

# An Equivalent Source Dipole Model of the Martian Crustal Magnetic Field Using MGS, MAVEN, and Tianwen-1 Data

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## Key Points:

- We developed a new Martian crustal magnetic field model, incorporating data from the MGS, MAVEN and the Tianwen-1 mission.
- The model is based on equivalent source dipoles of the Fibonacci sphere distribution and is derived from a two-step inversion process.
- The model's predictions are compared with ground-based magnetic measurements and magnetization estimates from the InSight lander and Zhurong rover.

## Abstract

Mars does not possess a present-day global magnetic field generated by an active dynamo; however, it exhibits intriguing localized crustal magnetic anomalies. High-resolution and reliable models of these remnant magnetic fields can offer critical insights into the planet's internal dynamics and evolutionary history. This study presents a novel model of the Martian crustal magnetic field, derived from magnetic field data acquired by MGS, MAVEN, and the orbiter of Tianwen-1 mission. The model is constructed using the Equivalent Source Dipole method, with dipole sources uniformly distributed on a spherical surface according to the Fibonacci Sphere Distribution. The inversion of the magnetic dipole moments involves two key steps. First, the radial components of the magnetic dipole moments are inverted using only the radial magnetic observations. Subsequently, the inverted radial magnetic moments serve as the initial model to derive the three-component magnetic moments through the iteratively reweighted least squares method with Huber weights. Our inversion scheme incorporates rigorous data selection criteria and novel mathematical methods to mitigate the influence of external magnetic sources. The inverted magnetic dipole moments are transformed into a spherical harmonic model, which demonstrates convergence of the magnetic power spectrum at the surface of Mars up to spherical harmonic degree 110, corresponding to a surface spatial resolution of approximately 190 km. The predicted values from our model are compared with on-site magnetic measurements taken by the InSight lander and the Zhurong Rover of the Tianwen-1 mission.

**Keywords:** Planetology; Mars; magnetic fields; magnetic anomalies

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## 1. Introduction

The Mars Global Surveyor (MGS) spacecraft conducted extensive measurements of the Martian crustal magnetic field from 1996 to 2006. During its initial aerobraking and science phase orbits (AB/SPO) phase, the eccentric orbit resulted in sparse data collection, primarily on the dayside of Mars. However, the subsequent Mapping Phase Orbit (MPO) involved a large number of measurements at a constant altitude of approximately 400 km (Acuña et al., 1998), providing global coverage data on both the dayside and nightside for Martian crustal magnetic field modeling. The MGS revealed that Mars does not possess a global magnetic field generated by an active dynamo. Instead, it exhibits globally distributed small-scale remanent magnetic fields that originate in the Martian crust. This magnetization is believed to have been acquired before the cessation of an ancient Martian dynamo (Lillis et al., 2013; Vervelidou et al., 2017). These magnetic anomalies on Mars are 1 to 3 orders of magnitude larger than those observed on Earth, the Moon (Purucker and Nicholas, 2010), and Mercury (Johnson et al., 2015). This difference may be attributed to a higher volume of magnetic minerals in the Martian crust or a greater magnetic susceptibility of these minerals (McEnroe et al., 2004). High-resolution and robust models of Martian crustal magnetic fields can provide important constraints on the internal structure and evolution of the planet (Mittelholz et al., 2020). Moreover, these models are pivotal in understanding atmospheric retention and space environments of Mars (Jakosky et al., 2018; Lammer et al., 2013). In particular, the crustal fields affect the large-scale geometry of the solar wind interaction. The Martian magnetotail is twisted, with the sense of the twist set by the orientation of the interplanetary magnetic field (IMF) (DiBraccio et al., 2018; Zhou JY et al., 2024), and reconnection between the strong anomalies and the draped IMF opens field lines along which ionospheric heavy ions escape (Cheng SW et al., 2025), although whether the crustal fields increase or reduce the total escape rate remains debated (Zhou JY et al., 2026). A crustal field model is an input to each of these studies, and its accuracy and spatial resolution limit the conclusions that can be drawn from them.

The earliest models of the Martian crustal magnetic field were based solely on measurements obtained from the MGS (Arkani-Hamed, 2001; Cain et al., 2003; Chiao et al., 2006; Langlais et al., 2004; Lillis et al., 2008; Morschhauser et al., 2014; Purucker et al., 2000; Whaler and Purucker, 2005). These models exhibit a dichotomy similar to the Martian topography (Acuña et al., 1999). The most intense magnetic anomalies are found in the southern highlands, while weaker anomalies are observed in the northern lowlands, which could be the result of impacts or large-scale mantle convection (Andrews-Hanna et al., 2008; Roberts and Zhong SJ, 2006). Notably, weak magnetic anomalies have also been detected in regions with large impact craters and volcanic provinces (Arkani-Hamed, 2001). These models suggest that Mars once possessed a planetary dynamo comparable in scale to Earth's, which may have eventually ceased due to the subcritical regime of its dynamo and giant impacts (Kuang WJ et al., 2008; Roberts et al., 2009).

Despite the unique features of the Martian crustal magnetic field depicted by these models, the small-scale magnetic field features are still poorly constrained. Most magnetic field measurements from the Mars Global Surveyor (MGS) were obtained at an altitude of approximately 400 km, while the data from the AB/SPO low-orbit phase are sparse and predominantly limited to the dayside, where the influence of solar wind is significant. When spherical harmonic (SH) models derived from 400 km altitude data are extrapolated downward to the surface, this can result in potential misinterpretation of small-scale magnetic field signals.

This is evident from the fact that the downward-continued magnetic fields from these models are an order of magnitude weaker than the surface measurements recorded by the InSight lander (Johnson et al., 2020). This limitation underscores the need for lower-altitude data collection to improve the resolution and accuracy of future models, and the latter MAVEN mission provides the required measurements to help address this shortcoming.

The Mars Atmosphere and Volatile Evolution (MAVEN) mission, launched in December 2013, has significantly mitigated the aforementioned challenges. MAVEN is equipped with two high-precision triaxial fluxgate magnetometers (Connerney et al., 2015). Mars hosts a relatively small magnetosphere, a result of strong remanent crustal magnetism in the absence of a present-day global intrinsic magnetic field. The MAVEN mission was specifically designed to investigate the interaction between the solar wind and Mars. As a result, MAVEN is capable of crossing the bow shock, providing partial upstream solar wind data, as well as low-altitude magnetic field measurements (Jakosky et al., 2015). The combination of MGS and MAVEN data provides a great opportunity to refine the Martian crustal magnetic field models. Langlais et al. (2019) constructed a new model of the Martian crustal magnetic field using magnetic data from MGS and MAVEN, spanning altitudes from 90 km to 600 km. Their model, based on the Equivalent Source Dipole (ESD) method, achieves a spatial resolution of approximately 160 km at the surface, corresponding to spherical harmonic degree 134. Gao JW et al. (2021) established a spherical harmonic model truncated to degree 110, utilizing magnetic field data from MGS and MAVEN collected at altitudes below 500 km. Most recently, Delcourt and Mittelholz (2025) modeled the crustal field with physics-informed neural networks, a mesh-free representation trained directly on the spacecraft data with the equations of magnetostatics imposed during training. Studies of impact crater demagnetization have estimated the dynamo shutdown time to likely be before 4.1 Ga (Lillis et al., 2013; Vervelidou et al., 2017), and Mittelholz et al. (2020) used the local crustal field model based on the ESD method to illustrate that the Martian dynamo was probably operating 3.7 Ga ago.

Each of the three approaches has its own limitation. Spherical harmonic models use a global basis whose effective resolution is set by the observation altitude and whose regularization acts globally. Equivalent source dipole models adapt to uneven data coverage, but the non-uniqueness of the source parameterization has to be handled explicitly (Section 3.3 and Appendix A). Network representations continue the field below the sampled altitudes through the trained network rather than through an analytic continuation.

More recently, Tianwen-1, China's first Mars exploration mission, has provided new opportunities for magnetic field studies. The mission consists of an orbiter, a lander, and the Zhurong rover, enabling both orbital and surface investigations. The orbiter was in a highly eccentric orbit with an altitude ranging from about 270 km at periaresis to about 13,000 km at apoapsis (Liu K et al., 2020), while Zhurong has contributed in situ magnetic measurements at the surface. In preparation for the mission, Li XZ et al. (2020) constructed a local ESD model of the candidate landing area, an application of the ESD approach at regional scale. Together, these datasets expand the scope for refining crustal magnetic field models.

The primary source of interference in modeling the Martian crustal magnetic field stems from external magnetic fields. However, there has been limited analysis focusing on the impact of these external fields on crustal magnetic field modeling. Satellite observations have identified a dawn-to-dusk asymmetry in the magnetic tail of Mars, with a pronounced preference for the

dawn side (Halekas et al., 2006). This phenomenon has been supported by space magnetohydrodynamics (MHD) simulations of Mars (Luhmann et al., 2015). The influence of external fields on the Martian crustal magnetic field also varies with the intensity of the upstream IMF (Mittelholz et al., 2018). Accordingly, upstream IMF proxies from different orbital datasets can be employed to identify and select intervals corresponding to relatively quiet conditions, thereby minimizing contamination from external disturbances. Additionally, identifying regions on Mars with strong or weak remanent magnetism will significantly affect the positioning of the ionosphere. Given the complexity of these interactions, applying simplistic selection criteria to Martian crustal magnetic field data, as done in previous studies, is inadequate. Instead, it is essential to establish more refined and diverse selection criteria tailored to different spatial and environmental factors.

To address these issues, we construct a new model by incorporating novel methods, including a grid distribution based on the Fibonacci sphere distribution and a two-step inversion approach to enhance accuracy and stability. Our new ESD model employs the Mars magnetic field proxy to minimize the influence of external fields. In addition to integrating data from MGS and MAVEN, we also incorporated magnetic field measurements from Tianwen-1. To address the non-uniqueness problem inherent in the ESD method, we adopted a two-step inversion strategy, while Iteratively Reweighted Least Squares (IRLS) with Huber weighting was applied to mitigate external field effects and noise, and to prevent overfitting at higher orders. Furthermore, a temporal and spatial analysis of the misfit was conducted to refine criteria for magnetic field data selection, which may provide guidance for future studies. The remainder of this paper is organized as follows: Section 2 describes the datasets and selection criteria; Section 3 outlines the modeling and inversion methodologies; Section 4 presents the results and comparisons with existing models; and Section 5 concludes with a summary and discussion.

## 2. Data Sets and Data Selection

### 2.1 Data sets

The Aerobraking and Science Phase Orbit (AB/SPO) of MGS, which lasted from September 1997 to March 1999, was characterized by an eccentric orbit. During the AB/SPO phase, the MGS primarily acquired low-altitude measurements on the dayside, with periapsis altitudes ranging between 90 km and 169 km, which resulted in a limited temporal distribution of data. Subsequently, the Mapping Phase Orbit (MPO) commenced in March 1999 and continued until November 2006 (Albee et al., 2001). This phase featured a nearly circular orbit, during which a significant volume of magnetic field data was collected at altitudes between 368 km and 438 km, with an inclination of  $87^\circ$  and local times at either 2 a.m. or 2 p.m. The MPO phase provided an abundance of nightside observations at similar local times. These repeated and extensive measurements significantly contribute to the construction of a stable model at an altitude of approximately 400 km. The MGS magnetic field data were originally sampled at a frequency of 32 Hz (Acuña et al., 2001). To reduce computational costs, we resampled the AB/SPO and MPO data to one sample per minute.

For models constructed solely from MGS data, the reliability of downward continuation to the surface is constrained by the limited observation altitude. To address this limitation, it is essential to incorporate MAVEN datasets, which offer superior coverage at lower altitudes. The MAVEN

spacecraft was launched on November 18, 2013, and began its scientific operations in November 2014. It follows an eccentric orbit with a periapsis altitude of approximately 140 km, an inclination of  $74^\circ$ , and an orbital period of 4.5 hours. While MAVEN does not provide polar magnetic field observations, it offers extensive coverage in terms of altitude, latitude, longitude, and local time. To extend its low-altitude sampling, the periapsis is periodically lowered to 120 km during “deep-dip” campaigns. While MAVEN observations provide valuable low-altitude data, they lack the repetition found in MGS measurements and are potentially subject to significant noise interference. Therefore, integrating data from both MGS and MAVEN is crucial to mitigate these challenges and improve the overall robustness and reliability of the model. The MAVEN magnetic field data are sampled at a frequency of 32 Hz (Jakosky et al., 2015). For this study, we utilize MAVEN measurements up to December 31, 2022, and resample the raw data to one sample every 4 seconds to balance data resolution and computational efficiency.

Tianwen-1 is equipped with a Mars Magnetic Field Probe designed to study complex space phenomena on Mars, including ion escape and magnetic tail dynamics (Zou ZX et al., 2023). Its orbit is highly eccentric, similar to MAVEN’s, spanning from the bow shock to the Martian magnetic tail. Tianwen-1 began recording magnetic field data on November 13, 2021, primarily in the pile-up region (Wang YM et al., 2023). With an orbital period of 7.8 hours, most of its time is spent in the solar wind. Although the volume of night-side data at lower altitudes is limited, Tianwen-1, one of only two magnetometer-equipped orbiters still in operation, provides distinctive observations, particularly within the Martian magnetotail region. These measurements have significantly advanced our understanding of solar wind interactions and magnetic field dynamics (Guo ZZ et al., 2024). Integrating Tianwen-1 data into the modeling framework enables a practical evaluation of its utility and quality, offering insights into both its strengths and inherent limitations. Tianwen-1 is included as a complementary dataset in this study, but its practical contribution to resolving short-wavelength crustal signals is limited by the coverage of the retained data. The magnetic field data are sampled at 32 Hz when the orbiter is near the periareon (Wang GQ et al., 2024). For this study, we resampled the vector magnetic field data from Tianwen-1 to one sample per second.

## 2.2 Data selection

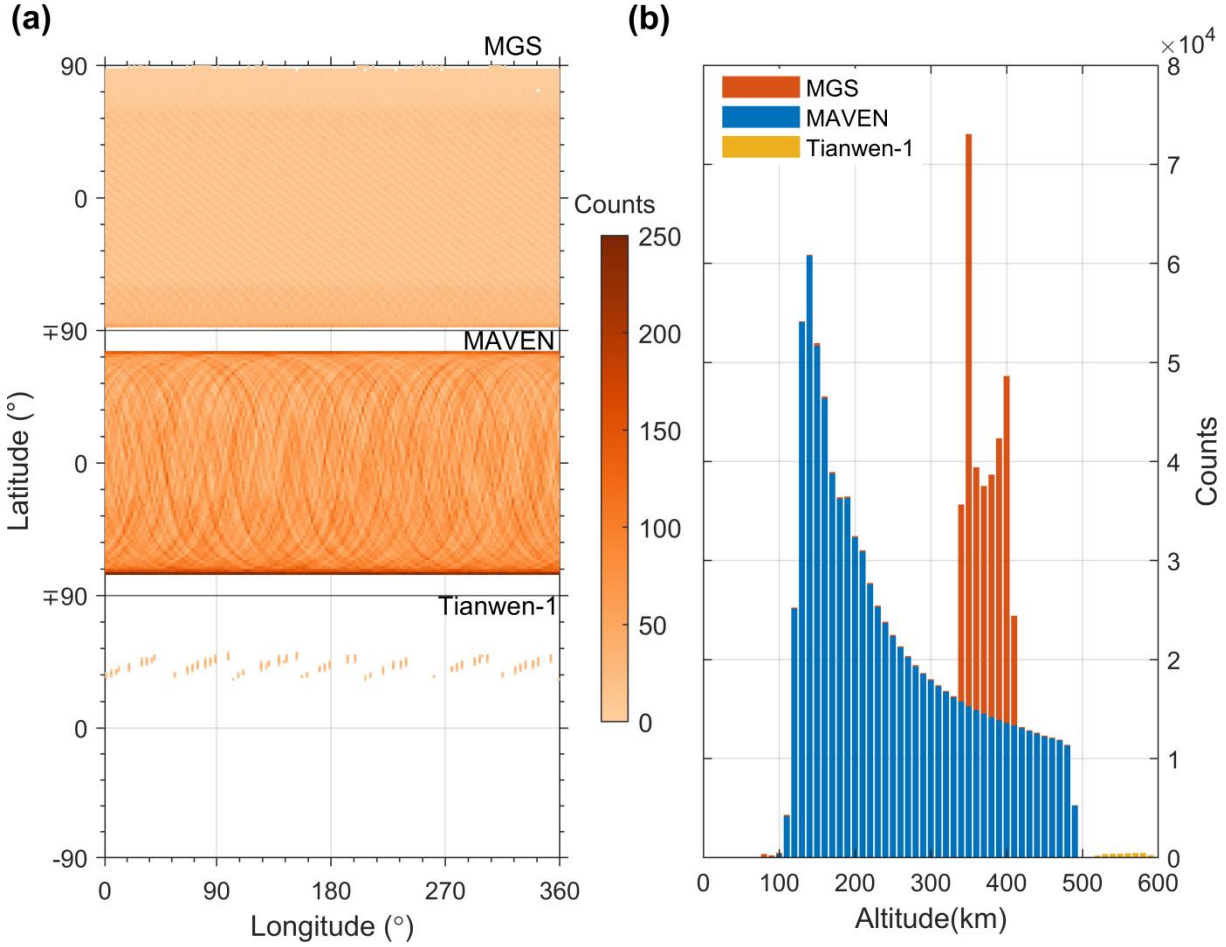
Since external magnetic sources are not explicitly modeled, it is crucial to apply specific selection criteria to identify data with minimal contributions from these external fields. Careful data selection not only improves the reliability of crustal magnetic field models but also significantly reduces computational costs when processing large datasets. We apply the following selection criteria, which are largely based on previous studies (Gao JW et al., 2021; Langlais et al., 2019).

1. At higher altitudes, the magnetic field is significantly influenced by the solar wind. Therefore, we select data collected below 500 km for MGS and MAVEN, and below 600 km for Tianwen-1.
2. In order to avoid the direct influence of large-scale ionospheric current and solar wind, we only select the night-side data with an asymmetrical edge, which refers to a longitude range in the Mars Solar Orbital (MSO) coordinate system that is not symmetrically distributed, because the interaction of the solar wind and bow shock is asymmetrical (Fang XH et al., 2023). The range of the nightside is defined as longitudes  $\phi$  in MSO

coordinate systems between  $105^\circ$  and  $260^\circ$ . For Tianwen-1, the nightside range is defined as  $\phi$  in MSO between  $90^\circ$  and  $270^\circ$ . This enables the procurement of a modest dataset from Tianwen-1 without being significantly impacted by external fields.

3. The intensity of the upstream IMF is strongly correlated to the noise levels (Gao JW et al., 2024; Mittelholz et al., 2018). When the upstream IMF intensity is high, the misfit between model predictions and observations increases significantly. For a given orbit, the influence of space weather can be assessed by examining the upstream IMF data as the spacecraft crosses the bow shock. Therefore, for MAVEN data, we exclude data from orbits where the upstream IMF intensity exceeds a predefined threshold 2.6 nT, following conclusions of Gao JW et al. (2021). In addition, orbits without available IMF intensity measurements are also removed from the modeling. The bow shock boundary is defined according to Vignes et al. (2000), and we extend this range by 0.3 Martian radii (rm) to ensure robustness. Additionally, we use the average IMF intensity as a proxy only when the spacecraft spends more than 30 minutes in the upstream solar wind for one orbit. Using this selection criterion, a total of 2,829 orbits were retained for further analysis. For Tianwen-1 data, although its orbit is similarly eccentric to MAVEN's, it rarely crosses the bow shock. As a result, we cannot apply the IMF-based magnetic field proxy for data selection in this case.
4. Only for the MPO dataset, we select data from periods of minimum solar activity to significantly reduce computational costs. In this study, we focus on data from the year 2006, as it aligns most closely with the solar minimum that occurred in 2008.

By applying the aforementioned criteria, we selected a total of 869,868 MAVEN data points, 25,206 AB/SPO data points, 203,832 MPO data points, and 3,060 Tianwen-1 data points for crustal magnetic field modeling. Figure 1 illustrates the data distribution and coverage for MAVEN, MGS, and Tianwen-1. The altitude of MGS varies nearly uniformly with longitude, providing good global coverage at the altitude around 400 km. MAVEN offers superior longitude and altitude coverage due to its orbital precession, although it lacks data at high latitudes. Due to its orbital design, Tianwen-1's low-altitude data is predominantly concentrated on the dayside, resulting in limited spatial coverage and a small dataset around  $45^\circ\text{N}$  latitude, while the altitude coverage is inadequate to provide small-wavelength information.



**Figure 1.** The spatial distribution of the selected data sets after selection. (a) The latitude-longitude distributions for MGS, MAVEN and Tianwen-1. The bin is set as  $2^\circ$  longitude  $\times$   $1^\circ$  latitude. (b) The histogram of the altitude for different data sets. The bin is set as 10 km for the altitude.

### 3. Modeling Scheme

Our model follows and extends the modeling scheme by Purucker et al. (2000) and Langlais et al. (2019) based on the ESD method, incorporating a novel dipole mesh distribution and an improved inversion scheme. The ESD method has been extensively utilized in lithospheric magnetic field modeling (Kother et al., 2015; Mayhew et al., 1978; Purucker, 2008). The ESD method is particularly advantageous for its sensitivity to non-uniform magnetic fields, making it well-suited for the heterogeneous magnetic conditions observed on Mars. Unlike the SH method, the ESD approach does not require formal regularization (Wahler and Gubbins, 1981), thereby simplifying the modeling process while maintaining accuracy.

#### 3.1 Equivalent Source Dipole Method

In the ESD Method, the magnetic field at an observation point is modeled as the superposition of magnetic fields generated by a large number of magnetic dipoles. The magnetic potential at any source-free location  $(r, \theta, \phi)$  due to these dipoles located at  $(r_d, \theta_d, \phi_d)$  can be expressed as follows:

$$V = \sum_{i=1}^{N_{\text{dip}}} \left( -\mathbf{M}_i \cdot \nabla \frac{1}{l_i} \right), \quad (1)$$

where  $N_{\text{dip}}$  represents the total number of dipoles,  $\mathbf{M}_i$  represents the magnetization of the  $i$ th magnetic dipole and  $l_i$  is the distance between the observation point and the  $i$ th magnetic dipole:

$$l_i = [r_d^2 + r^2 - 2r_d r (\cos(\theta)\cos(\theta_d) + \sin(\theta)\sin(\theta_d)\cos(\phi - \phi_d))]^{\frac{1}{2}}, \quad (2)$$

in the planetocentric spherical coordinate system. The vector magnetic field  $\mathbf{B}$  can be represented as the gradient of the scalar magnetic potential  $V$ :

$$\mathbf{B} = -\nabla V. \quad (3)$$

It is evident that when  $l_i$  exceeds a certain threshold, the influence of the magnetic dipole on the observation point becomes negligible. Therefore, in this study, we select a maximum threshold of 1,700 km. This choice reduces computational demands at negligible cost in accuracy. Evaluated at the recovered model, the field neglected by the threshold has an rms of 0.1–0.7 nT per component at observation altitudes, an order of magnitude below the observational uncertainties (Table 1). The total is this small because the far-field contributions largely cancel. The inversion problem can be formulated as:

$$\mathbf{d} = \mathbf{G}\mathbf{m} + \mathbf{v}, \quad (4)$$

where  $\mathbf{d}$  represents the magnetic field observation vector,  $\mathbf{m}$  is the vector of model parameters, namely the magnetization of the magnetic dipoles, and  $\mathbf{G}$  is the model matrix, which is dependent on the position of observation points and dipoles. The term  $\mathbf{v}$  represents the noise vector of observation. Assuming that the noise follows a multivariate Gaussian distribution of zero mean, the inversion problem can be reformulated as the minimization of the noise amount

$$w_i = \begin{cases} 1 & |\mathbf{v}_i| < \delta, \\ \delta/|\mathbf{v}_i| & |\mathbf{v}_i| > \delta, \end{cases} \quad (5)$$

This formulation corresponds to the least squares solution for  $\mathbf{m}$ , which can be expressed as:

$$\mathbf{m} = (\mathbf{G}^T \mathbf{G})^{-1} \mathbf{G}^T \mathbf{d}. \quad (6)$$

We solve this inversion problem using the conjugate gradient method, which provides an efficient and robust approach to obtaining the optimal model parameters.

### 3.2 Source Distribution

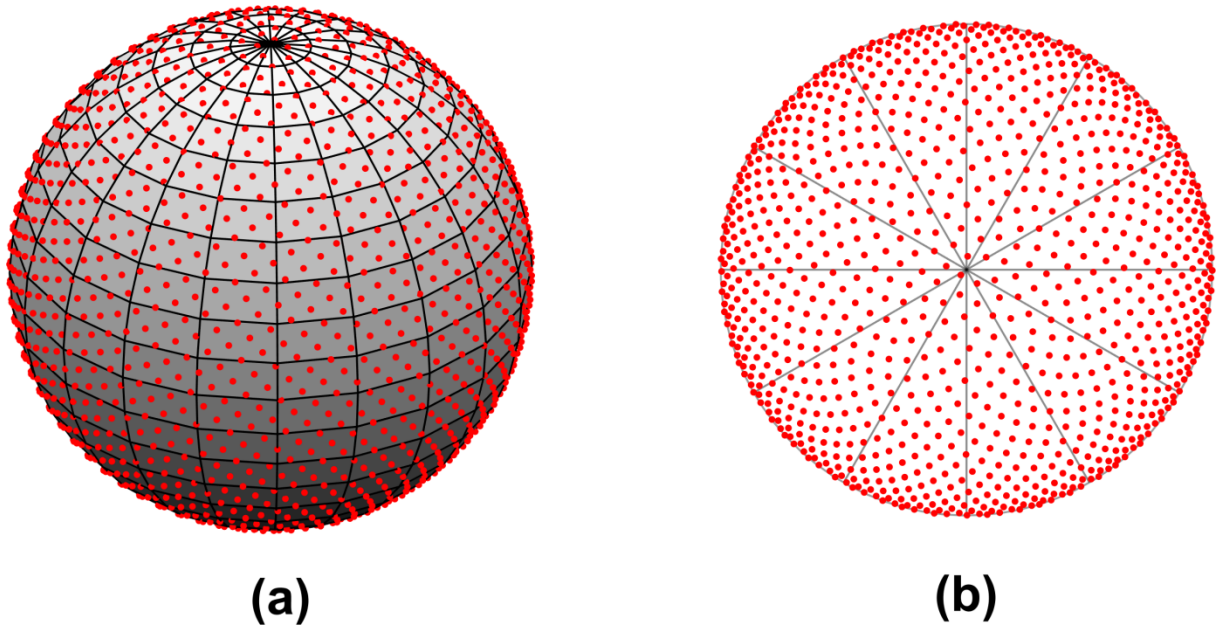
The ESD method necessitates that the dipoles achieve maximum uniform distribution across the sphere. Previous studies have employed geometrically uniform spherical meshes for this purpose (Kother et al., 2015; Oliveira et al., 2015). In this study, considering the theoretical implications of Runcorn's theorem, which states that if the magnetization of a spherical shell is entirely induced by its internal field, then any ideal dipole distribution on the shell's surface should produce no external magnetic field (Runcorn, 1975). By comparing the average radial magnetic field intensity at 150 km altitude produced by dipole sources of uniform susceptibility magnetized by a central axial dipole (Langlais et al., 2019), we employed the Fibonacci sphere distribution shown as Figure 2, which provides a geometrically uniform grid on the spherical surface. On a sphere of radius 1, the grid in Cartesian coordinate system is represented as follows

$$x = (\sqrt{1 - z^2})\cos((\sqrt{5} - 1)(n_{\text{dip}} + 1)\pi), \quad (7)$$

$$y = (\sqrt{1 - z^2})\sin((\sqrt{5} - 1)(n_{\text{dip}} + 1)\pi), \quad (8)$$

$$z = (2 * n_{\text{dip}} + 1)/N_{\text{dip}} - 1. \quad (9)$$

$N_{\text{dip}}$  is the number of grid points,  $n_{\text{dip}}$  from 0 to  $N_{\text{dip}} - 1$ . We used 15,000 dipoles at a depth of 20 km following Langlais et al. (2004) and the reference radius  $a$  of Mars is set to 3,393.5 km. The magnetic dipole position is projected into the spherical coordinate system before calculating the model matrix  $\mathbf{G}$ .



**Figure 2.** Illustration of the Fibonacci sphere distribution of dipoles. (a) 3D distribution of dipoles on a spherical surface. (b) Polar distribution.

### 3.3 Inversion Method

The ESD method is inherently non-unique due to the fundamental limitation that external magnetic field observations cannot uniquely constrain the underlying magnetization distribution. This non-uniqueness arises particularly in cases where the magnetization follows specific distributions, such as uniform magnetization or toroidal magnetization within a uniform spherical shell known as magnetic annihilators (Gubbins et al., 2011). Purucker et al. (2000) inverted the Martian crustal magnetic field using the ESD method, but restricted their analysis to the radial magnetic field component to recover radial magnetization. This approach was motivated by the observation that external fields dominated by radial magnetization tend to minimize the required magnetization intensity. In contrast, horizontal magnetic components are more susceptible to disturbances from external fields, making them less reliable for inversion (Langlais et al., 2004; Morschhauser et al., 2014). More recently, Langlais et al. (2019) employed a combined vector and scalar inversion approach, incorporating a large dataset with an improved selection scheme. To obtain a three-component magnetization distribution, we adopt a two-step inversion strategy. In the first step the data are restricted to the radial component  $B_r$  and the model to the radial dipole moments  $M_r$ , and the normal equations are solved by conjugate-gradient iteration from a zero starting model. The restriction on the data follows from the sensitivity of the horizontal components to external fields noted above; the restriction on the model rests on potential theory. A purely radial source layer reproduces every external field that a general three-component layer can produce, and reproduces it uniquely. Fixing  $M_r$  first therefore discards no observable information. It selects one well-defined representative of the infinite family of magnetizations that fit the data equally well, the family generated by the annihilators (Appendix A). In the second step all three moment components and all three field components are used, starting from the first-step solution with the horizontal moments  $M_\theta$  and  $M_\phi$  initialized to zero and proceeding by iteratively reweighted least squares (IRLS) with the Huber weights described below. Because the conjugate-gradient iterates differ from the starting model only within the row space of the weighted operator, the representative selected in the first step is preserved through the second (Appendix A). The sensitivity of conjugate-gradient solutions to the initial model is a consequence of this non-uniqueness; the two-step scheme uses that sensitivity to fix the representative rather than leaving it to an arbitrary initialization. The misfit of the horizontal components falls substantially during the first step even though those components are not fitted there. Only internal potential-field structure links the three components in this way, so this reduction indicates that the first step recovers crustal signal rather than external contamination.

Throughout the three-component inversion process, the overall misfit of the radial magnetic field component continues to decline, accompanied by a reduction in the misfit of the horizontal components. This suggests that the inclusion of horizontal magnetic field does not lead to increased uncertainty in the inversion results. By implementing this approach, we aim to enhance the precision and reliability of the inversion results, thereby improving the robustness of our overall analysis.

We use IRLS during the three-component inversion to estimate model parameters while downweighting the influence of outliers with Huber weights, which applies squared misfit for small residuals and absolute misfit for large residuals, reducing overfitting at small scales and increasing robustness (Huber, 1964). In the IRLS framework, the amount of noise  $L$  is redefined as:

$$\mathbf{L}(\mathbf{m}) = \mathbf{v}^T \mathbf{W} \mathbf{v}, \quad (10)$$

where  $\mathbf{W}$  is a diagonal weight matrix using the standard Huber weight function. The corresponding solution for  $\mathbf{m}$  is given by:

$$\mathbf{m} = (\mathbf{G}^T \mathbf{W} \mathbf{G})^{-1} \mathbf{G}^T \mathbf{W} \mathbf{d}. \quad (11)$$

The  $i$ th diagonal element  $w_i$  of the weight matrix  $\mathbf{W}$  is determined by the Huber weight:

$$w_i = \begin{cases} 1 & |\mathbf{v}_i| < \delta, \\ \delta/|\mathbf{v}_i| & |\mathbf{v}_i| > \delta, \end{cases} \quad (12)$$

where  $\delta$  is a prior threshold misfit value that remains fixed throughout the inversion, and  $\mathbf{v}_i$  for each sample point is obtained from the residuals of the current iteration. We set the threshold  $\delta$  to 1.32 times the standard deviation  $\sigma$  of the misfit, corresponding to approximately 90% of the data in a Gaussian distribution (Devore, 2000). To apply the Huber weight in the inversion process, it is necessary to prescribe  $\sigma$  based on prior knowledge of the error distribution. Here we assigned prior parameters of  $\sigma = 6$  nT for MGS data,  $\sigma = 7$  nT for MAVEN data, and  $\sigma = 10$  nT for Tianwen-1 orbiter data based on the misfit values from our initial unweighted inversion. When the misfit values of specific sample points exceed the threshold  $\delta$ , their contribution to the noise term  $\mathbf{L}$  transitions from a quadratic to a linear function. This transition is based on the assumption that the misfit distribution follows a Gaussian distribution within the interval  $[-\delta, \delta]$ , while values outside this interval conform to a Laplacian distribution. Consequently, the fitting curve primarily aligns with the majority of the sample points while still accounting for the presence of outliers. In cases where outliers are significant and must be incorporated into the modeling process, the Huber weight provides robustness against outliers while preserving sensitivity to small residuals.

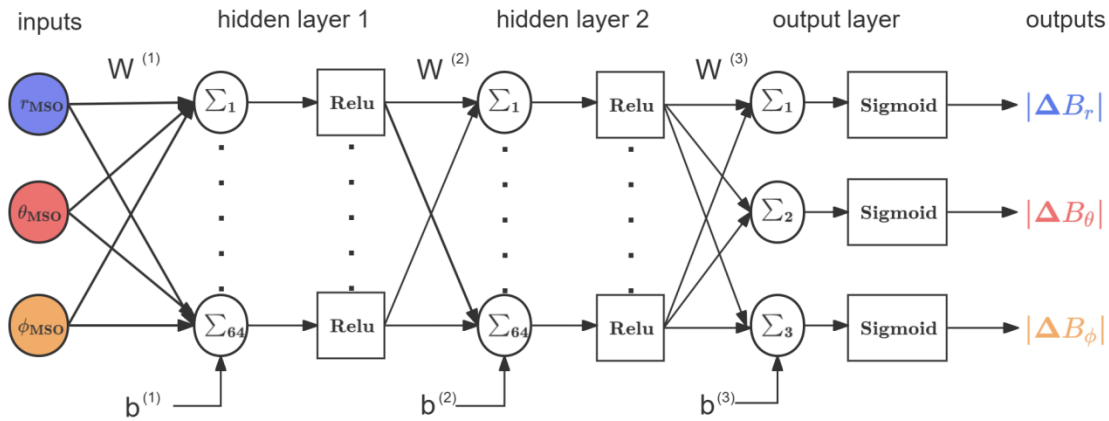
### 3.4 Residual Modeling

The residual, defined as the difference between the observed and predicted magnetic fields ( $\Delta \mathbf{B} = \mathbf{B}_{\text{obs}} - \mathbf{B}_{\text{pre}}$ ), quantifies the discrepancy between the model predictions and actual measurements. In this study the residual is given by the observational vector minus our ESD model prediction ( $\Delta \mathbf{B} = \mathbf{d} - \mathbf{G}\mathbf{m}$ ). In the modeling described earlier in this study, only the internal field was represented, while external field contributions were entirely neglected. As a result, the residuals essentially represent an estimate of the external field, obtained by subtracting the large-scale lithospheric magnetic contribution from the total observed field. For Martian crustal magnetic field modeling, the dominant source of residuals originates from these external field influences. Consequently, explicitly modeling external fields could significantly improve data screening strategies and enable more refined crustal field models. It may exhibit correlations with variables such as time and position, providing insights into the underlying physical processes. This residual analysis is an auxiliary diagnostic. It characterizes the external-field contamination present in the data and is not used in constructing the ESD model presented above. Due to the significant variations in spacecraft altitude and the complexity of external field

sources, the spatial coverage of the data is incomplete, making direct analysis and visualization challenging. Therefore, we employ an artificial neural network (ANN) using PyTorch (Paszke et al., 2019) to model the residuals, which are derived from the MAVEN dataset. The choice of MAVEN data is motivated by its comprehensive coverage in local time and the availability of IMF proxies. This analysis aims to investigate the interrelationship between the residual and other variables, providing guidance for subsequent data selection or modeling methods.

For residual modeling, we use MAVEN magnetic field data resampled to one sample every 4 seconds, up to December 31, 2022. The data are filtered to exclude observations at altitudes above 500 km, with the additional selection being the orbital criteria under different IMF conditions described in Section 2. Using our ESD model, we calculate the residuals for all resampled MAVEN data under both high ( $|B_{\text{IMF}}| > 2.6$  nT) and low IMF ( $|B_{\text{IMF}}| < 2.6$  nT) conditions. The residuals are then normalized by dividing by  $3\sigma = 21$  nT, a value chosen based on the standard deviation of the residuals in the training set. This normalization ensures that the residuals are scaled appropriately for model training. To focus on the magnitude of the residuals rather than their sign, we take their absolute values. To limit the impact of extreme values, any normalized residual exceeding 1 is truncated to 1. These processed residuals are used as the target outputs for training an ANN for further analysis.

We implement a fully connected feedforward ANN with two hidden layers (64 neurons each) as shown in Figure 3, where each neuron computes a weighted sum followed by a nonlinear activation. The input layer receives normalized MSO coordinates ( $r_{\text{MSO}}$ ,  $\theta_{\text{MSO}}$ ,  $\phi_{\text{MSO}}$ ), followed by hidden layers using Rectified Linear Unit (ReLU) activation, defined as  $\text{ReLU}(x) = \max(0, x)$ . The pre-activation  $x = \sum h \cdot W + b$ , where  $h$  are the input values,  $W$  are the corresponding weights and  $b$  are the bias values. The output layer employs a Sigmoid activation  $\sigma(x) = 1/(1 + e^{-x})$  to predict the absolute values of the residuals for magnetic field components ( $|\Delta B_r|$ ,  $|\Delta B_\theta|$ ,  $|\Delta B_\phi|$ ). This design allows the model to directly estimate the magnitude of the residuals, which is the primary focus of our analysis.



**Figure 3.** The architecture of the ANN used for inferring misfit using spacecraft location. The ANN has 3 inputs for spacecraft location and 3 outputs for residuals. The ANN consists of two hidden layers consisting of 64 neurons in both first and second hidden layers. The neurons in the hidden layers have ReLU nonlinear transfer functions, and the output layer has Sigmoid nonlinear transfer functions.

The model is trained using the Adam optimizer which adaptively adjusts parameter update step sizes (Kingma and Ba, 2014), with a learning rate of 0.001; training minimizes the mean squared error (MSE) between predicted and target values. To prevent overfitting, we apply early stopping with a patience of 10 epochs, using a validation set comprising 10% of the data to monitor the model's performance during training, and an independent test set comprising 10% of the data is held out for final evaluation. Training is performed separately on low and high IMF subsets to assess their distinct misfit patterns. The ANN is chosen for residual analysis due to its strong fitting capability, especially when the underlying physical processes are not fully understood. The ANN effectively identifies spatial and temporal patterns in the data; after training it enables computationally efficient forward evaluations for rapid screening and localization of times and regions with strong external noise. By augmenting or modifying the network inputs one can implement more comprehensive data selection criteria and thereby improve recovery of small scale crustal signals.

## 4. Results

### 4.1 Model Statistics

The final model undergoes 20 iterations in the radial inversion and 10 iterations in the vector inversion. The final residuals  $\mathbf{v}$  for different components across various datasets are summarized in Table 1. The formula for average misfit is as  $\mu = \sum(\mathbf{d} - \mathbf{Gm})/N_{\text{obs}}$ . Assuming this misfit is of zero mean, the standard deviation of the misfit  $\sigma$  which is also rms misfit can be calculated as

$$\sigma = \sqrt{\frac{(\mathbf{d} - \mathbf{Gm})^T (\mathbf{d} - \mathbf{Gm})}{N_{\text{obs}}}}, \quad (13)$$

where  $N_{\text{obs}}$  is the number of measured components. And a similar rms value  $\sigma_m$  associated with the magnetization distribution is defined as:

$$\sigma_m = \sqrt{\frac{\mathbf{m}^T \mathbf{m}}{N_{\text{dip}}}}. \quad (14)$$

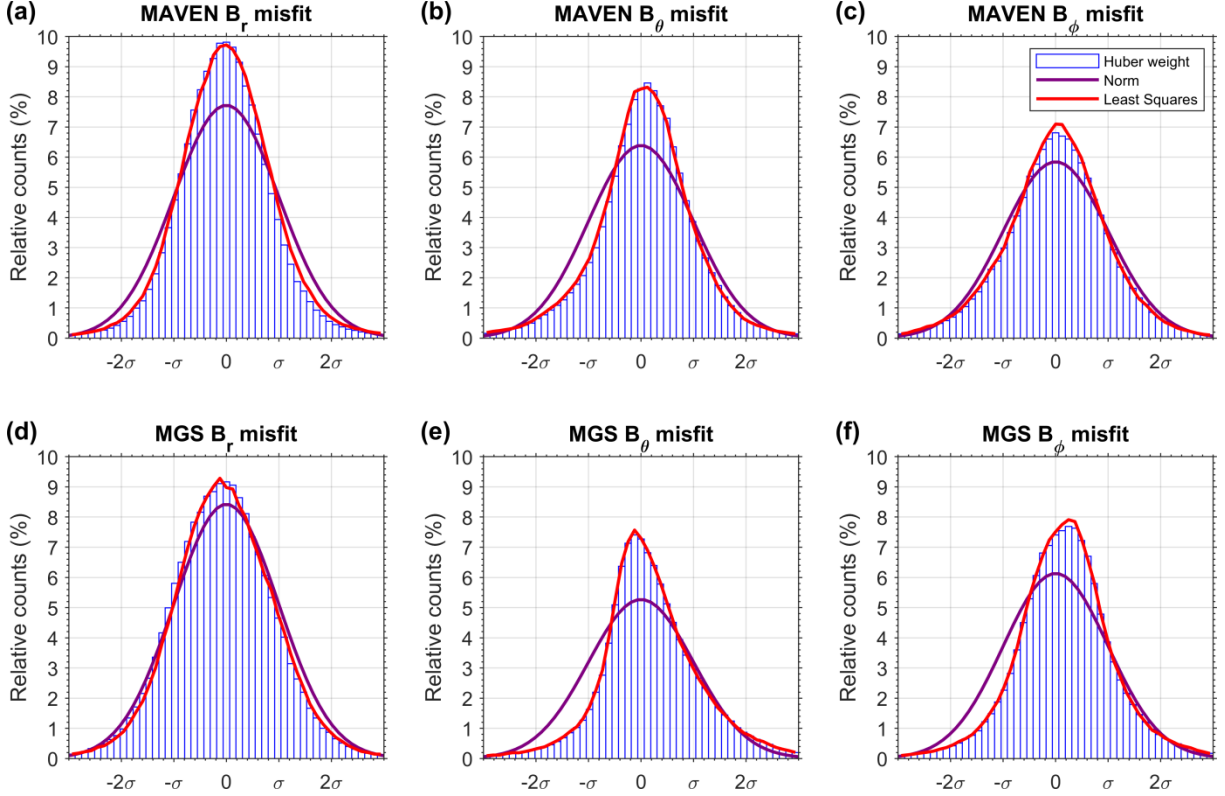
From the residual statistics, it is evident that the residuals in the radial component are smaller compared to the horizontal components. This is consistent with our inversion strategy, which prioritizes the radial component due to its lower susceptibility to disturbances from external sources such as the solar wind. Additionally, the MGS dataset, which was collected at a relatively fixed and high orbital altitude, exhibits smaller misfits. In contrast, the Tianwen-1 dataset, despite having a similar orbit to MAVEN, exhibits larger misfits. This discrepancy is likely due to the data being predominantly collected near the dayside, where ionospheric effects are more pronounced. Two further factors contribute to this discrepancy. No upstream IMF proxy is available for Tianwen-1 (Section 2), so disturbed intervals cannot be identified and removed before the inversion, and the 3,060 retained observations include no repeated passes over which the time-varying external contributions would average toward zero. The pattern of

the residuals is consistent with an external origin: the excess misfit is confined to the horizontal components, while the Tianwen-1 radial component has the smallest rms misfit of any dataset (Table 1). In the case of the MAVEN dataset, which consists of low-altitude observations with higher spatial resolution, our model achieves misfits comparable to those obtained for the MGS dataset. The performance on the MAVEN dataset may be attributed to the specific inversion scheme and data selection criteria used in our approach.

**Table 1.** *Residual Statistics of Mean and rms Misfits for Different Data Sets and Magnetic Field Components  $B_r$  (Radial),  $B_\theta$  (Horizontal South),  $B_\phi$  (Horizontal East).*

| Data set  | rms $\sigma$ $B_r$<br>(nT) | rms $\sigma$ $B_\theta$<br>(nT) | rms $\sigma$ $B_\phi$<br>(nT) | Mean $\mu$ $B_r$<br>(nT) | Mean $\mu$ $B_\theta$<br>(nT) | Mean $\mu$ $B_\phi$<br>(nT) |
|-----------|----------------------------|---------------------------------|-------------------------------|--------------------------|-------------------------------|-----------------------------|
| AB/SPO    | 6.55                       | 9.72                            | 9.06                          | 1.58                     | 0.45                          | -1.32                       |
| MPO       | 4.44                       | 7.22                            | 6.11                          | 0.18                     | -1.75                         | -1.12                       |
| MAVEN     | 5.17                       | 6.24                            | 6.82                          | 0.10                     | -0.52                         | -0.11                       |
| Tianwen-1 | 4.06                       | 11.74                           | 14.29                         | 1.67                     | -2.77                         | -8.94                       |

To further validate our approach, we also performed 10 iterations of the vector inversion using the Least Squares method and compared its residual distribution with a normal distribution. The residual histograms for different datasets and components are shown in Figure 4. Most of the residuals fall within the interval  $[-\sigma, \sigma]$  with the radial component  $B_r$  exhibiting the most concentrated distribution. However, the residuals from the Least Squares method do not follow a normal distribution but instead exhibit a long-tail distribution. This behavior underscores the necessity of incorporating Huber weighting to mitigate the influence of outliers and improve the robustness of the model. At the same time, assuming Gaussian residuals is a reasonable and practical approximation because the L2 formulation is computationally efficient and with dense data coverage a small number of large residuals typically do not materially affect the large-scale magnetic field.



**Figure 4.** Histograms of the residuals for MAVEN (a-c) and MGS (d-f) with  $B_r$ ,  $B_\theta$  and  $B_\phi$  component respectively. The residual histograms of the Huber weight model (blue cubes), the Least Squares model (red lines) and Normal distribution with the same standard deviation  $\sigma$  as our Least Squares model (purple lines) are overlapped for reference. Misfits are normalized with respect to individual  $\sigma$ .

#### 4.2 Transformation into SH Model

The ESD method provides an equivalent representation of physical fields, which can also be expressed in alternative mathematical formulations and translated into other representations. For instance, the ESD model can be converted into a SH model. To achieve this, we first project the ESD model onto the physical field and compute the magnetic field at 100,000 points distributed according to the Fibonacci Sphere Distribution at an altitude of 150 km, the grid resolution is defined to ensure that the model can be reliably estimated without aliasing effects up to at least spherical harmonic degree 250 (Langlais et al., 2019). The magnetic potential in the SH method is

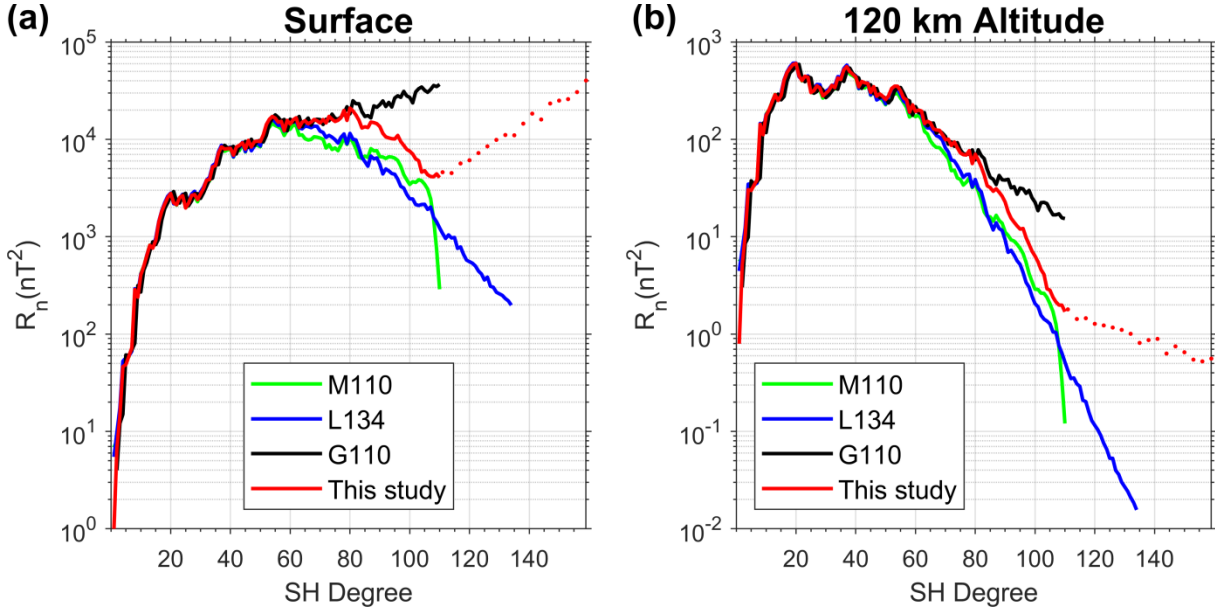
$$V = a \sum_{n=1}^N \left(\frac{a}{r}\right)^{n+1} \sum_{m=0}^n (g_n^m \cos m\phi + h_n^m \sin m\phi) P_n^m(\cos\theta), \quad (15)$$

where  $N$  is the maximum SH degree,  $P_n^m$  represents the Schmidt semi-normalized associated Legendre function for degree  $n$  and order  $m$ . The Gaussian coefficients  $(g_n^m, h_n^m)$  of SH model up to degree 160 are then derived using least squares without regularization or iterative refinement.

Once the Gaussian coefficients are obtained, we can calculate the Mauersberger-Lowes spatial power spectrum of the model (Lowes, 1966; Mauersberger, 1956). The power spectrum of the model at degree  $n$  is expressed as

$$R_n(r) = (n+1) \left(\frac{a}{r}\right)^{2n+4} \sum_{m=0}^n ((g_n^m)^2 + (h_n^m)^2), \quad (16)$$

where  $r$  is the radius above the reference radius  $a$ . In Figure 5, we compare the power spectra of our model with those from Gao JW et al. (2021), Langlais et al. (2019), and Morschhauser et al. (2014). The power spectra are evaluated at both the Martian surface and an altitude of 120 km.



**Figure 5.** Comparison of the power spectra of our model with previous models (M110 from Morschhauser et al. (2014), L134 from Langlais et al. (2019), G110 from Gao JW et al. (2021)) at the surface (a) and at the altitude of 120 km (b). Dashed lines of our model represent the magnetic power beyond the spherical harmonic degree of 110.

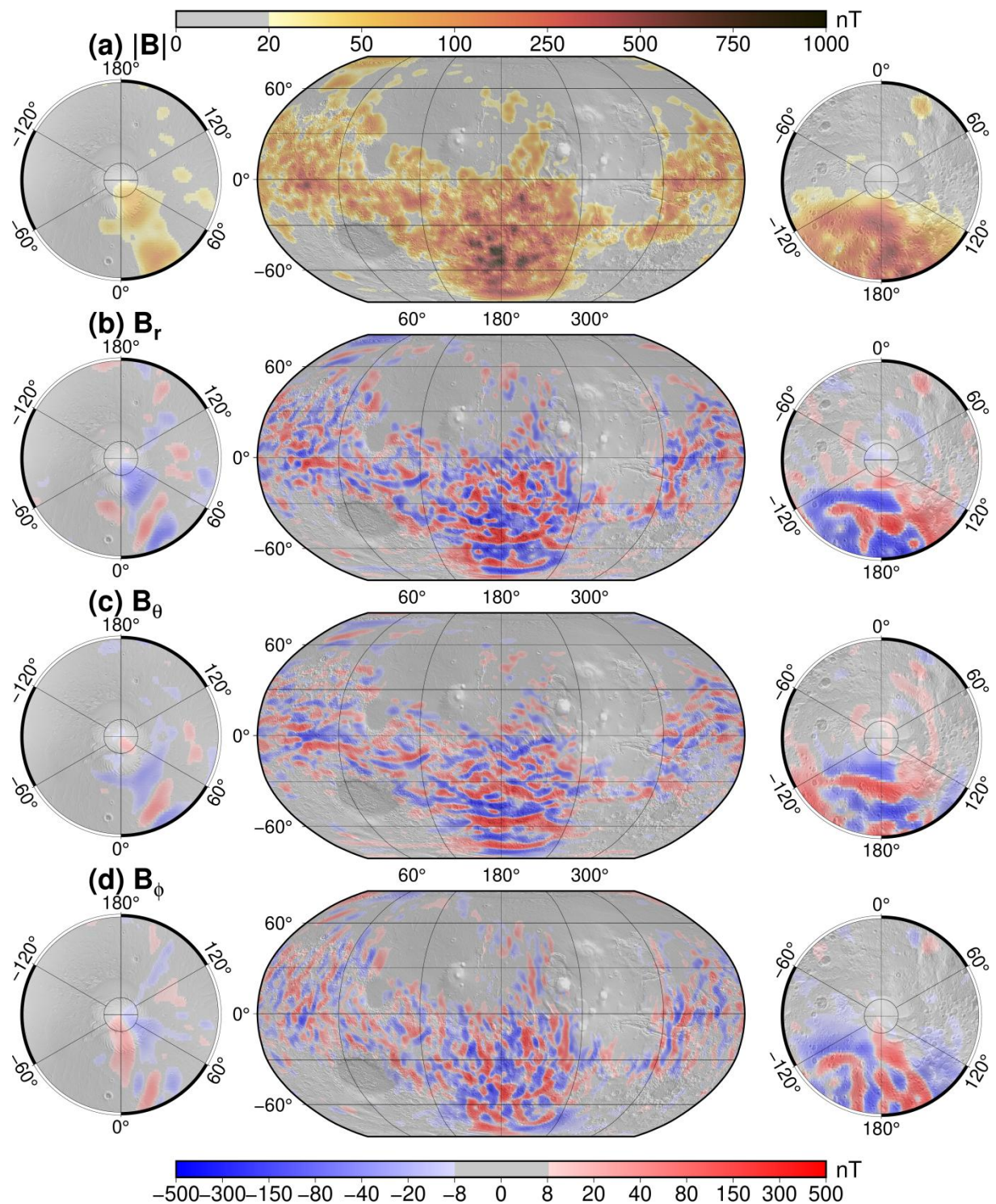
At degree 110, the energy of our SH model at the Martian surface begins to diverge, while its spectrum slope flattens at an altitude of 120 km. This behavior corresponds to a maximum resolution of our ESD model, leading us to truncate the model at  $n = 110$ . The primary differences between our model and others become apparent for degrees  $n > 70$ , which can be attributed to uncertainties in small-scale magnetic features. At higher degrees ( $n > 80$ ), the spectral energy of our model exceeds that of L134 and M110, approaching the levels of G110, before gradually declining. Gao JW et al. (2021) truncate their spherical harmonic model, built from the same MGS and MAVEN observations, at the same degree, on the different ground of a trade-off between model roughness and misfit. The agreement suggests that the limit is set by the observations rather than by either method.

The M110 model, which utilized all available vector measurements from MGS and applied regularization through an L1 norm of horizontal gradients, exhibits a sharp drop in spectral

energy for  $n > 106$ . In contrast, G110 did not employ regularization, resulting in a lack of convergence in its power spectrum at high degrees. These comparisons highlight the advantages of our approach in balancing resolution and robustness, particularly in capturing small-scale magnetic anomalies.

#### 4.3 Magnetic Field Map

We used our SH model to generate maps of the three components of the magnetic field at an altitude of 120 km on Mars, as illustrated in Figure 6. The large-scale magnetic features are consistent with those observed in previously published models. Notably, the strongest magnetic anomaly is located in the southern hemisphere near  $180^\circ$  longitude, while the northern hemisphere exhibits significantly weaker magnetic anomalies, indicating a pronounced hemispheric asymmetry. Additionally, the  $B_\theta$  displays distinct banding characteristics. Connerney et al. (1999) have proposed that the observed polarity alternations may be attributed to seafloor spreading processes, with repeated dike intrusion suggested as an alternative mechanism (Nimmo, 2000); Bouley et al. (2020) instead interpret the area as a collection of crustal blocks that may have formed by different processes.



**Figure 6.** Magnetic field maps of the spherical harmonic model up to  $n, m = 110$  at an altitude of 120 km for (a)  $|B|$ , (b)  $B_r$ , (c)  $B_\theta$  and (d)  $B_\phi$  from top to bottom. The left column shows polar magnetic field map between  $60^\circ\text{N}$  and  $90^\circ\text{N}$ . The right column shows polar magnetic field map between  $60^\circ\text{S}$  and  $90^\circ\text{S}$ . The middle panel displays the global magnetic field map using the Robinson projection. The black lines indicate the zero-elevation contour of the MOLA topography.

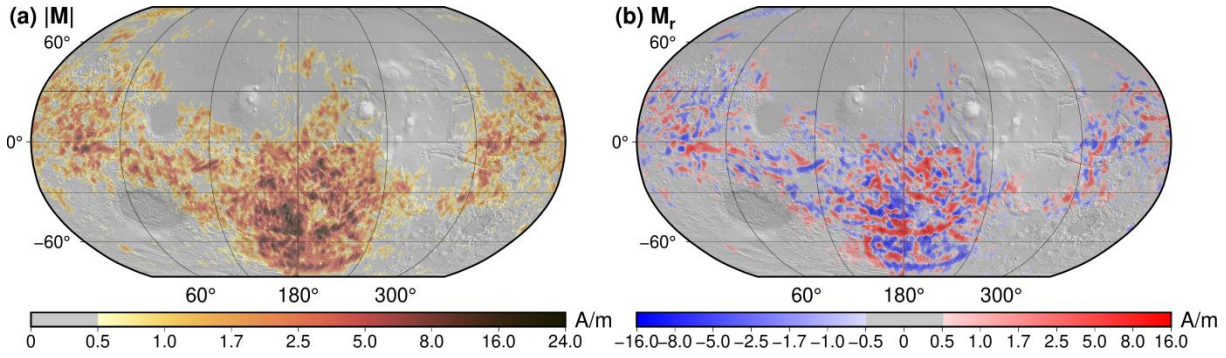
We further employed our model to map the magnetic field distribution within the polar regions, specifically between 60° and 90° latitudes in both the northern and southern hemispheres, as shown in Figure 6. The magnetic field intensity at the South Pole is approximately an order of magnitude greater than that at the North Pole. This asymmetry may be attributed to processes such as impact demagnetization or resurfacing events in the northern hemisphere. Previous research suggested that this asymmetry could be the result of a hemispherical dynamo (Landeau and Aubert, 2011). However, such a dynamo would require a high Rayleigh number, indicative of vigorous thermal convection, which is inconsistent with the conditions expected during the cessation of Martian dynamo activity. An alternative explanation posits that the hemispherical dynamo was driven by asymmetric heat flux at the core-mantle boundary (CMB) between the northern and southern hemispheres (Stanley et al., 2008). Nevertheless, this type of hemispherical dynamo would frequently reverse and fail to produce the remanent magnetization observed in the southern hemisphere (Yan C et al., 2023).

Large impact basins, such as Utopia, Argyre, and Isidis, which formed after the cessation of the Martian dynamo, exhibit weak magnetic anomalies. These anomalies may provide valuable constraints on the geological timing of their formation, although recent work suggests that weak basin magnetism could instead reflect cooling in a reversing dynamo (Steele et al., 2024). Volcanic provinces, such as Elysium and Tharsis, also display weak magnetic anomalies. These are typically associated with the demagnetization induced by volcanic activity.

#### 4.4 Magnetization Map

The ESD method employed in this study directly parameterizes the magnetic field in terms of the magnetic moments and spatial positions of individual dipoles. This representation can be converted into a magnetization distribution under the assumption of a uniformly magnetized layer with a predefined thickness of 40 km. While this conversion provides an approximate magnetization map, it is crucial to emphasize that this does not represent the true magnetization structure of the Martian crust. However, it serves as a valuable proxy for analyzing spatial trends and directional variations in magnetization.

As summarized in Table 2, the average total magnetization magnitude is 1.13 A/m, and the rms value is 2.41 A/m. The radial magnetization component  $M_r$  dominates the distribution, displaying a mean value of 0.005 A/m and an rms value of 2.41 A/m. In addition, the mean magnetization in the southern hemisphere is 1.65 A/m, which is significantly larger than the 0.65 A/m observed in the northern hemisphere, indicating a clear hemispheric asymmetry. Figure 7 illustrates the spatial distribution of the radial component and total intensity of the magnetization using interpolation. Magnetization exhibits very similar characteristics to magnetic fields, especially in the radial direction. Notably, both the average and rms values of the total magnetization in the southern hemisphere are significantly higher than those in the northern hemisphere, revealing a distinct hemispheric dichotomy with stronger magnetization in the south. In addition, at longitudes 60° to 300° and latitudes 60°S to 75°S, our magnetization does not show the weak magnetization pattern that produces no visible magnetic field reported in Langlais et al. (2019). This distinction suggests that our inversion strategy effectively mitigates annihilators inherent to the ESD method.



**Figure 7.** The equivalent magnetization intensity map of the ESD model, interpolated onto a  $0.4^\circ \times 0.4^\circ$  grid. (a) Total magnetization intensity  $|M|$ . (b) Radial magnetization component  $M_r$ . Both maps use the Robinson projection.

**Table 2.** Statistics of the Magnetization Model (A/m)

|            | $M_r$ | $M_\theta$ | $M_\phi$ | $ M $ | $ M_{\text{south}} $ | $ M_{\text{north}} $ |
|------------|-------|------------|----------|-------|----------------------|----------------------|
| Mean       | 0.005 | 0.013      | 0.006    | 1.13  | 1.65                 | 0.60                 |
| $\sigma_m$ | 2.41  | 0.04       | 0.05     | 2.41  | 3.25                 | 1.04                 |

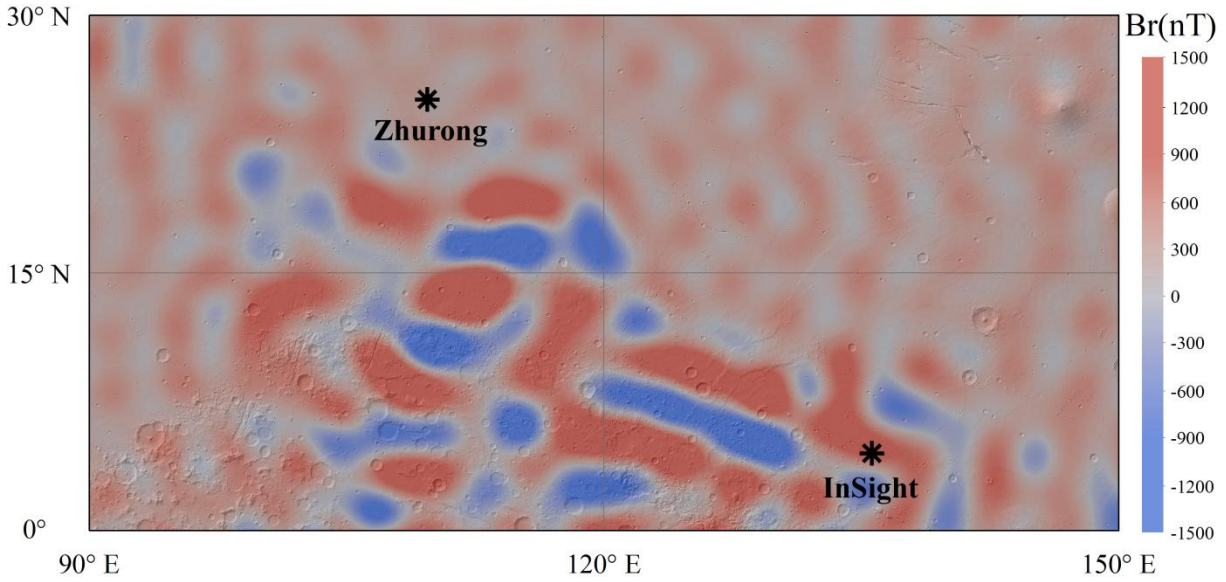
The global magnetization distribution shares morphological similarities with the surface magnetic field but exhibits differences in spatial patterns. For instance, at 120 km altitude magnetic anomalies are detected along the peripheries of impact basins such as Argyre, Isidis, and Hellas; the corresponding magnetization is located farther from the basin centers (Figure 7). This spatial decoupling demonstrates that our ESD method effectively constrains the approximate source distribution of crustal magnetization rather than merely replicating surface field patterns. The resulting magnetization model facilitates exploration of correlations between magnetic source characteristics and geological features, and probably provides new insights into crustal composition, impact history, and volcanic resurfacing processes.

#### 4.5 Comparisons with Ground-based Observations

The NASA InSight mission landed at  $4.5^\circ\text{N}$ ,  $135.6^\circ\text{E}$  in western Elysium Planitia in 2018 (Banerdt et al., 2020). The Zhurong rover of Tianwen-1 mission landed in the southern Utopia Basin at  $25.065^\circ\text{N}$ ,  $109.926^\circ\text{E}$  in 2021 (Wu B et al., 2022). Both missions have taken magnetic field measurements at their locations, which show significant differences. InSight measured a magnetic field intensity of about 2,000 nT, while the measured average horizontal magnetic field intensity of Zhurong on its path was less than 20 nT.

Figure 8 shows the predicted crustal field  $B_r$  regional maps in the vicinity of the InSight landing site and Zhurong at Martian surface by our model. Table 3 compares model predictions with ground measurements. For the G110 model, we truncate it to degree 80, since Gao JW et al. (2021) recommended downward continuation to the surface of Mars using degree 80. For InSight, our model predicts a magnetic field intensity of 547 nT, closer to the observed 2,013 nT than previous models, though all models underestimate the measured intensity by an order of magnitude. At Zhurong's location, the rover's 1,089 m traverse yielded 16 discrete measurements using its magnetometer. Our model predicts a horizontal magnetic intensity of 35.4 nT, which is the same order of magnitude as the measured average of 11.2 nT ( $\sigma = 10.9$

nT). The vertical component measurements could not be reliably calibrated, and therefore only the horizontal magnetic intensity is presented. The measured northward component averaged  $-5.1$  nT ( $\sigma = 13.6$  nT) while the eastward component averaged  $9.9$  nT ( $\sigma = 11.0$  nT), compared to our model predictions of  $-5.2$  nT (northward) and  $35.1$  nT (eastward) at the landing site. Meanwhile, other models except G110 also give relatively low horizontal component prediction misfits, less than the error, when the model is continued downward. These comparisons are order-of-magnitude checks rather than quantitative benchmarks. Truncated at degree 110, the model represents the field averaged over roughly  $190$  km at the surface, whereas the rover samples point values along a  $1,089$  m traverse; the rover's own measurements vary by a comparable factor within that traverse. A point measurement therefore records local crustal heterogeneity that satellite data cannot resolve, and a mismatch of the size seen here is expected in either direction.



**Figure 8.** Martian regional crustal field map of the spherical harmonic model up to  $n, m = 110$  in the vicinity of InSight and Zhurong landing sites at the surface. The radial magnetic field  $B_r$  is superimposed on a regional shaded relief map of Mars.

**Table 3.** Crustal field prediction and measurement of our model and other models in the vicinity of the InSight and Zhurong landing sites at the surface.

|                             | Observation | M110 | L134 | G110  | This study |
|-----------------------------|-------------|------|------|-------|------------|
| InSight $ \mathbf{B} $ (nT) | 2013        | 101  | 312  | 287   | 547        |
| Zhurong $B_h$ (nT)          | 11.2        | 30.9 | 51.8 | 148.6 | 35.5       |

Despite the improved performance of our models, there remains a consistent underestimation of intense surface magnetic fields, which underscores a fundamental limitation in current crustal field modeling. Specifically, the absence of low-altitude magnetic data hinders the resolution of small-scale and high-intensity anomalies near the Martian surface. The discrepancies observed between model predictions and measurements at the InSight landing site may indicate the presence of localized magnetization structures that lie below the resolution capabilities of existing datasets. Compounding these limitations, Zhurong's in-situ measurements suggest that

the downward extrapolation of high degree spherical harmonics inherently magnifies errors, further restricting the reliability of near-surface magnetic field reconstruction.

The InSight landing site is located approximately 500 km away from the Martian crustal dichotomy boundary. Geological evidence suggests that the crust beneath this site consists of volcanic flows from the Early Amazonian to Hesperian periods, dating between 3.6 and 1.5 billion years ago, with an estimated depth not exceeding 170 m (Knapmeyer-Endrun et al., 2025). Beneath these flows lies the Hesperian–Noachian transition unit, which is thought to extend to depths of roughly 1 to 5 km (Pan L et al., 2020). Further below, the basement rock is likely a continuation of the Noachian unit up to 10 km depth; beneath that, it could be related to fractured impact ejecta from Utopia, extending to depths of up to 20 km (Knapmeyer-Endrun et al., 2025). The deepest layer (20–39 km) is interpreted as primary crust formed by differentiation of the Borealis impact melt sheet (Wieczorek et al., 2022). Johnson et al. (2020) estimated the magnetization required to account for the local magnetic field observed at the InSight landing site, assuming it is carried by a horizontally layered structure beginning at a burial depth of 200 m and extending through a 40-km-thick magnetized section. This assumed thickness is comparable to the magnetized layer thickness adopted in our model. Under these assumptions, the minimum magnetization necessary to reproduce the observed field is approximately 0.4 A/m. In comparison, our model yields a local magnetization estimate of 0.42 A/m with an  $\sim 110$  km spatial scale. The shallow burial depth of 200 m suggests that at least part of the Hesperian–Noachian transition unit, or possibly the early Hesperian terrain, may be magnetized. This implication raises the possibility that Mars’s internal dynamo persisted beyond the traditionally accepted termination time, potentially extending into the Hesperian period (Johnson et al., 2020).

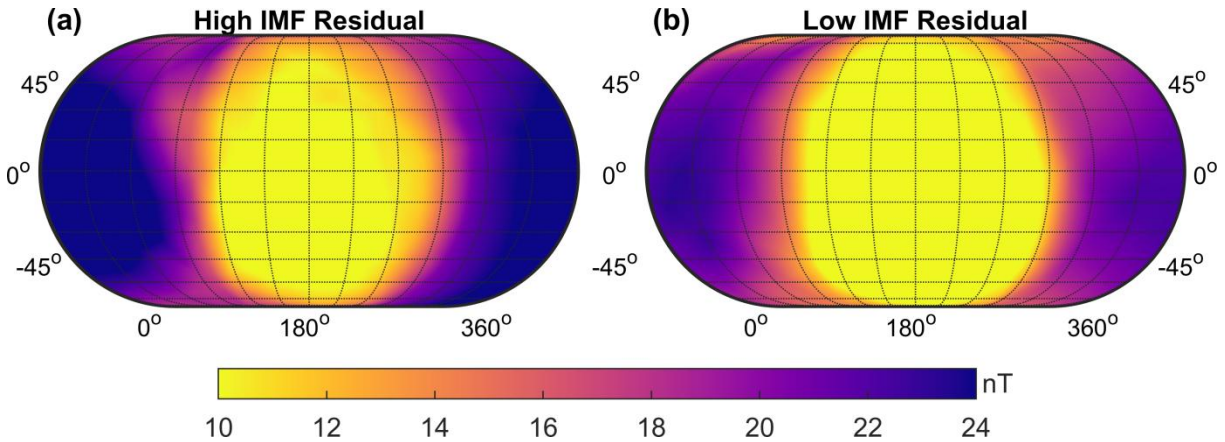
The Zhurong rover landed in a geologically simple region within Utopia Planitia. Based on estimates of crater density and surface morphology, the terrain is dated to the Hesperian to Amazonian periods (Tanaka et al., 2014). The stratigraphic data suggest that the ancient Noachian basement, lying several kilometers beneath the surface, was inundated by Hesperian-aged volcanic lava flows with an estimated thickness of 1–2 kilometers (Du AM et al., 2023). Overlying this volcanic plain is the Vastitas Borealis Formation (VBF), composed of late Hesperian sediments. Crater counting analysis reveals that the emplacement of VBF sediments and volcanic deposits may have transformed older Hesperian craters into ghost craters by approximately 3.45 billion years ago (Wu X et al., 2021). Subsurface radar data from Zhurong shows that the site is covered by Amazonian-aged materials tens of meters thick, topped by a regolith layer several meters deep (Li C et al., 2022). Du AM et al. (2023) propose that the extremely low magnetic anomaly observed at the Zhurong rover’s landing site may be attributed to the ghost crater with a radius of 5 km (Zhao JN et al., 2021). To test this hypothesis, they modeled a magnetized layer with dimensions of  $1,000 \text{ km} \times 1,000 \text{ km} \times 1 \text{ km}$ , a horizontal magnetization of 0.1 A/m and a vertical magnetization of 0.1 A/m. A cylindrical void was introduced at the center of the ghost crater to simulate the demagnetized region. The resulting horizontal magnetic field strength at an altitude of 200 m above the crater center was calculated to be 6 nT, while the field strength at the crater rim reached 41 nT, closely matching the maximum horizontal field strength observed by Zhurong rover. Based on these simulations, they estimate that the upper limit of magnetization required to reproduce the observed anomaly is approximately 0.14 A/m. In comparison, our ESD model yields a localized magnetization of 0.16 A/m, of the same order as this upper bound. We note, however, that the two estimates are not directly comparable: our value assumes a uniformly magnetized layer 40 km thick, whereas the

estimate of Du AM et al. (2023) assumes a 1-km-thick layer, and the inferred magnetization scales with the assumed source thickness and geometry. A magnetized layer with a thickness of 1 km corresponds to Hesperian-aged volcanic lava flows, whereas prevailing studies attribute the primary crustal magnetization to deeper Noachian basement. Nevertheless, satellite-derived models remain valuable for comparison with in-situ measurements and can provide partial constraints on the Martian dynamo and its geological evolution.

#### 4.6 Residual Analysis

Through a residual distribution map of the MSO coordinate system in Figure 9, we visualize the intensity of predicted residuals, defined as the magnitude of the vector residual:  $|\Delta \mathbf{B}| =$

$\sqrt{(|\Delta B_r|^2 + |\Delta B_\theta|^2 + |\Delta B_\phi|^2)}$ . A notable observation is the asymmetric influence of the solar wind between the dusk (longitudes 90°) and dawn (longitudes 270°) sides of Mars. Under low upstream IMF conditions, this asymmetry manifests as a pronounced difference in the spatial distribution of residuals, with the dusk side exhibiting larger residuals compared to the dawn side. This finding necessitates a re-evaluation of data selection criteria, suggesting that dusk-side observations under low IMF conditions require stricter spatial constraints to minimize bias. Furthermore, we observe that higher IMF intensities significantly reduce the spatial extent of low-residual regions, implying that stronger IMF amplifies external noise contributions to the magnetic field measurements.



**Figure 9.** Residual distribution of magnetic field data in the MSO coordinate system at 400 km altitude under varying IMF conditions. (a) High IMF ( $|\mathbf{B}_{IMF}| > 2.6$  nT); (b) Low IMF ( $|\mathbf{B}_{IMF}| < 2.6$  nT).

The observed residual asymmetry may be linked to asymmetries in the Martian magnetotail. As demonstrated in global magnetohydrodynamic (MHD) simulations of Mars (Liemohn et al., 2007), the IMF in its Parker spiral configuration generates an S-shaped topology near the planet's equatorial plane. This configuration preferentially enhances solar wind-magnetosphere interactions on the dusk side, while suppressing them on the dawn side, consistent with our residual distribution patterns. This agreement validates our initial approach of selecting night-side data using an asymmetrical edge, which refers to a longitude range in the MSO coordinate system that is not symmetrically distributed. Notably, when IMF intensities are elevated, the residuals increase, indicating heightened complexity in Martian space weather under such conditions.

Residual analysis serves as a powerful tool for understanding the interplay between external fields and crustal magnetic sources. By correlating residuals with variables such as location (MSO coordinates), season, solar activity phase, crustal field strength, and terrain, we can establish data selection criteria that account for external perturbations. Additionally, machine learning techniques offer a pathway to develop rapid data filtering methods without requiring explicit physical models of noise sources. Crucially, residual modeling enables the isolation of external field contributions, providing insights into solar wind-crustal field interactions. This approach can be effectively combined with orbit-based screening strategies, such as those introduced in Section 2, to further enhance data reliability and exclude segments influenced by non-crustal disturbances. However, our results highlight the need for more robust inversion methods to separate noise from multiple sources and improve the accuracy of crustal magnetic field models.

## 5. Conclusions

In this study, we have developed a new equivalent source dipole model of the Martian crustal magnetic field by combining magnetic field data from the MGS, MAVEN, and China's Tianwen-1 mission. The model is transformed into a spherical harmonic representation, demonstrating convergence up to degree 110. Our model is also compared with ground-based magnetic field measurements from NASA's InSight lander and the Zhurong rover of the Tianwen-1 mission. Several novel elements distinguish our model from existing Martian crustal magnetic field models (e.g. Morschhauser et al., 2014; Gao JW et al., 2021; Langlais et al., 2019), both in terms of data and inversion methodology:

1. This is the first Martian crustal field model to incorporate Tianwen-1 magnetometer observations. Our model draws on magnetic field observations from three orbiters and is based on rigorous, carefully designed data-selection criteria. The retained Tianwen-1 data provide limited low-altitude coverage after screening. They therefore broaden the multi-mission basis of the model and support a practical assessment of the dataset for crustal field modeling, rather than improving the global model.
2. We utilize the Fibonacci Sphere Distribution for dipole placement, ensuring a geometrically uniform grid that enhances the model's accuracy and stability.
3. The model employs a two-step inversion process. First, the radial component of the magnetic dipole moments is inverted. These results are then used as initial conditions for a full three-component inversion, implemented via the Iteratively Reweighted Least Squares (IRLS) method with Huber weighting. This approach reduces the influence of external magnetic sources, and prevents overfitting, thereby improving the robustness of the model.

We compared our model's predictions at the surface with ground-based magnetic measurements and magnetization estimates from the InSight lander (Johnson et al., 2020) and the Zhurong Rover (Du AM et al., 2023). At the InSight landing site, our model predicts a magnetic field intensity of 547 nT, which is closer to the measured value of 2,013 nT than predictions from previous models. At the Zhurong landing site, our model predicts a horizontal magnetic field intensity of 35.4 nT, showing general agreement in magnitude with the rover's measured value

of 11.2 nT. Furthermore, the magnetization strength inferred from our model at both landing sites is comparable to the estimated crustal values. However, intense regional magnetic fields near the surface are still not fully resolved, likely due to the altitude limitations of the available data. In addition, small-scale noise fields tend to be amplified during downward continuation, obscuring the true magnetic field distribution.

Finally, we performed a residual analysis that reveals important insights into the influence of external magnetic fields on the Martian crustal magnetic field. We observed a dawn-to-dusk asymmetry in the residual distribution, particularly under low interplanetary magnetic field (IMF) conditions. The residual analysis highlights the importance of considering external field sources in future modeling efforts. The use of Huber weighting in our inversion process effectively reduces the impact of outliers and external noise, but the complexity of Mars' space weather, requires more sophisticated methods to fully account for external field contributions. Future studies could benefit from incorporating more detailed external field models or using machine learning techniques to better filter out noise and improve data selection.

## 6. Software and third party data repository citations

The MGS and MAVEN magnetic field data are available at the NASA Planetary Data System (<https://pds-ppi.igpp.ucla.edu>); specifically, the MAVEN MAG data (Connerney, 2017), the MGS AB/SPO MAG data (Connerney, 2025b) and the MGS MPO MAG data (Connerney, 2025a) are available. The Tianwen-1 magnetic field data are available at the Lunar and Planetary Data Release System (<https://moon.bao.ac.cn>). The numerical code used in this study is available at <https://github.com/southminanmi/A-Martian-ESD-Model>. The Martian crustal magnetic field model and relevant data (Tang XF and Lin YF, 2025) produced in this study is available at <https://doi.org/10.5281/zenodo.15049439>.

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## Appendix A. Non-uniqueness of the ESD parameterization and the two-step inversion

The dipole sources lie on a spherical layer of radius  $c = a - 20$  km (Section 3.2). Any magnetization distribution on such a layer can be written as a radial part plus the Helmholtz decomposition of its tangential part,

$$\mathbf{m}(\Omega) = m_r(\Omega) \hat{\mathbf{r}} + \nabla_s \Phi(\Omega) + \hat{\mathbf{r}} \times \nabla_s \Psi(\Omega), \quad (\text{A1})$$

where  $\nabla_s$  is the surface gradient and  $\Phi, \Psi$  are scalar fields on the sphere. Writing the layer potential as  $V(\mathbf{r}) = \int \mathbf{m}(\Omega') \cdot \nabla' |\mathbf{r} - \mathbf{r}'|^{-1} dS'$  (constant physical factors omitted), inserting the exterior expansion of  $|\mathbf{r} - \mathbf{r}'|^{-1}$  and integrating the tangential term by parts on the sphere gives, for  $r > c$ ,

$$V(r, \Omega) = \sum_{l \geq 1} \sum_m \frac{4\pi}{2l+1} \frac{c^{l+1}}{r^{l+1}} l [(m_r)_{lm} + (l+1) \Phi_{lm}] Y_{lm}(\Omega). \quad (\text{A2})$$

The toroidal scalar  $\Psi$  does not appear, because  $\nabla_s \cdot (\hat{\mathbf{r}} \times \nabla_s \Psi) \equiv 0$ : a toroidal magnetization carries no effective magnetic charge and produces no external field. Toroidal distributions are therefore one family of annihilators (Gubbins et al., 2011). Equation (A2) shows further that the external field constrains, per degree and order, only the single combination

$$u_{lm} = (m_r)_{lm} + (l+1) \Phi_{lm}, \quad (\text{A3})$$

so two of the three scalar fields describing  $\mathbf{m}$  cannot be determined from exterior data. Equation (A3) also identifies the second family of annihilators, the distributions with  $u_{lm} = 0$  for every degree and order. Runcorn's theorem, invoked in Section 3.2, is an instance of this second family rather than of the toroidal one: a shell magnetized by a field from an internal source produces no exterior field (Runcorn, 1975), and by (A2) its  $u_{lm}$  therefore vanish. Together the two families are the ESD non-uniqueness referred to in Section 3.3. None of this means that tangential magnetization produces no exterior field: a tangential poloidal distribution does produce one, and by (A3) it is as effective a source as a radial one at high degree. What the data leave open is how  $u_{lm}$  divides between  $m_r$  and  $\Phi$ .

Because only  $u_{lm}$  is observable, a purely radial layer can reproduce any field that an arbitrary three-component layer produces: given  $\mathbf{m}$ , the radial distribution defined by

$$(m'_r)_{lm} = (m_r)_{lm} + (l+1) \Phi_{lm}, \quad \Phi' = \Psi' = 0 \quad (\text{A4})$$

has the same exterior potential. The radial representative is also unique. With  $\Phi = \Psi = 0$ , equation (A2) reduces to  $V_{lm} = \alpha_l (m_r)_{lm}$  with  $\alpha_l = 4\pi l c^{l+1}/(2l+1) \neq 0$  for  $l \geq 1$ , a one-to-one map on magnetizations of zero spherical mean, the mean itself producing no external field. Restricting the first inversion step to  $M_r$  is therefore not a physical assumption that the crust is radially magnetized. It is a choice of representative, a gauge, among the infinitely many magnetizations consistent with the same observations, and it discards no observable information while removing the ambiguity of (A3) completely. The recovered  $M_r$  should accordingly be read as the radial equivalent layer that reproduces the observed field, not as the radial component of the true crustal magnetization, which exterior data cannot determine.

The full-vector step solves the weighted normal equations by conjugate gradients starting from  $\mathbf{m}_0 = (M_r^*, 0, 0)$ . Every iterate satisfies  $\mathbf{m}_k - \mathbf{m}_0 \in \text{Range}(\mathbf{G}^T \mathbf{W} \mathbf{G}) \subseteq \text{Null}(\mathbf{G})^\perp$ , and Huber reweighting by any positive diagonal  $\mathbf{W}$  leaves  $\text{Null}(\mathbf{G})$  unchanged, since  $\mathbf{W}^{1/2} \mathbf{G} \mathbf{x} = \mathbf{0}$  if and only if  $\mathbf{G} \mathbf{x} = \mathbf{0}$ . The component of the solution in  $\text{Null}(\mathbf{G})$  is therefore fixed by the starting model for any number of iterations and reweightings, and equals that of the radial model obtained in the first step. If the iteration is run to convergence, the scheme returns the data-fitting model closest to  $\mathbf{m}_0$ , which is the zero-damping limit of Tikhonov regularization towards  $\mathbf{m}_0$ . These statements are exact for the continuous layer and, in exact arithmetic, for whatever null space the discretized operator possesses. For a finite dipole set the annihilator directions

correspond to very small rather than exactly zero singular values of  $\mathbf{G}$ , and the corresponding practical statement is that a finite number of conjugate-gradient iterations leaves these weakly constrained directions close to their initial values. This is the sense in which the two-step procedure mitigates the magnetic annihilators and removes the dependence on an arbitrary initial model (Section 3.3): the initial model still selects the null-space content, as it must, but it now selects the unique, fully specified radial-gauge member of the solution family.

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