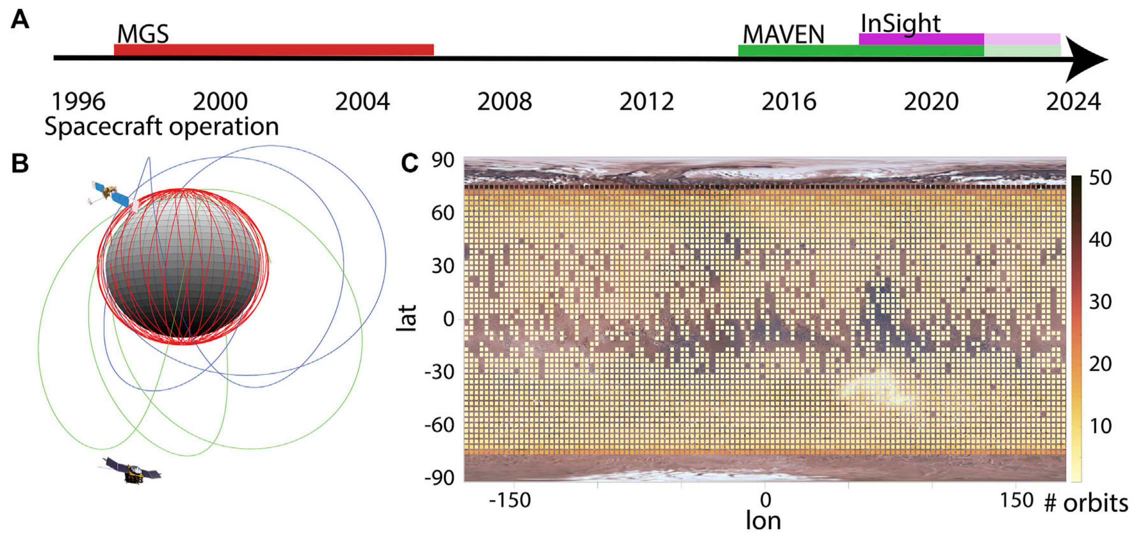

FIGURE 2 | (A) Timeline for missions that have returned information on Mars' crustal magnetic field. **(B)** Representative orbits for MGS (red) and MAVEN (green, blue) **(C)** MAVEN data coverage at altitudes less than 150 km, for data taken until October 2021. Color scale indicates the number of MAVEN orbits crossing each 3° latitude by 3° longitude bin.

complete coverage is available at 150 km altitude (**Figure 2C**). The orbital inclination is 75°, and so data over the polar regions are not available. The MAVEN data set provides a wider range of altitude and local time coverage over any given region of the planet than that of MGS, specifically substantial data at altitudes less than 400 km. However, it does not provide multiple repeat measurements at the same altitude and local time, and so data contaminated by ionospheric fields are less easily identified.

2.1.3 Crustal Magnetic Field Models

Satellite observations have led to the generation of many crustal magnetic field models, that allow projection of data onto a constant altitude and downward-continuation to predict the surface magnetic field. MGS data have enabled global (e.g., Cain, 2003; Langlais et al., 2004; Morschhauser et al., 2014) and local (Plattner and Simons, 2015) models of the field using different techniques such as equivalent source dipoles (e.g., Langlais et al., 2004) or spherical harmonics (SH; e.g., Morschhauser et al., 2014). Lower altitude night-time MAVEN data have resulted in models with improved spatial resolution (Mittelholz et al., 2018a; Langlais et al., 2019; Gao et al., 2021). The minimum wavelength in the field visible from orbit can be approximated by the orbital altitude and so models including MAVEN data currently resolve structure globally at ~150 km (equivalent to SH degree 142). The measured crustal field is proportional to the product of the thickness of the magnetized layer and the magnetization, and so for a given layer thickness, magnetization maps are a further outcome from modeling studies (Vervelidou et al., 2017; Langlais et al., 2019). However, in general, the thickness of the magnetized layer is unknown and most models assume that the magnetization is uniformly distributed over a 40-km-thick crustal column.

2.2 Surface Data

InSight landed in November 2018 and as part of the Auxiliary Payload Sensor System (APSS) which characterizes the environment around the lander, it carries the InSight Fluxgate Magnetometer (IFG) (Banfield et al., 2018). The IFG is a 3-component sensor and has provided the first surface measurements of the magnetic field (Banfield et al., 2018; Johnson et al., 2020). The average measured field strength is ~2000 nT, interpreted to reflect the ambient static crustal field (Johnson et al., 2020). This is superposed by time-varying fields of a few to tens of nT that are of non-crustal origin (Mittelholz et al., 2020b; Johnson et al., 2020). InSight thus provides a crustal field measurement at a single location that can be compared with satellite-based predictions, and in particular provides important information on the contribution of magnetization at spatial scales too small to be detectable at satellite altitudes (**Section 3.2**).

2.3 Meteorites

To date, and until the return of martian samples, laboratory analyses of martian rocks have relied on meteorites. Only a few samples have provided plausible paleointensity estimates, the orthopyroxenite ALH 84001, shergottites Shergotty and Tissinit and the naklites Nakhla and Miller Range 03346 (Cisowski, 1986; Kirschvink et al., 1997; Shaw et al., 2001; Antretter et al., 2003; Gattacceca et al., 2013; Volk et al., 2021); for most of these rocks the young formation ages (< 1,300 Ma) suggest that the magnetization of a few μT was not acquired in a dynamo field (Cisowski, 1986; Shaw et al., 2001; Gattacceca et al., 2013; Volk et al., 2021).

The paleofield estimates derived from these meteorites are consistent with satellite-based predictions for the surface crustal magnetic field in some regions of Mars (mostly in the southern hemisphere), but exceed satellite-based predictions in the proposed meteorite source regions (Langlais et al., 2019).

However, additional contributions to the local crustal field from magnetizations unresolved at satellite altitudes have been observed at the InSight landing site (see **Section 3.2**) and are likely elsewhere. Furthermore, models for magnetization of younger overlying material by older subsurface magnetized layers can explain the paleointensities inferred from the Miller Range 03346 (Volk et al., 2021). One of the oldest known meteorites (formed before ~4 Ga; Treiman, 2021) is ALH 84001 and in contrast to others mentioned above it contains magnetizations interpreted to have been acquired in a martian paleofield at ~4 Ga, that had an Earth-like paleointensity (Antretter et al., 2003; Weiss et al., 2002, 2008).

In general, studies of the magnetic properties of martian meteorites have yielded important insight, despite limited information on their origin and on the timing of magnetization. A range of magnetic carriers have been identified including magnetite, hematite, titanohematite, titanomagnetite, pyrrhotite, and maghemite (Dunlop and Arkani-Hamed, 2005; Gattacceca et al., 2014; Rochette et al., 2005, 2001). Further possible candidate mineralogies are titanomagnetite-ilmenite exsolution lamellae which have been shown to acquire strong magnetizations on Earth (McEnroe et al., 2004). The relevant rock magnetic properties vary among these minerals, and include the Curie/blocking temperatures, saturation magnetization and the ability to retain stable magnetizations over long periods of time. Specifically the Noachian-aged breccia "Black Beauty", NW 7034 (~4.4 Ga), appears to be unique in terms of its magnetic properties. It is highly oxidized (possibly the result of near-surface hydrothermal alteration), has a high unblocking temperature and high saturation isothermal remanence magnetization, and thus may be a candidate source lithology for strong magnetizations (Gattacceca et al., 2014). Ancient clasts of this meteorite experienced Fe-Ti oxide fractionation implying progressive oxidation where some rock forming material might have been directly implanted by chondritic projectiles (Deng et al., 2020). However, although oxygen isotopic composition indicates rock interaction with ^{17}O-rich water reservoirs, the data do not allow distinguishing between water equilibrating with photochemical products from an early atmosphere or water delivered by impact events. Deng et al. (2020) highlight that the enrichment in siderophile elements and fracturing of the martian crust as inferred at the InSight landing site (Lognonné et al., 2020) is consistent with the latter origin.

3 CHALLENGES

Some of the most fundamental issues in understanding the record of martian magnetization are concerned with its nature, specifically the spatial scales, depth extents and mineralogical carriers of coherent magnetization and how they vary geographically. These, in turn, are linked to the broader issues of the conditions under which magnetization has been acquired or lost over time, and hence to both the crustal and dynamo histories. In what follows we outline some of the general challenges to understanding martian magnetization and the dynamo history, imposed mainly by current data limitations.

3.1 Meteorites

Because meteorites are currently the only direct source of knowledge of magnetic mineralogy on Mars, they play a key role in interpretation of global and local magnetic field data. However, although some studies have associated individual meteorites with specific locales on Mars (e.g., Werner et al., 2014; Kereszturi and Chatzitheodoridis, 2016; Lagain et al., 2021), in general the unknown provenance of samples presents a fundamental challenge. Further, the original natural remanent magnetization (NRM) has often been adversely affected by the impact process on Mars, weathering at Earth's surface, and strong magnets used in meteorite hunting/collection.

3.2 Magnetic Field Spatial Resolution

As noted earlier, satellite data provide information on spatial scales comparable to, and larger than, the spacecraft altitude. This limits geographical correlations of magnetization with geological features that can provide clues regarding the magnetization source regions and, if a dateable surface layer is magnetized, can further constrain dynamo timing. An outstanding question is thus: how much magnetization is carried at scale lengths currently unresolved in orbital data, i.e., less than ~150 km?

An indication that substantial magnetization is carried at these smaller spatial scales is the surface field strength at the InSight landing site: the measured ~2000 nT (**Figure 3**), is about an order of magnitude larger than predictions from models based on satellite data (Mittelholz et al., 2018a; Smrekar et al., 2018; Johnson et al., 2020). Even comparisons among data from different satellite altitudes show the effect of resolving shorter wavelength structure. For example, a model that includes MAVEN data taken at altitudes as low as 150 km (Langlais et al., 2019), c. f. a model that is primarily based on data collected at 400 km altitude (Morschhauser et al., 2014), shows an approximately three-fold increase in surface field strength (**Figure 3**).

The importance of being able to detect shorter wavelength magnetization is show-cased by the size-frequency distribution of craters. The presence or absence of magnetization in the interior of craters c. f. the surrounding terrain has been used to constrain dynamo timing (Vervelidou et al., 2017; Lillis et al., 2013) (**Section 4.2.1**). Many small craters exist, yet the spatial resolution of magnetization models means that only craters with diameters larger than 300 km have been investigated in such studies (Lillis et al., 2013; Vervelidou et al., 2017). For example, out of the 384,343 craters catalogued by Robbins et al. (2013), only 21 have diameters larger than 300 km (excluding the 4 basins HUIA) and are resolvable in MGS data, whereas 104 have diameters larger than 150 km, resolvable with MAVEN data. Data at even higher resolution are needed to associate spatial variations in magnetization with smaller craters, individual volcanic units and constructs, and across tectonic features (e.g., Milbury et al., 2007, 2012; Langlais et al., 2019; Mittelholz et al., 2020a). Spatial correlations with specific geological features at these smaller spatial scales are important because they are suggestive of

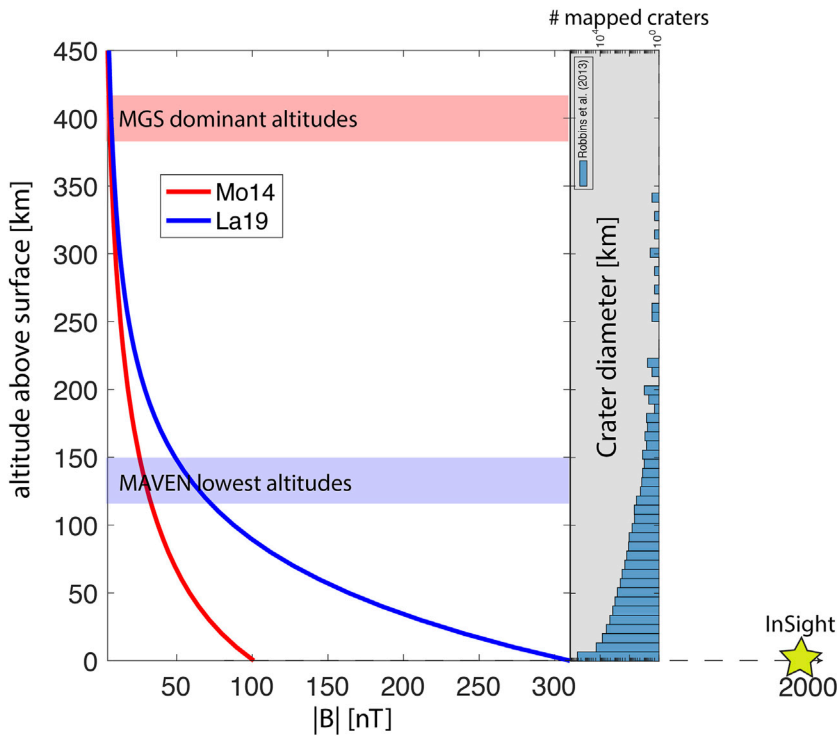

FIGURE 3 | Models of the crustal magnetic field predicted at the InSight landing site. Mo14 (red) includes only MGS data (Morschhauser et al., 2014), La19 also includes MAVEN data (Langlais et al., 2019). The shaded red/blue zones highlight the altitudes of data acquisition to indicate approximate spatial resolution of respective models. The surface measurement is shown by the yellow star. A crater size distribution from Robbins et al. (2013) shows an example of the increase in the number of smaller surface features for which magnetic analyses are only possible with improvements in data resolution.

magnetization (or demagnetization) ages that correspond to the age of the feature, and are thus key to understanding dynamo timing.

3.3 Magnetization Source Depths

A related issue stems from information about the geometry and depth of the magnetization sources that is contained in the power spectrum of a magnetic field. Statistical models in which the magnetization is confined to randomly-placed magnetized dipoles, prisms, or spherical caps (e.g., Gong and Wieczorek, 2021; Lewis and Simons, 2012; Voorhies, 2008), allow a theoretical power spectrum to be developed and fit to observation-derived magnetic field models. Source parameters include magnetization depth (either a single depth, or upper and lower limits of a magnetized layer), the spatial scale(s) of coherently magnetized regions, and the magnetization strength (or equivalently its depth- or volume-integral). In such approaches only resolvable wavelengths in the magnetic field can be exploited in the fitting procedure. Moreover, investigations of source depths on a region-by-region basis require localization of global magnetic field models before the fitting is implemented, and the localization process further limits the maximum resolution available (Lewis and Simons, 2012; Wieczorek, 2018; Gong and Wieczorek, 2021). For the Moon, high resolution mapping of the magnetic field has led to estimates of magnetization depth that are well-constrained, including the upper surface of the magnetized layer and its thickness (Wieczorek, 2018). For Mars, source depth estimation using such a spectral approach was first attempted by Lewis and Simons (2012), using an early MGS-based spherical harmonic model (Cain, 2003). More recent spherical harmonic models have not only extended the spectrum to higher resolution (Langlais et al., 2019) but also revised the spectral slope of the earlier model (Cain, 2003), showing the importance of higher spatial resolution data. However, even the recent magnetic field model resolution (Langlais et al., 2019) is insufficient to allow estimation of the top and bottom of the magnetized layer, but yields a single depth estimate for approximately the layer midpoint, often with large uncertainties (Gong and Wieczorek, 2021). Thus other information is needed to constrain the magnetized layer thickness, and in turn the magnetization strength.

Knowledge of magnetization source depths is key to identifying the host units, and constraining the timing of magnetization and plausible magnetic mineralogies. Establishing where, and how much, magnetization is carried by near-surface units is critical, because in such cases the surface age can be used to establish a magnetization chronology. Shallow magnetization could be carried by surface or near-surface lava flows, impact sheets or basin ejecta. Deep magnetic sources could be indicative of magnetization acquisition during the formation of primordial crust, or by later magmatic intrusions. Retention of a magnetization over time, requires that the magnetic minerals have not been reheated above their Curie temperature. In a general 1D sense, planetary heat flow decreases with time (e.g., Schubert et al., 2000; Plesa et al., 2018), and so the depth to the Curie isotherm increases for a given mineralogy. However, magmatic intrusions can result in local or regional reheating. For Mars, this has been suggested to have been important, specifically as a mechanism for demagnetizing the lower crust, in some regions until as recently as 1 Ga. Assuming global monotonic cooling after 3.5 Ga, TRM has been stable to depths of at least 25 km for pyrrhotite, 50 km for magnetite, and 60 km for hematite (Dunlop and Arkani-Hamed, 2005). These depth estimates are inferred from 1D parameterized thermal evolution models, and they could locally be thinner or thicker depending on the local heat flow. Combining estimates of source depth with mineralogical constraints can thus inform on likely magnetic carriers for individual regions. Improved source depth information can also better elucidate the relative contributions of chemical, shock and thermal remanence in different regions, discussed more in **Section 3.4** below.

3.4 Contributions of Different Remanence Acquisition and Modification Processes

The relative importance of thermal, chemical, or shock processes to the acquisition and modification of remanent magnetization in different geologic settings is intimately linked to the thermal, climate and impact history of Mars. Thermoremanence is typically assumed to be the main type of remanence (e.g., Acuna et al., 1999; Bowles et al., 2009; Milbury and Schubert, 2010) motivated by the evidence for widespread volcanic activity on early Mars, the efficiency of TRM in producing strong magnetizations (Dunlop and Arkani-Hamed, 2005), and the theoretical basis for understanding TRM acquisition by some likely carriers (e.g., single domain or pseudo-single domain magnetite grains; Dunlop and Özdemir, 2001). SRM (Hood et al., 2003) and CRM (Harrison and Grimm, 2002; Scott and Fuller, 2004; Quesnel et al., 2009) have also been proposed, the extent to which they play a role in a given region is unknown but key to unraveling the magnetization history of impact craters (**Section 4.1.2**) and volcanic regions (**Section 4.1.3**).

Impacts can demagnetize formerly magnetized crust in the absence of a dynamo field or (re)magnetize crust in the presence of a dynamo field through both thermal and shock effects (Halekas et al., 2002; Hood et al., 2003; Kletetschka et al., 2004; Mohit and Arkani-Hamed, 2004; Langlais and Thébault, 2011). To assess the extent to which reheating and pressure-related resetting of magnetization might have occurred, models for the temperature and pressure effects of a given impact are needed, as well as knowledge of the relevant rock magnetic properties–the Curie temperature and coercivity. In general, for large basins, thermal resetting of magnetization is restricted to basin interiors for any magnetic mineralogy (Mohit and Arkani-Hamed, 2004); however the effects of shock are less easy to determine. Shock associated with basin-forming impact events is poorly understood due to a lack of similarly sized analogue sites on Earth and challenges in numerical modeling of such impacts (e.g. Marinova et al., 2008). Uncertainties in impact parameters and target (i.e. crustal) properties lead to a range of predictions for the pressure decay of the shock wave from the impact site. For example, Mohit and Arkani-Hamed (2004) compare the predictions of empirical scalings for pressure for Hellas-, Argyre- and Isidis-forming impacts with the inferred 0.8–1.4 basin radii demagnetization region from MGS-derived crustal field models. They infer that high coercivity carriers such as single domain magnetite or multi-domain hematite are required to explain the observations. In contrast, Hood et al. (2003) suggest that pressure effects have modified the crustal magnetization record out to 3-4 basin radii at Argyre and Hellas, requiring the presence of lower coercivity magnetic carriers such as pyrrhotite.

CRM is particularly interesting because most CRM processes considered for Mars involve hydration reactions and are thus linked to early climate, and because CRM has been suggested to have a sufficiently high magnetization efficiency to explain the strong crustal fields over the southern hemisphere (McClelland, 1996; Scott and Fuller, 2004). Aqueous alteration processes mobilize metals in the matrices of rocks and can alter mineral phases hosting magnetic properties if the alteration process is sufficiently pervasive. Morphological and compositional evidence suggests that water-rock reactions were common during the Noachian (Amador et al., 2018; Ehlmann et al., 2011; Hynek et al., 2010; Fassett and Head, 2008). Specifically, the identification of extensive valley networks and morphological features suggest that liquid water once flowed on Mars under very different climatic conditions from present (Hynek et al., 2010; Fassett and Head, 2008). Laboratory studies of meteorite ALH84001 confirm early extensive aqueous alteration, and several alteration stages while it was part of the crust (Moyano-Cambero et al., 2017). For example, carbonate globules that formed by precipitation from a Mg- and Fe-rich aqueous solution, are found in cracks in the rock. There is also ample evidence for hydrothermal and deep aqueous alteration (Ehlmann et al., 2011; Ojha et al., 2021; Viviano et al., 2013), that could have allowed serpentinization (Quesnel et al., 2009; Amador et al., 2018) of ultramafic olivine-rich rock to produce magnetite. The associated release of H_2, followed by H escape to space has been suggested as an important atmospheric loss process for early Mars (Chassefière et al., 2013). As noted in **Section 1**, spatial correlations of valley networks and magnetic fields (**Figure 4B**) have been used to suggest that CRM might have played an important role in magnetization of the crust (Harrison and Grimm, 2002). Key intertwined issues in understanding the role of CRM are thus the depths of magnetization, the magnetic minerals present and the maximum depth of hydrothermal circulation. The latter depends on the depth of pore closure, which varies with time during martian history (Gyalay et al., 2020). Viscous pore closure is highly dependent on heat flow and the maximum depth of pore closure increases as the planet cools. As a result, pore closure depth today is larger than in the past, i.e., when the dynamo operated, and hydrothermal circulation would have been constrained to a thinner layer. For example, Gyalay et al. (2020) found that for a range of crustal properties at the InSight landing site, the pore closure depth at 4 Ga was 10–30 km.

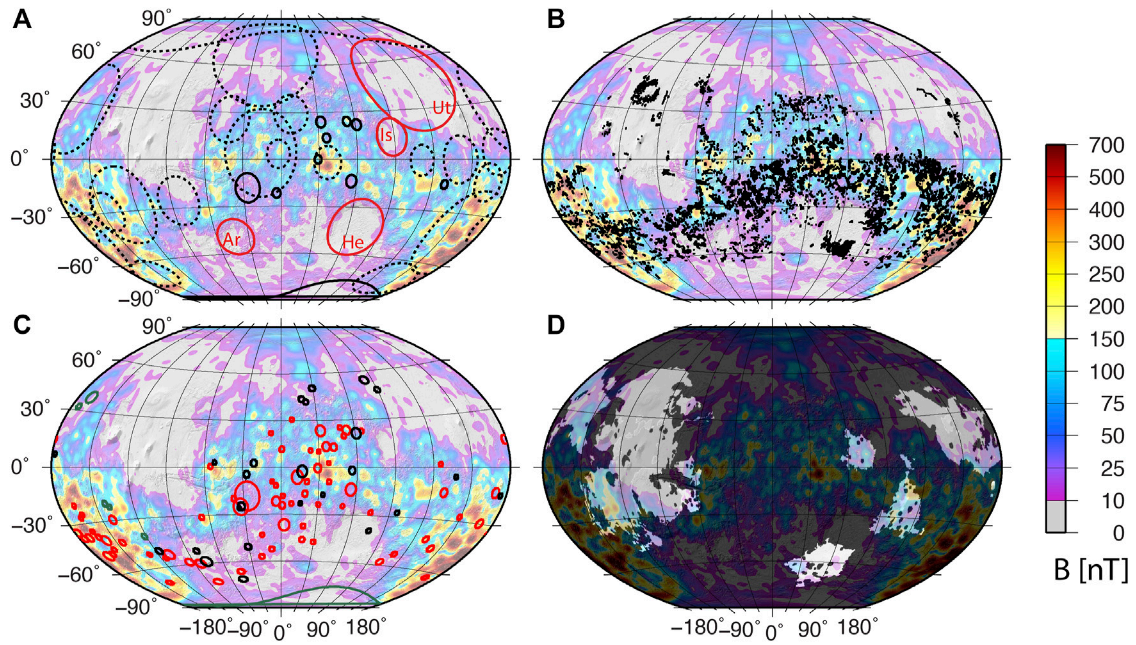

FIGURE 4 | The amplitude |B| of the crustal magnetic field as shown in **Figure 1** overlain by **(A)** large impact basins from Robbins et al. (2013) (solid black) and Frey (2008) (dashed black) and the basins Hellas (He), Utopia (Ut), Isidis (Is) and Argyre (Ar) highlighted in red **(B)** the valley network (Hynek et al., 2010) **(C)** craters larger than 150 km (Robbins and Hynek, 2012) color-coded by underlying layer age (Tanaka et al., 2014) (Pre-)Noachian = red, Hesperian = black, Amazonian = green **(D)** highlights volcanic areas as defined by Tanaka et al. (2014)

4 IMPLICATIONS

Although there are limitations associated with current data sets, crustal magnetic measurements and their distribution have contributed significantly to understanding the geophysical evolution of Mars. We next highlight outstanding questions (also summarized in **Table 1**) and discuss recent studies related to the distribution and origin of magnetization in the crust (**Section 4.1**) and the ancient dynamo (**Section 4.2**). Collectively, these motivate future avenues of research outlined in **Section 5**.

4.1 Distribution and Origin of Magnetization

4.1.1 Hemispheric Dichotomy

The large-scale pattern of magnetization follows the hemispheric dichotomy, suggesting that the origin of the dichotomy and crustal magnetization are linked, and raising several interconnected questions (**Table 1**): (H1) Was a martian dynamo active at the time the dichotomy formed? (H2) Does the hemispheric difference in magnetic fields reflect differences in the distribution and/or types of magnetic minerals and/or the properties of the dynamo field? (H3) Can the magnetic field record help elucidate the origin (exogenic versus endogenic) of the dichotomy?

Over the past couple of years, the global data set of lower altitude MAVEN observations has enabled progress on each of these questions. MGS data alone could not conclusively demonstrate the presence of magnetic fields of crustal origin over the northern lowlands; leaving open the possibility of either no magnetization or magnetization that was unresolvable by MGS. MAVEN data have allowed identification of shorter-wavelength magnetic signatures that do not show clear correlation with any other data set such as gravity or topography (Langlais et al., 2019; Mittelholz et al., 2020a). These are inferred to reflect magnetization acquired in an active dynamo period at the time of formation of the dichotomy (Mittelholz et al., 2020a). Furthermore, Mittelholz et al. (2020a) suggested that the distribution of magnetization globally, as well as the newly-detected weak northern hemisphere anomalies could be explained by different magnetic mineralogies in the two hemispheres that would result from formation of the northern lowlands crust following a giant impact.

Source depth investigations (Lewis and Simons, 2012; Gong and Wieczorek, 2021) show that magnetization depths follow the dichotomy, and are on average 9 km in the Northern hemisphere and 32 km in the southern hemisphere (Gong and Wieczorek, 2021). Gong and Wieczorek (2021) suggest that the strongly magnetized, deep layer in the South might be ancient (pre-dichotomy) units, buried by ejecta material from the large impact that formed the Borealis Basin. In their scenario, the shallow northern hemisphere material magnetization was acquired either by the newly formed post-impact crust or by later lava flows (i.e., its precise timing is still unknown, but must be syn- or post- Borealis impact). In contrast, a degree-1 dynamo origin (see **Section 4.2**) for the magnetic signature of the dichotomy would predict hemispherical differences in surface strengths but similar magnetization depths in the South and North. However, further work is needed, in particular to narrow

TABLE 1 | Summary of key questions magnetic field data can address.

Distribution and Origin of Magnetization
(H) Hemispheric Dichotomy
• (H1) Was a martian dynamo active at the time the dichotomy formed?
• (H2) Does the hemispheric difference in magnetic fields reflect differences in the distribution and/or types of magnetic minerals and/or the properties of the dynamo field?
• (H3) Can the magnetic field record help elucidate the origin (exogenic versus endogenic) of the dichotomy?
(C) Craters
• (C1) How reliable is the large basin catalog used in previous crustal magnetic field studies?
• (C2) Does the absence of magnetic fields at satellite altitudes over crater interiors imply that they are unmagnetized, or simply magnetized below the resolution of satellite data?
• (C3) Do craters lacking magnetic fields over their interiors require the absence of a dynamo at the time of their formation, or are there alternative explanations?
• (C4) Are shock effects important to the interpretation of the crater magnetic field record?
(V) Volcanoes
• (V1) (How) can we reliably use the surface volcanic record for dating magnetization?
• (V2) Is the remanent magnetization (or absence thereof) in volcanic regions solely a TRM or a combination of TRM/CRM?
(S) Strong Magnetization
• (S1) What gives rise to the strong magnetization? Is it a thick magnetized layer, distinct mineralogy, or a combination thereof?
• (S2) What are the scale lengths and origins of coherent blocks of magnetization? Are any indicative of reversals, versus units with different lithologies, post emplacement rotations?
The Ancient Dynamo
(DT) Timing
• (DT1) When did the martian dynamo operate? Was it continuous or intermittent?
• (DT2) When did it cease permanently and how quickly did this happen?
• (DT3) How does this fit in with models for early Mars thermal evolution and, atmospheric escape and climate?
(DM) Mechanism
• (DM1) What was the driving mechanism and did it change over time?
(DC) Dynamo Characteristics
• (DC1) What was the average surface field strength, and did it vary over time?
• (DC2) Was it dominantly dipolar or did it have substantial non-dipolar contributions? If dipolar, how closely was the dipole axis aligned with the rotation axis?
• (DC3) Is there evidence in the magnetic field record for major crustal reorganization (e.g. continental drift) during the period of operation of the dynamo?
• (DC4) Did the martian dynamo reverse polarity and, if so, how frequently and what were the characteristics of the reversals (duration, global field weakening)?

source depth uncertainties and to understand the nature of the southern hemisphere magnetization (**Section 4.1.4**).

4.1.2 Basin and Crater Record

Magnetic anomalies associated with a crater are a record of the ambient magnetic field at the time at which the crater formed/cooled. The general absence of magnetic anomalies and inferred lack of strong magnetization associated with the interiors of the HUIA basins (craters >500 km diameter; Robbins et al., 2013) was observed with the first MGS magnetic field data (Acuna et al., 1999). Subsequent analysis of the entire MGS mission data over craters with diameters greater than 300 km, indicated that others also show signs of temperature- and shock-induced (de)magnetization (Lillis et al., 2013; Vervelidou et al., 2017). The ages of large craters and basins with, or without, interior magnetizations have been used to infer a dynamo cessation by 4.1–4.0 Ga (Lillis et al., 2013; Vervelidou et al., 2017). Although a clear magnetic field signature, or absence thereof, associated with a crater provides valuable information, interpretation of the cratering record requires caution. Questions that persist regarding the crater record include (**Table 1**): (C1) How reliable is the large basin catalog used in previous studies? (C2) Does the absence of magnetic fields at satellite altitudes over crater interiors imply that they are unmagnetized, or simply magnetized below the resolution of satellite data? (C3) Do craters lacking magnetic fields over their interiors require the absence of a dynamo at the time of their formation, or are there alternative explanations? (C4) Are shock effects important to the interpretation of the crater magnetic field record?

Not many large craters exist (**Figure 4A**) and they are generally old, implying that identification of young magnetization is not possible. The ages of the large craters in **Figure 4A** are 3.91–4.18 Ga (mean of 4.06 Ga) and 4.25–4.11 Ga (mean of 4.17 Ga) according to Robbins et al. (2013) and Frey (2008), respectively. Question (C1) above arises because the latter record includes 19 basins that were inferred from their muted topographic expression (referred to as quasi-circular depressions, QCDs), and were suggested to be buried by younger units leaving no expression in visible imagery and only partially intact rims. However, the existence of what was identified as a large ancient buried crater (Frey, 2008); dashed in **Figure 4A**) has since been challenged (Bottke and Andrews-Hanna, 2017). Furthermore, Bottke and Andrews-Hanna (2017) suggest that between 4 (HUIA) and at most 12 basins post-date the dichotomy, substantially reducing the number of large (>300 km) ancient craters that can provide dynamo timing constraints. Many craters of smaller diameters occur on young and old terrains (**Figure 4C**), and can be investigated with higher resolution data from MAVEN (**Section 3.2**) and potential future missions. In addition, the presence or absence of weak fields in the lower altitude MAVEN data c. f. MGS data is an important step toward addressing question (C2) above. Related to this, although the canonical explanation for the absence of magnetic fields above major basins and large craters has been the absence of a dynamo at the time they formed (question (C3) above), recent work has proposed alternative scenarios, such as excavation of "magnetizable" crustal material or episodes of weaker dynamo activity (Mittelholz et al., 2020a; Hemingway and Driscoll, 2021).

Improved understanding of the extent to which shock or thermal effects result in demagnetization of existing magnetized layers requires improved constraints on magnetization depths. For example, a strongly magnetized layer at depths greater than those affected by the shock wave and/or reheating, could retain a pre-impact magnetization. For the large basins, where transient crater diameters range from 750–1,400 km (Melosh, 1989), it is also clear that crustal modification and/or removal would be enormous. In addition, for such large basins, estimation of shock demagnetization based on scaling arguments derived for smaller impactors is likely not appropriate (similar to the Moon (Halekas et al., 2002)). However, for smaller craters, SRM could be important and future analysis over craters sufficiently large to be magnetically resolvable, but with excavation depths that do not exceed the crustal thickness locally, is required. Because of the systematic increase in excavation depths of craters with increasing diameter, future investigations of craters with a range of diameters will allow the magnetic record of the crust at different depths to be assessed.

4.1.3 Volcanoes

Analysis of volcanic regions and nearby magnetic field signatures have been used to place constraints on dynamo timing (**Section 4.2**; e.g., Hood et al., 2010; Lillis et al., 2006, 2008; Milbury et al., 2012). However, unlike the case for impacts, surface ages of volcanic constructs or flows do not necessarily record the magnetization age because "magnetic resetting" of intrusions at depth can occur with no surface record. Further, although magnetization in volcanic areas is generally inferred to be a TRM, prolonged volcanism can also fuel hydrothermal activity and chemical reactions leading to a CRM (Harrison and Grimm, 2002; Chassefière et al., 2013). Important questions regarding the volcanic record are thus (V1) (How) can we reliably use the surface volcanic record for dating magnetization? (V2) Is the remanent magnetization (or absence thereof) in volcanic regions solely a TRM or a combination of TRM/CRM?

Recent work has started to address the first of these questions. Clearly, additional source depth constraints are needed to understand whether the bulk of the magnetization is carried by near-surface or deep units. For example, regional source depth estimates for the southern hemisphere point to much of the magnetization being carried in the deeper crust (Gong and Wieczorek, 2021); however whether some magnetization could more locally be carried at shallow depths in the vicinity of volcanoes studied previously for dynamo timing (e.g., Tyrrhenus Mons; Milbury et al., 2012) is unknown. On such local scales clear indications that at least some of the magnetization is carried by a surficial unit such as the identification of a magnetized pyroclasic flow in Mittelholz et al. (2020a), require the lowest altitude MAVEN data and future high resolution studies.

4.1.4 Origin of the Strong Southern Hemisphere Magnetization

Locally, and mostly concentrated in Terra Sirenum (TS) and Terra Cimmeria (TC) (see **Figure 1**), crustal fields are much stronger than in other regions on Mars or terrestrial crustal fields measured from Earth orbit (Langlais et al., 2010; Thébault et al., 2010). These require large amplitude magnetizations even if tens of kms of the crust are magnetized. For example, Parker (2003) showed that a minimum magnetization of 4.76 A/m was required throughout 50 km of crust to explain the strong southern hemisphere fields. The vertically integrated magnetization is 238 kA, compared with at most ~10 kA for freshly magnetized mid-ocean ridge basalts (10 A/m for a ~1 km magnetized layer). Pervasive in the interpretation of martian magnetization is that the magnetized layer may be tens of km thick; however in the ocean basins most of the magnetization is carried in Layer 2A of the crust which is ~400–600 m thick (Sandwell, 2022). The main questions here are: (S1) What gives rise to the strong magnetization? Is it a thick magnetized layer, distinct mineralogy, or a combination thereof? (S2) What are the scale lengths and origins of coherent blocks of magnetization? Are there any indications of a record of reversals, versus magnetization contrasts that reflect units with different lithologies and/or post emplacement rotations?

Although the first orbital observations over the TS/TC region were inferred to result from linear magnetization contrasts akin to those in the ocean basins (Acuna et al., 1999; Connerney et al., 1999), later modeling and addition of MAVEN data revealed a more blocky magnetic field pattern (**Figure 1**; Langlais et al., 2019), in which anomalies do not have obvious correlations with surface features. Bouley et al. (2020) inferred the region to consist of discrete crustal blocks, associated with thicker crust and enriched in potassium and thorium. Because the terrain is overprinted by the big basins Hellas and Argyre, they suggest that TS/TC represents some of the oldest crustal material and possibly a distinct crustal formation mechanism compared with other regions. Source depths in the region are found to be exceptionally deep, specifically larger than 40 km (Gong and Wieczorek, 2021). However, analysis of satellite compositional data also indicates surficial iron enhancements, that have been suggested to contribute to the strong magnetizations (AlHantoobi et al., 2021).

4.2 The Ancient Dynamo

Mars' ancient dynamo has been discussed widely, because its timing and nature are directly linked to interior compositional and thermal properties. We thus review current hypotheses for dynamo timing, their implications for potential driving mechanisms, and constraints on the paleofield characteristics, in particular strength and geometry.

4.2.1 Timing

Dynamo timing raises some of the most-discussed questions regarding the martian magnetic field: (DT1) When did the martian dynamo operate? Was it continuous or intermittent? (DT2) When did it cease permanently and how quickly did this

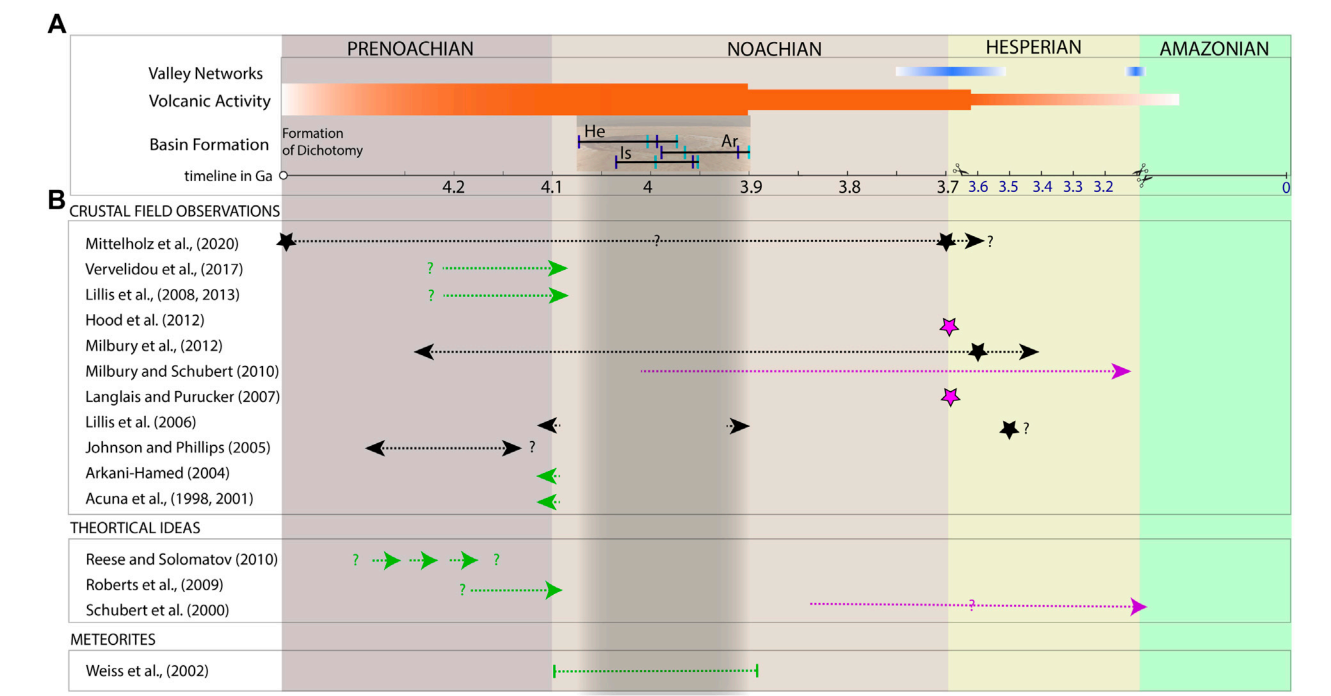

FIGURE 5 | (A) A summary timeline for major relevant volcanic, fluvial activity and basin formation events on Mars and **(B)** literature addressing dynamo timing: **(A)** Error bars represent isochron (cyan) and N (50) (blue) age estimates (Robbins et al., 2013) for Hellas (He), Isidis (Is) and Argyre (Ar) and the grey shaded area in **(A)** and **(B)** highlights the maximum time interval over which basin forming events occurred according to those estimates. **(B)** Stars represent magnetized features that suggest an active dynamo at a specific time. Arrows indicate "later/earlier than" arguments and "?" refer to unknown starting/end points. Purple/green arrows/stars indicate studies that have been used to argue for an early/late dynamo. Black arrows indicate studies that have not been used to argue for either late or early.

happen? (DT3) How does the dynamo timing fit with models for early Mars thermal evolution and atmospheric and surface conditions?

The starting point for most hypotheses regarding timing, is the assumption that the absence of substantial crustal field anomalies above the large impact basins implies that no dynamo operated when the basins formed. The traditionally most-accepted suite of studies argued that a dynamo operated prior to basin formation, the "early dynamo" hypothesis and that shock and thermal effects during basin formation erased any magnetization present. This was supported by the early observation that most magnetic signatures are seen over (Pre-)Noachian terrain (Acuña et al., 2001), on which the major basins Hellas and Argyre are superposed. The argument was further bolstered by analyses of magnetic fields over craters larger than 300 km diameter, from which the oldest magnetized crater was found to be ~4.1 Ga, older than the large basins (Lillis et al., 2013; Vervelidou et al., 2017). Other studies have argued for a "late dynamo", that started after the basins were emplaced. Magnetic signatures over Noachian terrains were interpreted as magnetized intrusions postdating the surface units. Identification of magnetic signatures above young volcanoes and lava flows also supported the "late dynamo" hypothesis (Hood et al., 2010; Langlais and Purucker, 2007; Lillis et al., 2006). **Figure 5** shows a compilation of literature addressing dynamo timing to emphasize the breadth of studies and suggested scenarios. Recently, low altitude MAVEN data have allowed identification of magnetic signatures over the northern hemisphere (**Section 3.2**) and a pyroclastic flow (**Section 4.1.3**) indicating an active dynamo at 4.5 Ga and at 3.7 Ga, respectively (Mittelholz et al., 2020a). One interpretation of these results is that a dynamo was active from 4.5 to at least 3.7 Ga, including during the period of basin formation. This was also previously suggested by Milbury et al. (2012) who hypothesized a dynamo shutdown around 3.6 Ga while still arguing for an early dynamo. An active dynamo during basin formation leaving small amplitude signatures not visible from orbit is plausible, given the large amount of crustal material excavated, and subsequent changes in amount and type of magnetic minerals in the basin floors (Mittelholz et al., 2020a). An alternative scenario is that of changing dynamo processes during and immediately after basin formation, specifically the possibility of decreased dynamo activity during the basin forming events (Hemingway and Driscoll, 2021; Mittelholz et al., 2020a).

Dynamo timing has been suggested to affect climate evolution on early Mars (Jakosky and Phillips, 2001), linking interior and surface/atmospheric processes. The change in ancient climate conditions from a thick atmosphere that allowed liquid water to persist (at least temporarily or intermittently) on the martian surface to a much thinner atmosphere (e.g., Chassefière et al., 2007; Wordsworth et al., 2015), could suggest a causal relation between atmospheric escape and a global dynamo field. However, this notion has been questioned (Brain et al., 2013, 2016). Pinning down the active periods of the dynamo, including temporary and permanent cessation, will allow this issue to be addressed further.

4.2.2 Mechanisms

The timing results summarized above prompt the major question: (DM1) What was the driving mechanism and did it change over time? Although a thermally-driven dynamo is able to explain a very early, but limited-duration dynamo (~500 Ma), extending the dynamo lifetime to 3.7 Ga and longer is challenging. Early in a planet's history, the core is much hotter than the mantle driving superadiabatic heat loss from the core that enables sufficiently vigorous convection to generate a dynamo (Breuer and Spohn, 2003; Williams and Nimmo, 2004). When the heat loss is no longer superadiabatic, additional sources of buoyancy must be found to sustain a convectively-driven dynamo. Once the core temperature drops below the solidus and inner core crystallization starts, both the latent heat (additional thermal buoyancy) and the release of light elements (compositional buoyancy) become important. For the Earth, inner core solidification (i.e., bottom-up crystallization) is the dominant driver of the present-day dynamo (Gubbins, 1977; Gubbins et al., 2003; Nimmo, 2007). Recent InSight results indicate a relatively large martian core (~1860 km), requiring a low density, and thus large concentrations of light elements (Stähler et al., 2021). While the exact amount and types of light elements in the core and mantle are currently debated, a solid inner core has not been seismically detected, and is now thought to be unlikely because it is incompatible with proposed compositional and thermal models (Stähler et al., 2021; Khan et al., 2022). The latter suggest melting temperatures that are below core temperatures inferred from thermal evolution models (e.g., Plesa et al., 2018).

Hemingway and Driscoll (2021) explore a range of dynamo scenarios for early Mars by varying parameters such as core sulfur mass fraction and thermal conductivity, light element partitioning or initial temperature. They find a wide range of possibilities for the evolution of the past and future martian dynamo, that given the large parameter space, can satisfy any of the scenarios proposed in **Figure 5**. In particular, the thermal conductivity of the core affects the duration of an early thermal dynamo. Recent experiments (Pommier, 2018; Pommier et al., 2020) suggest that the thermal conductivity of a sulfur-rich core is lower and between 5 and 30 $\mathrm{Wm^{-1}K^{-1}}$ depending on the iron alloy composition in the Fe-S and Fe-S-O-Mg-Si systems. This is lower than the 30–120 $\mathrm{Wm^{-1}K^{-1}}$ typically assumed in previous studies (e.g., Nimmo and Stevenson, 2000; Williams and Nimmo, 2004; Arkani-Hamed, 2012; Hemingway and Driscoll, 2021). Thermal evolution modeling (Greenwood et al., 2021) confirms that this has an important effect on the dynamo duration, with lower conductivity values allowing longer thermally-driven dynamo activity. In conclusion, the evolution of the core and possible dynamo mechanisms critically depend on the light element content, and thermal state and core thermal conductivity. The recent InSight results for core size and composition (Stähler et al., 2021; Khan et al., 2022) suggest the need to re-examine alternative crystallization scenarios, e.g., iron snow from possible top-down (near the core-mantle boundary) crystallization and subsequent remelting deeper in the core as well as the possibility that different driving mechanisms might have operated at different periods in martian history and whether future dynamo episodes are a possibility (Stewart et al., 2007; Hemingway and Driscoll, 2021). A further consideration is whether the thermal consequences of one or more large impacts could have had a transient effect on the core heat flow pattern that was sufficient to either initiate (Reese and Solomatov, 2010) or inhibit (Roberts et al., 2009) dynamo generation.

4.2.3 Paleodynamo Characteristics

A range of questions regarding the characteristics of the martian paleodynamo exist: (DC1) What was the average surface field strength, and did it vary over time? (DC2) Was the field dominantly dipolar or did it have substantial non-dipolar contributions? If dipolar, how closely was the dipole axis aligned with the rotation axis? (DC3) Did the martian dynamo reverse polarity and, if so, how frequently and what were the characteristics of the reversals (duration, global field weakening)?

The strength of the paleodynamo, B_{ancient}, in which a crustal magnetization, M, was acquired, is related to M, and the thermoremanent magnetic susceptibility, χ_{TRM}, by $B_{\mathrm{ancient}} = M\mu_0/\chi_{\mathrm{TRM}}$, where μ_0 is the magnetic permeability of free space. Magnetization inferred from satellite data is inversely related to the thickness of the magnetized layer; as described in **Section 3.3** the latter is commonly difficult to assess. Mittelholz et al. (2020a) identified a magnetic anomaly associated with a crater in Lucus Planum; using the crater depth, combined with geological constraints, to infer the thickness of the magnetized layer, they were able to conclude that the 3.7-Ga-old lava flow could have been magnetized in an Earth-like ancient field. This is consistent with conclusions from paleomagnetic studies of meteorite ALH84001 (Weiss et al., 2008). However, the strength of the field might have varied through time and therefore such inferences are recordings of the ancient field strength at the time at which the respective magnetization was acquired.

The ancient dynamo might also have reversed polarity, often studied in planetary magnetism via paleopole locations inferred from satellite-derived magnetic field maps. Different methods and assumptions to derive paleopole locations significantly affect the results and lead to different conclusions (e.g., Milbury et al., 2012; Boutin and Arkani-Hamed, 2006; Thomas et al., 2018). Thomas et al. (2018) present an overview of published paleopole locations (their **Figure 1**) which show a spread across the planet. In their study, uncertainties are incorporated and admissible locations of paleopoles indicate that Mars must have reversed polarity at least once and that a substantial amount of true polar wander occurred. This is consistent with reorientation of the rotation axis in response to the emplacement of Tharsis (Zuber and Smith, 1997).

Most discussions to date concerning the ancient martian field, in particular paleopole studies, assume that it was a classic dipolar field. However, one hypothesis for the north/south dichotomy in the magnetic field is that the dynamo itself produced a hemispheric field, driven by hemispheric (spherical harmonic degree 1) variations in heat flow at the core-mantle boundary (Stanley et al., 2008). This large-scale pattern in heat flow has been invoked in endogenic models for the formation of the dichotomy (Zhong and Zuber, 2001; Roberts and Zhong, 2006). Previous (Stanley et al., 2008) and more

TABLE 2 | Future key advances are linked to specific questions outlined in **Table 1**, where H=Hemispheric Dichotomy, C=Craters, V=Volcanoes, S=Strong Magnetization, DT = Dynamo Timing, DM = Mechanism, DC = Dynamo Characteristics.

Key advances	Additional Information	Specific questions addressed
Laboratory analysis and return samples		
Advances in laboratory studies of meteorites and future returned samples.	High resolution instrumentation allows paleomagnetic analysis of smaller, possibly more stable mineral phases.	S depending on sample provenance: V2 or C2, C4; DT, possibly DC.
Return samples.	Martian sample rocks with geological and geochemical context and protected from shock and alteration while in transit to, and storage on, Earth. Samples can be dated.	S depending on sample: V2 or C2, C4; DT, possibly DC.
Data Analysis and Modelling Advances		
Information from other data sets.	(e.g.) InSight's new constraints on the martian interior through seismic and geodetic information.	DT, DM, DC, possibly H1, H2; provides additional information for interpreting depths of magnetization at a given locale.
Numerical modeling advances.	Dynamo modeling. Hydrocode simulations of impacts, in particular for large basins.	DT, DM, DC, C4, possibly H1, H2.
Magnetic field data from ongoing and future missions		
Magnetometers on landers.	Local magnetic field description and magnetic field variability at the surface.	S1, S2, possibly DT.
Low altitude missions such as a helicopter, drone, balloon or airplane.	Regional high resolution data to resolve challenges related to the lack of knowledge regarding small spatial scales of magnetization and source depths.	DT, related to DM, DC; depending on location H, C, V, S.

recent (Yan et al., 2021) numerical dynamo simulations indicate that such heat flow variations can result in either a stable or reversing hemispheric dynamo. However, Dietrich and Wicht (2013) have challenged this view and argued that frequent (~10 kyr timescales) dynamo reversals in these hemispherical models would not allow strong coherent magnetization to develop. This highlights the need for future constraints on any history of magnetic field reversals on Mars.

5 OUTLOOK

Future progress on the questions discussed in **Section 4** and summarized in **Table 1**, can be made from a variety of different angles, through laboratory measurements, new analyses and modeling approaches, as well as dedicated observations from ongoing and future missions (**Table 2**).

New laboratory investigations are now possible with the advent of significant increases in instrument resolution. For example, previous analyses of martian meteorite ALH84001 used SQUID microscopy with a spatial resolution of 140 μm (Weiss et al., 2007; Weiss et al., 2008). However, the recently-developed quantum diamond microscope enables high-sensitivity, high resolution mapping of magnetic fields at spatial scales of 1 μm (Glenn et al., 2017), a two order magnitude of improvement over existing studies. This allows more detailed characterization of the history of magnetization within a sample at the microscopic level. In particular, previously-unresolvable phases that might be more capable of retaining a magnetic record over billions of years, can now be accessed and current studies are exploiting such capabilities (Volk et al., 2021; Steele et al., 2022). Furthermore, activities such as the current acquisition of samples as part of the Mars 2020 mission and their return to Earth planned for 2031 (Farley et al., 2020) will revolutionize laboratory analyses of martian rocks. Return samples are collected from a known location and the geological context provides critical knowledge in both deciding on appropriate sample analysis approaches and in the interpretation of laboratory measurements. Careful handling in sealed shielded containers can ensure minimal magnetic contamination from chemical alteration or related to the transport and reentry in the Earth atmosphere. Future analyses of these samples will allow in-depth characterizations of mineralogy and magnetized phases and the possibility of deriving absolute ages. This will shed light on mineralogies giving rise to magnetization, processes that would have led to magnetization acquisition and modification, and constraints on the timing and characteristics of the martian dynamo (see discussion in Mittelholz et al. (2018b) for more details).

Geophysical parameters derived from satellite and surface missions to Mars, particularly the In Sight mission, have led to new understanding of interior structure and updated models for the planet's thermo-chemical interior evolution. For example, seismic measurements have recently led to updated estimates of core radius and density, constraining compositional models for the core (Stähler et al., 2021), and, in combination with cosmochemical constraints, for the mantle (Khan et al., 2022). Such models provide, in turn, important constraints for numerical simulations of the core dynamo (e.g., Yan et al., 2021). MAVEN data have already enabled key advancements (**Sections 3**, **4**). The periapsis of MAVEN was raised in 2020 to ~200 km, meaning that repeat observations at the prior lowest altitudes are not possible. However, continued development of magnetic data inversion techniques (e.g., Moore and Bloxham, 2017) together with ongoing data acquisition especially at nighttime below the altitudes of the MGS mapping orbit will continue to provide information on the crustal field on a global scale. Rover missions provide local geochemical context on current and ancient surface conditions. These are important inputs for

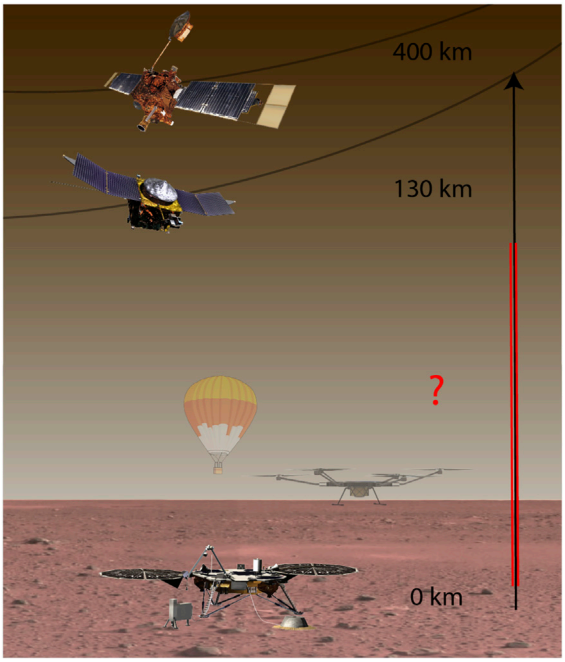

FIGURE 6 | Sketch highlighting the lack of data coverage between surface and orbital data.

understanding the crustal magnetic field as they constrain models for environmental conditions, in particular climate and water inventory (e.g., Scheller et al., 2021; Wordsworth et al., 2021), during and after the acquisition of magnetization.

Magnetic field data from planetary missions enable a variety of studies, and we now focus on what could be achieved with possible future missions. In addition to satellite coverage at orbital altitudes, InSight, Zhurong (Du et al., 2020) and ExoMars (Biele et al., 2007) (will) add local descriptions of the crustal magnetic field at their respective landing sites, specifically offering insight into contributions from short-wavelength magnetization. Sites for which geological context exists, would greatly benefit from magnetization source depth and thickness estimates. For example, even prior to detailed knowledge of the seismic structure of the crust at the InSight landing site, the local geological context including exposures of units at depth in nearby craters allowed more robust constraints on the possible depth(s) of burial and ages of magnetized units (Johnson et al., 2020). For a mission such as InSight seismic measurements have led to characterisation of the thickness of crustal layers (Knapmeyer-Endrun et al., 2021), further constraining magnetization. At the InSight landing site, assuming that a young upper layer of broken up regolith and unconsolidated material of approximately 10 km (depths at which layer transitioning has been seismically shown), overlies a magnetized layer of 10 km thickness, the magnetization would be approximately 3 A/m (Knapmeyer-Endrun et al., 2021) and would likely have been acquired in the Noachian (Johnson et al., 2020). Furthermore, we note that even magnetometers such as the one included on InSight that are not officially designated as science instruments, can, with some pre-launch calibration and magnetic cleanliness assessment, greatly contribute to magnetic field science.

Although not focus of this paper, crustal magnetic fields affect the extent to which external magnetic fields interact

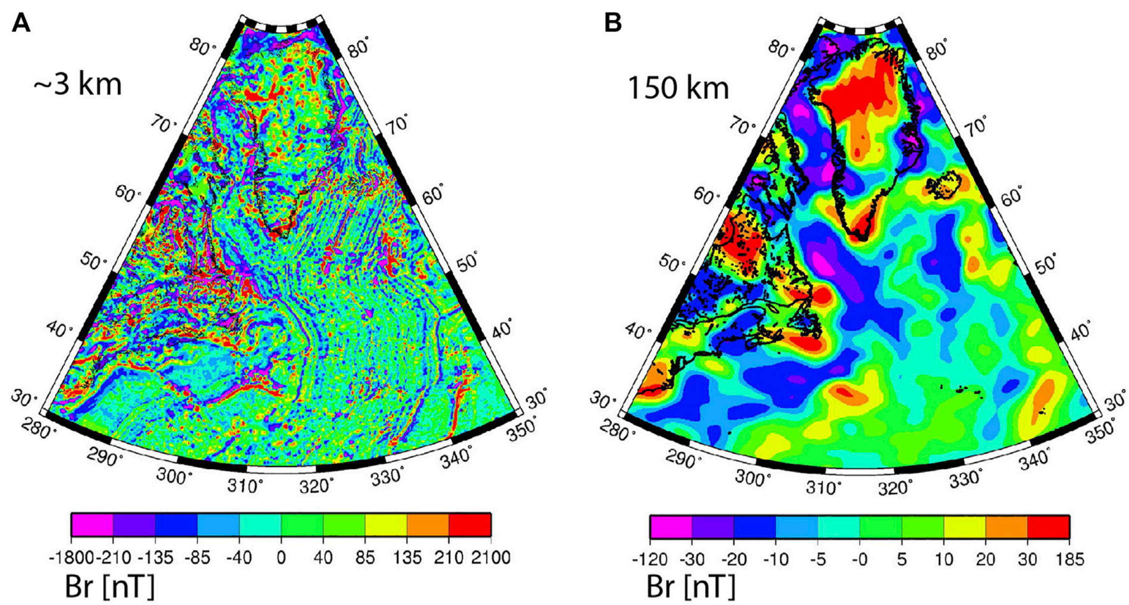

FIGURE 7 | The radial crustal magnetic field of the mid-oceanic ridge on Earth as seen **(A)** just above the surface **(B)** and from 150 km altitude. Adapted from Ravat (2011).

with the surface. While the details of this interaction are complex (e.g., Hara et al., 2018; Weber et al., 2019), it is clear that such processes are important. They can affect atmospheric escape (Fang et al., 2015; Jakosky, 2021), and also influence the radiation environment on the planetary surface which is of particular importance for future exploration efforts (Emoto et al., 2018). Hence understanding the topology of magnetic field lines and small-scale crustal field structure at a given location, as well as any changes in the local field line configuration in response to external fluctuations (Mittelholz et al., 2021b), is an important avenue for future studies. Local measurements from different locations with different crustal magnetic field geometries will be particularly valuable.

Investigations of the small-scale magnetic field structure to bridge the gap between surface and orbit observations requires increased spatial resolution and thus low altitude missions (Mittelholz et al., 2021a; **Figure 6**) enabled by drones, helicopters (Bapst et al., 2021), air planes (Braun et al., 2006) or balloons (Kerzhanovich et al., 2004; Hall et al., 2007). Mission designs such as the successful Ingenuity helicopter could potentially offer several tracks of low altitude coverage, together with multiple local surface measurements when the helicopter lands (Balaram et al., 2021; Bapst et al., 2021). **Figure 7** underscores the importance of improved resolution by comparing orbital with near-surface measurements over the ocean basins and mid-oceanic ridges on Earth. It is very obvious that a satellite-based model shown at 150 km altitude (**Figure 7B**), does not give any indication of the characteristic "stripe" pattern of magnetic anomalies in the ocean basins indicative of seafloor spreading that is recorded by the magnetic field. Ship-board data however yield crustal magnetic field models with the distinctive alternating polarity pattern, providing insight into reversals of the gloabl dipole field over time scales of up to ~200 Ma, and evidence for ongoing plate tectonics. Only near-surface measurements reveal these important features (**Figure 7A**). On Mars, any short wavelength information that could carry such information is unexplored and awaits discovery.

In **Sections 3**, **4**, we review limitations of current data sets and how it affects our current knowledge; all of the discussed studies are in some ways limited by available resolution (**Table 2**), horizontally (i.e., small wavelength structures) or vertically (i.e., source depth) and we argue that low altitude magnetometry studies would provide a leap forward in our understanding of the martian crustal field. While no such definite mission plans have been made so far, efforts in planning low altitude missions at Mars are ongoing and have been endorsed by the community (Mittelholz et al., 2021a; Bapst et al., 2021; Rapin et al., 2021). Future such endeavors will transform the view of crustal magnetism and Mars' history, by not only addressing existing key questions such as those outlined here, but identifying new ones.

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All authors listed have made a substantial, direct, and intellectual contribution to the work and approved it for publication.

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