



Magnesium isotopes constraint on the origin of stratiform chromitite

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Abstract

Resolving the debate on whether magmas are saturated solely in chromite or if chromite reaches a cotectic with other liquidus phases before segregating is crucial to understanding the origin of stratiform chromitite. Probing the Mg isotopic variations provide insight into this issue, as chromite crystallization significantly fractionates melt $\delta^{26}\text{Mg}$. Here, we present $\delta^{26}\text{Mg}$ data for whole rocks from the Stillwater Complex to assess their Mg isotopic evolution. The Banded Series are isotopically homogeneous with $\delta^{26}\text{Mg}$ averaging $-0.29 \pm 0.05\text{‰}$ ($n=10$), which is lighter than komatiitic melts. Rayleigh distillation modeling shows that generating this signature via chromite-only saturation would require 15 wt% crystallization. In contrast, cotectic crystallization of chromite and silicates rapidly depletes the melt's Mg budget, necessitating only ~ 1.7 wt% chromite to produce the same isotopic shift. This strongly supports a model where chromitite layers form by mechanical sorting of chromite from a cotectic crystal mush. Olivine cumulates and granular harzburgites have average $\delta^{26}\text{Mg}$ values of -0.17‰ , which are isotopically heavier than poikilitic harzburgites with $\delta^{26}\text{Mg}$ values of -0.32‰ . Chromitite samples generally exhibit lighter Mg isotopes, with a wider range from $-0.32 \pm 0.03\text{‰}$ to $+0.03 \pm 0.04\text{‰}$. Modeling results suggest that fluid migration between chromitite and silicate cumulates does not cause their Mg isotopic heterogeneity but rather serves as a critical medium for mechanical sorting. The extensive occurrence of Mg isotopic heterogeneity in Ultramafic Series implies that effects of fluid migration need to be considered in studies of basaltic magma chamber processes, evolution, and oxide ore formation.

Keywords Stillwater complex · Chromitite · Cotectic model · Fluid migration · Mg isotopes

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Introduction

Layered intrusions, characterized by alternating layers of different rock types, serve as natural laboratories for studying magma chamber processes. They also host precious metal deposits and, among them, major resources for chromium in chromite. The local extreme enrichment of chromite in layers is one of the most distinctive and enigmatic geological features in these intrusions. Traditionally, interpretations of its origin focus on mechanisms that suggest magmas are saturated solely in chromite through: (1) magma hybridization with differentiated melts (Campbell and Murck 1993), assimilated country rocks (Spandler et al. 2005), or variable mantle sources, yielding magmas with distinct initial compositions that favor chromite saturation (Kinnaird et al. 2002); (2) fluctuations in pressure or oxygen fugacity induced by magma-chamber deformation (Liu and Presnall 1990), bubble expansions (Lipin 1993), magmatic replenishment (Cawthorn 2012), and gas release (Ulmer 1969); (3) incongruent partial melting of pre-existing cumulates

(Nicholson and Mathez 1991) or by the assimilation and chemical upgrading of Fe-Ti oxide xenocrysts from country rocks (Leshner et al. 2019); (4) water-induced increases in oxygen fugacity that promote early spinel crystallization (Veksler and Hou 2020). In these models, the melt becoming supersaturated in chromite as the sole liquidus phase during initial fractional crystallization is considered crucial for the formation of stratiform chromitite.

Experimental studies have shown that Cr depletion in melts predominantly occurs during cotectic crystallization of oxides and silicates, making the idea of chromite supersaturation geologically unreasonable and transient (e.g., Leuthold et al. 2015; Drage and Brenan 2023). Consequently, alternative hypotheses have emerged that do not require the assumption of chromite being the sole liquidus phase. These alternative hypotheses propose that predominantly silicate phases, with minor chromite, are cotectic in proportions and form a slurry, either in shallow conduits (Jenkins and Mungall 2018) or in deep-seated vast staging chambers (Eales 2000; Mondal and Mathez 2007; Eales and Costin 2012; Mungall et al. 2016). Monomineralic chromitite layers would form through the separation of heavy, small-sized chromite from lighter, larger silicate minerals during crystal settling in flowing magmas (Eales 2000; Eales and Costin 2012), earthquakes with associated shock waves (Elsworth and Voight 1995; Nex 2004), or through processes related to crystal slurry slumping, such as viscous particle segregation (Forien et al. 2015) or kinetic sieving (Maier et al. 2018; Mungall et al. 2016; Otto et al. 2025). These models largely deny the necessity for a chromite-only saturated interval and even the substantial chromite fractional crystallization due to insufficient Cr-depleted material in upper zones of layered intrusions (Campbell and Murck 1993; Eales 2000).

The debate surrounding the monomineralic crystallization of chromite has emerged as a highly contentious and fundamental issue in recent studies of layered intrusions. Evidence in various methodologies has been presented by proponents of different views, including the morphological studies (e.g., Latypov et al. 2022), analogous experiments (e.g., Forien et al. 2015), and thermodynamic modelling (e.g., Yao et al. 2021). Geochemical studies are commonly supposed to play a crucial role in understanding geological processes. However, when it comes to chromite, these studies have been overlooked due to compositional modifications of chromite by post-cumulus processes (review in Barnes et al. 2023). Therefore, identifying a reliable geochemical proxy that is sensitive to chromite crystallization is essential for resolving ongoing controversies. The advancement of high-precision magnesium (Mg) isotopic analysis offers new constraints in this area. On one hand, high-temperature magmatic processes, including partial melting and magmatic differentiation, are generally not

expected to significantly fractionate Mg isotopes (Teng et al. 2007, 2010; Huang et al. 2013; Liu et al. 2011). Subsequent low-temperature alteration processes have also been demonstrated to have a negligible effect on the bulk-rock Mg isotopic compositions of mafic and ultramafic rocks, ensuring the preservation of primary magmatic signatures (e.g., Wang et al. 2014). Additionally, significant alteration of the melt Mg isotopes by contamination with surrounding crustal rocks is unlikely, due to the large difference in Mg concentration between high-Mg komatiitic magmas and low-Mg crustal rocks (Dauphas et al. 2010). On the other hand, the accumulation of oxides like chromite fractionates Mg isotopes by preferentially incorporating heavier isotopes relative to co-crystallizing silicates. This process creates a significant isotopic contrast, which is robustly preserved even against potential post-cumulus diffusion. For example, studies of the Stillwater Complex show that chromite $\delta^{26}\text{Mg}$ values are consistently more than 0.6‰ higher than in coexisting silicates (Bai et al. 2021). As a result, extensive chromite crystallization drives the residual melt towards an isotopically lighter composition, making $\delta^{26}\text{Mg}$ a sensitive tracer for this process (Xiao et al. 2016, 2023; Su et al. 2017, 2019; Wang et al. 2021; Liu et al. 2024).

In this paper, we present new whole-rock Mg isotope data on the Stillwater Complex, including chromitite, poikilitic harzburgite, olivine cumulate, granular harzburgite, and bronzitite from the Ultramafic Series, as well as troctolite, anorthosite and gabbro-norite samples from the Banded Series. The Stillwater Complex, with its exceptionally well-exposed and complete magmatic stratigraphy, is a classic locality for studying large layered intrusions. It is also at the forefront of fundamental and ongoing debates regarding magma chamber processes, including controversies over in-situ crystallization versus replenishment by crystal slurries, the validity of “big tank” models, and even apparent out-of-sequence age constraints (e.g., Boudreau 2016, 2025; Barnes et al. 2016, 2023; Wall et al. 2018; Parker et al. 2022; Latypov et al. 2022, 2024; Otto et al. 2025). The origin of its significant stratiform chromitite layers is central to these debates. This context makes the Stillwater Complex an ideal natural laboratory to test how novel geochemical systems can provide new constraints on these long-standing problems. We therefore use our results, combined with prior Mg isotopic compositions of mineral separates (Bai et al. 2021), to provide new insight into the origin of its stratiform chromitites.

Geological setting and samples

The ~2.7 Ga Stillwater Complex is a layered ultramafic-mafic intrusion located along the northern front of the Beartooth Mountains in Montana, the USA (e.g., Wall et al. 2018). It has a maximum exposed stratigraphic thickness of 6.5 km and an exposed strike length of ~45 km along a northwest-southeast orientation (McCallum 1996). The dip of the complex varies from nearly vertical or overturned in the eastern part to about 60° N in the western part. The emplacement depth of the intrusion is debated; while some interpretations suggest a shallow, sub-volcanic setting (Lipin and Zientek 2002), metamorphic studies of the contact aureole indicate a deeper crustal emplacement at 12–16 km (e.g., Hanley et al. 2008). The intrusion was emplaced into a 3.3-Ga metasedimentary sequence comprising quartz monzonite, hornfels, pelitic schists, iron formations and quartzites (Page 1977, 1979; Helz 1995). The complex has undergone faulting, tilting, and erosion prior to the Middle Cambrian, with further deformation and uplift during the Laramide orogeny bringing it to its present position (Polovina et al. 2004). The Stillwater Complex has been extensively studied for its mineralogy, petrology and geochemistry (e.g., Raedeke and McCallum 1984; McCallum 1996; Jenkins and Mungall 2018; Bai et al. 2021, 2024; Su et al. 2020), and is subdivided into three major series based on the nature and proportions of cumulus minerals.

- I. **Basal Series** - The base of the first laterally continuous norite in the Basal Series defines the bottom of the complex, separating the overlying cumulates from the hornfels underneath. The norite (with subordinate anorthosite, gabbro and peridotite) grades upward with an irregular decrease in plagioclase content into bronzitite (Page 1977). Small bodies of massive sulfide occur within the hornfels, with a marked decrease in sulfide abundance entering the Basal Series (Zientek and Ripley 1990). One sample (16SW-3–10) with disseminated sulfides was collected in this series (Table 1).
- II. **Ultramafic Series** - The first significant appearance of accumulated olivine marks the base of the Ultramafic Series, which comprises a lower Peridotite Zone and an upper Bronzitite Zone (Fig. 1). The Peridotite Zone is characterized by cyclic units of olivine-chromite-orthopyroxene layers (Cooper 1997). Each cyclic unit consists, from the base upward, of poikilitic harzburgite, chromitite, disseminated chromitite, poikilitic harzburgite, olivine cumulate, granular harzburgite and bronzitite. At least 15 cyclic units have been recognized in the Peridotite Zone, with 11 chromitite layers, designated A to K from bottom to top, at the base of each unit. Significant chromitite has been extracted from two

principal mines: Benbow (B cyclic unit) to the east and Mountain View (G cyclic unit) near the intrusive center (Fig. S1). A total of seventeen samples, including massive, anti-nodular and disseminated chromitite, were collected from these locations, along with six un-mineralized silicate cumulate samples from the A cyclic unit in the Gish area for comparison (see photomicrographs of representative samples in Fig. S2). The overlying Bronzitite Zone is a generally uniform orthopyroxene cumulate. One bronzitite sample (16SW-2-2) was collected in the Bronzitite Zone. It is important to note that the relative timing and emplacement of these zones are subjects of significant debate. While traditionally viewed as forming sequentially from bottom to top, recent high-precision U-Pb zircon geochronology by Wall et al. (2018) challenged this model. Their data suggest an “out-of-sequence” emplacement, proposing, for instance, that the upper Bronzitite Zone (~2711 Ma) is actually older than parts of the underlying Peridotite Zone (~2710 Ma). This model posits that the entire Stillwater Complex was assembled over ~3 million years as a stack of amalgamated sills, with similar age discrepancies observed in the Banded Series. However, this interpretation is not universally accepted. Critics (e.g., Latypov, 2019) argue that such a model is difficult to reconcile with the petrologically coherent stratigraphy of the complex, which largely follows a predictable magmatic crystallization sequence. Thus, the precise emplacement mechanism and the significance of these zircon ages remain an active area of discussion.

- III. **Banded Series** - The base of the Banded Series is defined by the appearance of cumulus-textured plagioclase. This series is subdivided into the Lower, Middle, and Upper Banded series and is composed mostly of gabbronorite, norite, and anorthosite. These are interspersed with several distinct olivine-bearing zones (termed OB-I to OB-V), which are composed of troctolite and olivine gabbronorite. The OB-I hosts J-M Reef with Pt-Pd mineralization, and is composed of a package of lithologies including troctolite, anorthosite, and pyroxenite, often with pegmatitic textures (Barnes et al. 2020; Jenkins et al. 2021). It is worth noting that while the Banded Series are often interpreted as originating from distinct magma lineages with the Ultramafic Series, gradational transitions from pyroxenite to gabbronorite have been observed in drill cores across the boundary. Throughout most of the Banded Series, oxides such as magnetite and ilmenite occur only as accessory minerals (Boudreau and McCallum 1986). Eight samples from the Banded Series, including gabbronorite, anorthosite and troctolite, were collected

Table 1 Locations, coordinates, mineral assemblages and $\delta^{26}\text{Mg}$ values of various rocks from the Stillwater Complex

Stratigraphic position	Outcrop	Latitude	Longitude	Sample	Lithology	Mineral assemblage	$\delta^{26}\text{Mg}$ (‰)	2SD
Upper Banded Series	Picket Pin Mountain	45.4529°N	110.0497°W	16SW-1-1	Gabbronorite	Opx + Cpx + Pl	−0.29	0.01
Middle Banded Series	Picket Pin Mountain	45.4529°N	110.0497°W	16SW-4-1	Troctolite	Pl + Ol	−0.31	0.04
Middle Banded Series	Picket Pin Mountain	45.4529°N	110.0497°W	16SW-1-2	Anorthosite	Pl	−0.18	0.04
Lower Banded Series	Mountain View	45.3818°N	109.8785°W	16SW-1-3	Gabbronorite	Opx + Cpx + Pl	−0.32	0.06
Lower Banded Series	Mountain View	45.3818°N	109.8785°W	16SW-1-5	Gabbronorite	Opx + Cpx + Pl	−0.30	0.05
Lower Banded Series	Mountain View	45.3818°N	109.8785°W	16SW-1-6	Gabbronorite	Opx + Cpx + Pl	−0.30	0.02
Lower Banded Series	East Boulder Mine	45.3818°N	109.8785°W	16SW-3-15	Troctolite	Pl + Ol	−0.25	0.03
Lower Banded Series	East Boulder Mine	45.3818°N	109.8785°W	16SW-3-16	Troctolite	Pl + Ol	−0.29	0.05
Bronzitite Zone	Benbow Mine	45.3601°N	109.8022°W	16SW-2-2	Bronzitite	Opx	−0.31	0.06
G cyclic unit	Mountain View	45.3818°N	109.8785°W	16SW-1-17	Poikilitic harzburgite	Ol + Opx + Cpx + Pl	−0.31	0.05
G cyclic unit	Mountain View	45.3818°N	109.8785°W	16SW-1-15	Poikilitic harzburgite	Chr + Ol + Opx + Pl	−0.29	0.06
G cyclic unit	Mountain View	45.3818°N	109.8785°W	16SW-1-11	Poikilitic harzburgite	Chr + Ol + Opx + Pl	−0.29	0.08
G cyclic unit	Mountain View	45.3818°N	109.8785°W	16SW-1-10	Poikilitic harzburgite	Chr + Ol + Opx	−0.35	0.07
G cyclic unit	Mountain View	45.3818°N	109.8785°W	16SW-1-9	Disseminated chromitite	Chr + Ol + Opx	−0.32	0.03
G cyclic unit	Mountain View	45.3818°N	109.8785°W	16SW-1-27	Disseminated chromitite	Chr + Ol + Opx	−0.26	0.02
G cyclic unit	Mountain View	45.3818°N	109.8785°W	16SW-1-8	Anti-nodular chromitite	Chr + Ol	−0.32	0.04
G cyclic unit	Mountain View	45.3818°N	109.8785°W	16SW-1-26	Anti-nodular chromitite	Chr + Ol	−0.16	0.04
G cyclic unit	Mountain View	45.3818°N	109.8785°W	16SW-1-30	Massive chromitite	Chr	−0.13	0.03
G cyclic unit	Mountain View	45.3818°N	109.8785°W	16SW-1-23	Massive chromitite	Chr	−0.14	0.01
B cyclic unit	Benbow Mine	45.3601°N	109.8022°W	16SW-2-12	Granular harzburgite	Ol + Chr + Opx + Cpx	−0.17	0.02
B cyclic unit	Benbow Mine	45.3601°N	109.8022°W	16SW-2-21	Poikilitic harzburgite	Chr + Ol + Opx	−0.31	0.04
B cyclic unit	Benbow Mine	45.3601°N	109.8022°W	16SW-2-4	Disseminated chromitite	Chr + Ol + Opx	−0.25	0.06
B cyclic unit	Benbow Mine	45.3601°N	109.8022°W	16SW-2-8	Disseminated chromitite	Chr + Ol + Opx	−0.13	0.05
B cyclic unit	Benbow Mine	45.3601°N	109.8022°W	16SW-2-14	Massive chromitite	Chr	0.02	0.01
B cyclic unit	Benbow Mine	45.3601°N	109.8022°W	16SW-2-6	Massive chromitite	Chr	0.03	0.04
B cyclic unit	Benbow Mine	45.3601°N	109.8022°W	16SW-2-13	Massive chromitite	Chr	0.03	0.03
A cyclic unit	Gish Mine	45.4811°N	110.2029°W	16SW-3-5	Granular harzburgite	Ol + Opx + Cpx	−0.21	0.03
A cyclic unit	Gish Mine	45.4811°N	110.2029°W	16SW-3-4	Olivine cumulate	Ol + Opx	−0.19	0.05
A cyclic unit	Gish Mine	45.4811°N	110.2029°W	16SW-3-2	Olivine cumulate	Ol + Opx + Cpx	−0.14	0.04
A cyclic unit	Gish Mine	45.4811°N	110.2029°W	16SW-3-3	Olivine cumulate	Ol + Opx	−0.13	0.03
A cyclic unit	Gish Mine	45.4811°N	110.2029°W	16SW-3-9	Granular harzburgite	Ol + Opx + Cpx	−0.14	0.05
A cyclic unit	Gish Mine	45.4811°N	110.2029°W	16SW-3-11	Granular harzburgite	Ol + Opx + Cpx + Pl	−0.18	0.03
Basal Series	Gish Mine	45.4811°N	110.2029°W	16SW-3-10	Disseminated sulfides	Py + Pn	−0.25	0.04

Chr, chromite; Cpx, clinopyroxene; Opx, orthopyroxene; Ol, olivine; Pl, plagioclase, Py, pyrite; Pn, pentlandite.

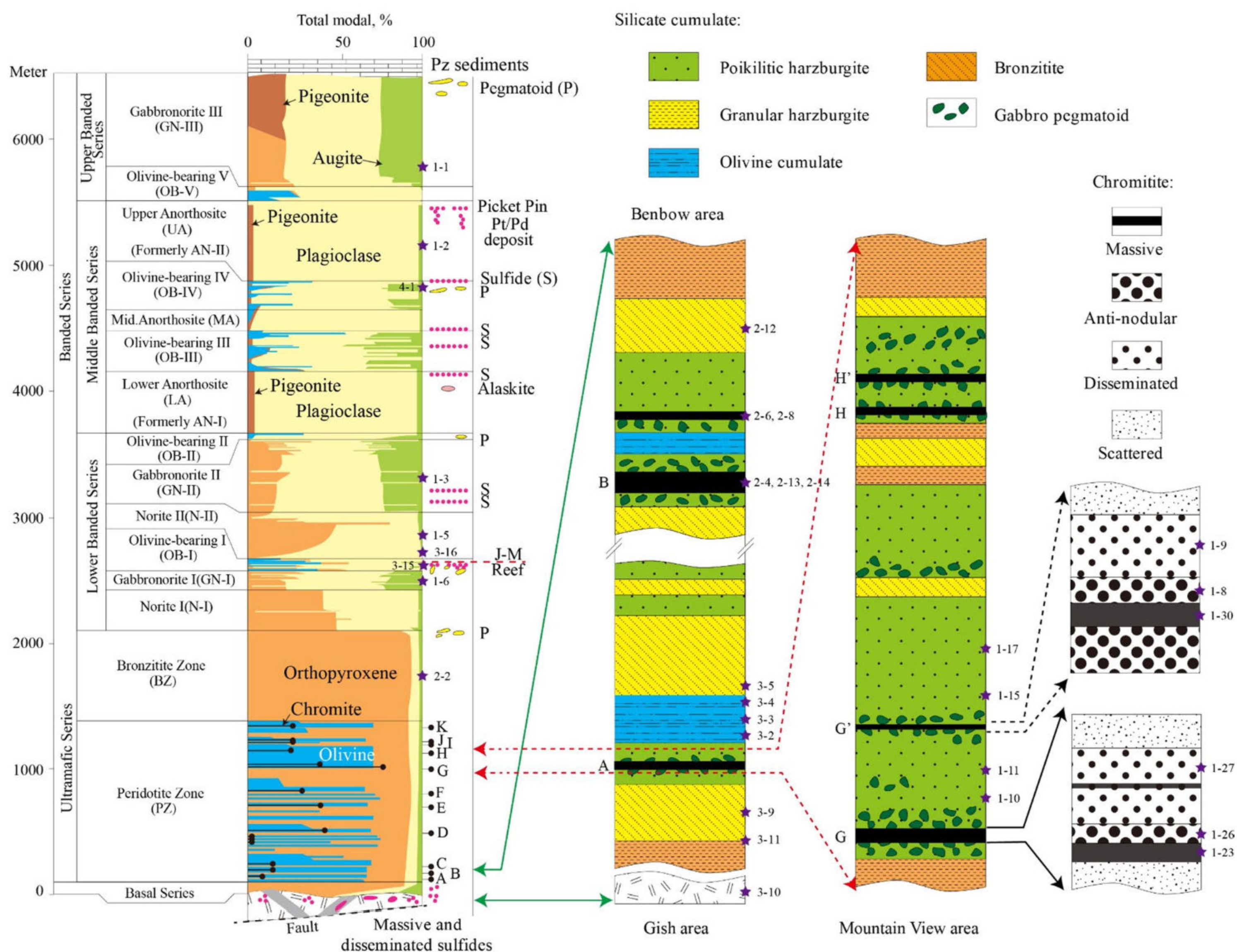


Fig. 1 Stratigraphy of the Stillwater Complex with major mineral modals as a function of height, modified from Boudreau (2016). Mineral modal content and stratigraphic units after Raedeke and McCal-

from East Boulder Mine, Mountain View and Picket Pin Mountain, respectively (Fig. S1).

Sample Preparation and analytical methods

Sample digestion and chemical separation

The procedures for sample dissolution, column chemistry, and instrumental analysis closely follow those reported in previous studies (e.g., Teng et al. 2010, 2015; Ke et al. 2016), and are briefly summarized here. Silicate cumulates were crushed to ~200 mesh and dissolved overnight in a mixture of HF, HCl, and HNO₃ at 130 °C in capped Savillex beakers. Chromitite samples were ground to a very fine powder and dissolved in concentrated 3:1 HF-HNO₃ in a microwave oven, with temperatures increasing to 180, 210, 220, and 225 °C at half-hour intervals. All solutions were dried on a hot plate and sequentially treated by aqua

regia and concentrated HNO₃. The samples were then dried and re-dissolved in 1 N HNO₃ in preparation for chromatographic separation. Mg was purified by cation exchange chromatography with Bio-Rad AG50W-X8 (200–400 mesh) resin at the Isotope Geochemistry Lab, China University of Geosciences (Beijing), following established procedures (e.g., Dauphas et al. 2010; Teng et al. 2010, 2015; Xiao et al. 2016; Ke et al. 2016). The purification procedure was performed twice to ensure the purity of sample Mg obtained. Routine recovery was >99%, and the whole procedure blank was monitored to be <10 ng, which is negligible compared to the ~20 µg of Mg processed.

Mass spectrometry

The Mg isotopic measurements were performed on Thermo Finnigan Neptune Plus multi-collector inductively coupled plasma mass spectrometers using the sample-standard bracketing method. Each measurement consisted of a 3-s idle time and 40 cycles of 4.194 s integration times. A 3% HNO₃ blank was measured at the beginning of each analytical session for on-peak-zero correction. Isotopic data are expressed in standard δ notation as per-mil deviations relative to the DSM-3 standards as:

δ²⁶Mg [0/00] = [(²⁶Mg/²⁴Mg)_{sample} / (²⁶Mg/²⁴Mg)_{DSM-3} - 1] × 1000

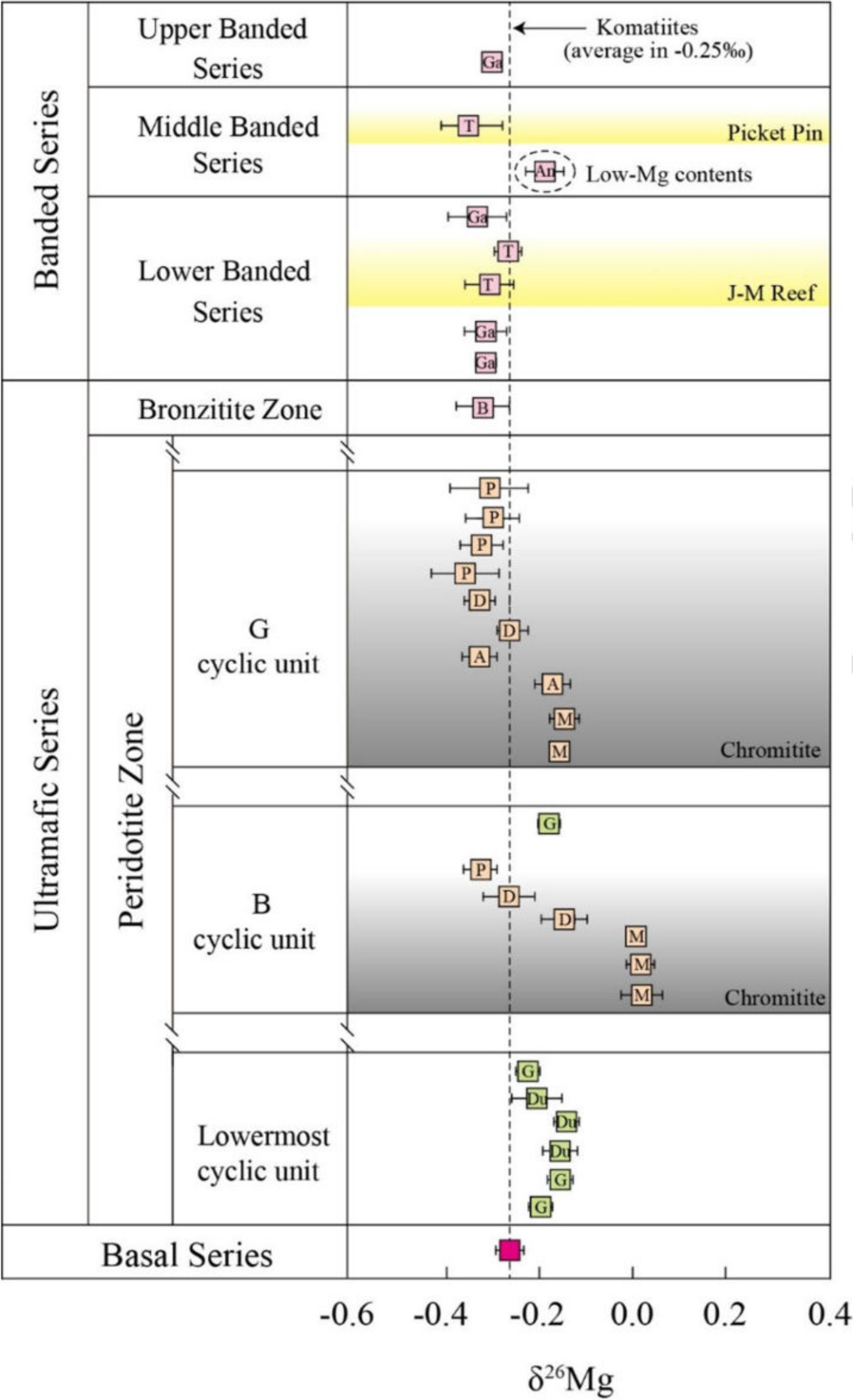


Fig. 2 Mg isotopic compositions for whole rocks in the stratigraphic section of the Stillwater Complex. M Massive chromitite, A Anti-nodular chromitite, D Disseminated chromitite, P Poikilitic harzburgite, G Granular harzburgite, B Bronzite, Oc Olivine cumulate, T Troctolite, Ga Gabbronorite, An Anorthosite

To enhance the reproducibility and accuracy of the measurements, the standard-sample sequences were repeated four times. The isotopic compositions reported herein are the average values of the repeated analyses, and uncertainties are reported as two standard deviations (2SD) for δ²⁶Mg. Two international rocky standards, BHVO-2 and JP-1, were processed along with unknowns to monitor accuracy throughout the column chemistry and isotopic analyses (Measured by the critical mixture double spike technique; He et al. 2022), and yielded δ²⁶Mg values (BHVO-2: δ²⁶Mg = -0.19 ± 0.04‰ (2SD); JP-1: δ²⁶Mg = -0.23 ± 0.13‰ (2SD) consistent with their recommended values (BHVO-2: δ²⁶Mg = -0.20 ± 0.07‰ (2SD); JP-1: δ²⁶Mg = -0.25 ± 0.02‰ (2SD) within quoted errors (Teng et al. 2015; Ke et al. 2016).

Results

The complete Mg isotopic results for the samples are reported in Table 1 and are presented in stratigraphic profiles in Fig. 2. As shown in Fig. 2, the Banded Series exhibits very limited Mg isotopic variability. Except for the anorthosite outlier (16SW-1-2), which has high δ²⁶Mg values of (-0.18 ± 0.04‰) and low MgO contents (3.18 wt%; Table S1 and Fig. S3), other samples from the Upper and Lower Banded Series have variable MgO contents ranging from 6.74 wt% to 24.4 wt% (average 13.4 wt%) and consistently exhibit limited values of δ²⁶Mg ranging from -0.32 ± 0.06‰ to -0.25 ± 0.03‰, with an average of -0.29 ± 0.05‰ (n = 10).

In contrast to the Banded Series, samples from the Ultramafic Series exhibit significant variations in δ²⁶Mg values within individual cyclic units (Fig. 2). In the B and G cyclic units, Mg isotopes of samples range from -0.31 ± 0.06‰ to +0.03 ± 0.04‰ and -0.32 ± 0.05‰ to -0.13 ± 0.03‰, respectively, indicating a gradual enrichment of ²⁴Mg from massive chromitite to anti-nodular chromitite to disseminated chromitite. Poikilitic harzburgite samples adjacent to chromitite layers in the G and B cyclic units have narrow range of δ²⁶Mg values from -0.35 ± 0.07‰ to -0.29 ± 0.06‰, with an average value of -0.32‰. The olivine cumulate and granular harzburgite samples from the B and A cyclic units display markedly heavier Mg isotopic compositions (-0.21 ± 0.03‰ to -0.13 ± 0.03‰; average δ²⁶Mg = -0.17‰) than poikilitic harzburgite, despite these silicate cumulates having similar mineral assemblages. Notably, although multiple cyclic units existed in the Ultramafic Series, their Mg isotopes in each layer appear independent and do not retain memory from the underlying cyclic unit.

Only one sample (16SW-3–10) derived from the Basal Series has $\delta^{26}\text{Mg}$ values of -0.25‰ (Table 1), making it unique, as it shares similar Mg isotopic compositions with the mean mafic Earth value ($\delta^{26}\text{Mg} = -0.25 \pm 0.07\text{‰}$; Teng et al. 2010) and komatiites ($\delta^{26}\text{Mg} = -0.25 \pm 0.04\text{‰}$; Dauphas et al. 2010).

Discussion

Mg isotopic evidence for Cotectic crystallization of chromite and silicates

While the composition of the parental magma of the Stillwater Complex is debated, a classic interpretation is that two distinct parental magmas derived from a primary komatiitic melt. A higher-MgO mafic norite (~13 wt% MgO), linked to the Ultramafic Series, is thought to result from contamination with upper crust, whereas a lower-MgO gabbronorite (~6–9 wt% MgO), parental to the Banded Series, is attributed to contamination with lower crust (Barnes et al. 2020; Jenkins et al. 2022). Crucially for this study, the Mg isotopic composition of these parental magmas is expected to remain largely unaffected by these different contamination pathways. This is because the upper crust contributes negligible Mg mass due to its low MgO content (~2.5 wt%; Rudnick and Fountain 1995; Gao et al. 1998), while the lower crust possesses a mantle-like Mg isotopic signature (Yang et al. 2016). The $\delta^{26}\text{Mg}$ values can therefore effectively trace the signature of the primary mantle source through these crustal processes. Our sample (16SW-3–10; $\delta^{26}\text{Mg} = -0.25 \pm 0.04\text{‰}$) from the Basal Series, identified as a chilled margin representing a primitive magma of the complex (Helz 1985; Premo et al. 1990), exhibits Mg isotopic compositions consistent with those of komatiites ($\delta^{26}\text{Mg} = -0.25 \pm 0.06\text{‰}$; Dauphas et al. 2010). In contrast, samples in the Banded Series, which represent the remnants of residual melts that underwent magmatic differentiation, show notably lighter Mg isotopic compositions (varying from -0.32‰ to -0.25‰ , with an average $\delta^{26}\text{Mg}$ of -0.29‰). Therefore, the magmatic differentiation process from komatiitic melts to mafic melts likely contributes to the variations in Mg isotopes observed in the Stillwater Complex.

Recent high-precision studies indicate that silicate crystallization can induce minor but resolvable Mg isotope fractionation ($\Delta^{26}\text{Mg}_{\text{Ol-melt}} \approx -0.10\text{‰}$; Wang et al. 2024), challenging the traditional view that this process leaves melts with a mantle-like $\delta^{26}\text{Mg}$ (Teng et al. 2010). However, in complex systems like layered intrusions, this primary magmatic signal is often obscured by post-cumulus processes and sub-solidus re-equilibration, which can create large, disequilibrium isotopic effects (Liu et al. 2025).

Therefore, while silicate fractionation may provide a background contribution, the crystallization of oxides (e.g., chromite, ilmenite, and titanomagnetite) remains the most accepted and dominant mechanism for producing significant Mg isotopic variations in evolved magmas (Su et al. 2019, 2021; Wang et al. 2021; Xiao et al. 2023). The rare presence of magnetite and ilmenite in the Banded Series of the Stillwater Complex is unlikely to significantly affect the Mg budget of the melts (Maze and Carlson 1982), consistent with the Mg isotopic homogeneity observed in samples from the Upper to Lower Banded Series. Therefore, the isotopically heavy chromite should be considered a key factor leading to the Mg isotopic deviations in the Banded Series.

Fractional crystallization models and Mg isotope fractionation

Two quantitative fractional crystallization models are presented here to explain how isotopically heavy chromite drives residual melts towards enrichment in light Mg isotopes, thereby reproducing the Mg isotopic variations observed in the Banded Series.

Model I: melts saturated solely in chromite

The equation governing Mg isotopic fractionation in a Rayleigh distillation process is:

$$\delta^{26}\text{Mg}_{\text{melt}} = (\delta^{26}\text{Mg}_0 + 1000) f^{(\alpha-1)} - 1000$$

where the $\delta^{26}\text{Mg}_0$ is the initial Mg isotope of the parental melt; in the case of the Stillwater, this value is consistent with the average Mg isotopes of komatiites (Dauphas et al. 2010). The fraction of Mg remaining in the melt is defined as $f = F \times C_{\text{melt}} / C_0$, where C_{melt} and C_0 represent the MgO concentration in the residual melts and the initial melts, respectively (Table S3), and F is the fraction of residual melt. The composition of the parental melts is taken from Jenkins and Mungall (2018), with an initial MgO concentration of 15.4 wt%. Considering that the majority of chromite grains are hosted in chromitite samples and the significant subsolidus diffusion in silicate cumulates affecting chromite initial compositions, we use the highest MgO values (11.9 wt%; 16SW-1–23) of chromite from the massive chromitite and lowest MgO (7.70 wt%; 16SW-2–4) of chromite from the disseminated chromitite for this model, respectively (Table S2). The α is the fractionation factor between chromite and melts, which is calculated by.

$$\Delta^{26}\text{Mg}_{\text{Chr-melt}} = \delta^{26}\text{Mg}_{\text{Chr}} - \delta^{26}\text{Mg}_{\text{melt}} \approx 10^3 \ln \alpha_{\text{Chr-melt}}$$

The equilibrated Mg isotopes of chromite ($\delta^{26}\text{Mg}_{\text{Chr}}$) can be theoretically calculated based on the ratios of octahedrally coordinated ions, i.e. Al: Cr: Fe^{3+} in spinel (Schauble 2011; Wang et al. 2023), along with an estimated temperature. The magnitude of inter-mineral fractionation varies from 0.5‰ for pure spinel (MgAl_2O_4) to 0.15‰ for magnesiochromite (MgCr_2O_4), to 0.08‰ for magnesioferrite (MgFe_2O_4) at 1200 °C; from 0.6‰ for pure spinel (MgAl_2O_4) to 0.2‰ for magnesiochromite (MgCr_2O_4), to 0.1‰ for magnesioferrite (MgFe_2O_4) at 1000 °C, and from 0.85‰ for pure spinel (MgAl_2O_4) to 0.25‰ for magnesiochromite (MgCr_2O_4), to 0.15‰ for magnesioferrite (MgFe_2O_4) at 800 °C, respectively (Young et al. 2009; Schauble 2011; Liu et al. 2011; Wang et al. 2023). Contributions from these three end members, pure spinel, magnesiochromite and magnesioferrite, must be considered based on their relative proportions. Samples of two chromite grains from the Stillwater Complex have general formulae of $(\text{Fe}_{0.64}\text{Mg}_{0.37}\text{Al}_{0.69}\text{Cr}_{1.00}\text{Fe}_{0.27}\text{O}_4)$ and $(\text{Fe}_{0.44}\text{Mg}_{0.56}\text{Al}_{0.65}\text{Cr}_{1.22}\text{Fe}_{0.11}\text{O}_4)$. Average chromite-melt fractionation factors ($\Delta^{26}\text{Mg}_{\text{Chr-melt}} = \delta^{26}\text{Mg}_{\text{Chr}} - \delta^{26}\text{Mg}_{\text{melt}}$) thus range from 0.264 and 0.261‰ at 1200 °C,

to 0.347 and 0.346‰ at 1000 °C, and to 0.441 and 0.447‰ at 800 °C (Table S2).

As predicted by the model, the removal of chromite can indeed result in isotopically light melts. However, significant Mg isotopic fractionation is only observed when a substantial amount of chromite crystallization occurs. For instance, to reach a Mg isotopic value of melts as -0.29 ‰, which is the average Mg isotopes of silicate cumulates in the Banded Series, 17%, 15%, and 11% modal abundances of chromite are required at liquidus temperatures ranging from 1200 °C to 1000 °C and then to 800 °C, respectively (Fig. 3a and Table S3). If we consider variations in Mg contents of chromite, such as the decreasing MgO contents of chromite in the model, higher modal abundances of chromite would be needed to lower the melt Mg isotopic value to -0.29 ‰. Actually, it is unrealistic for komatiitic melts to undergo such extensive chromite crystallization, as the melt would not provide a sufficient abundance of Cr to crystallize such a high proportion of chromite (Table S3; Liang and Elthon 1990; Barnes and Brand 1999). In other words, even with a continuous supply of komatiitic melts to provide the

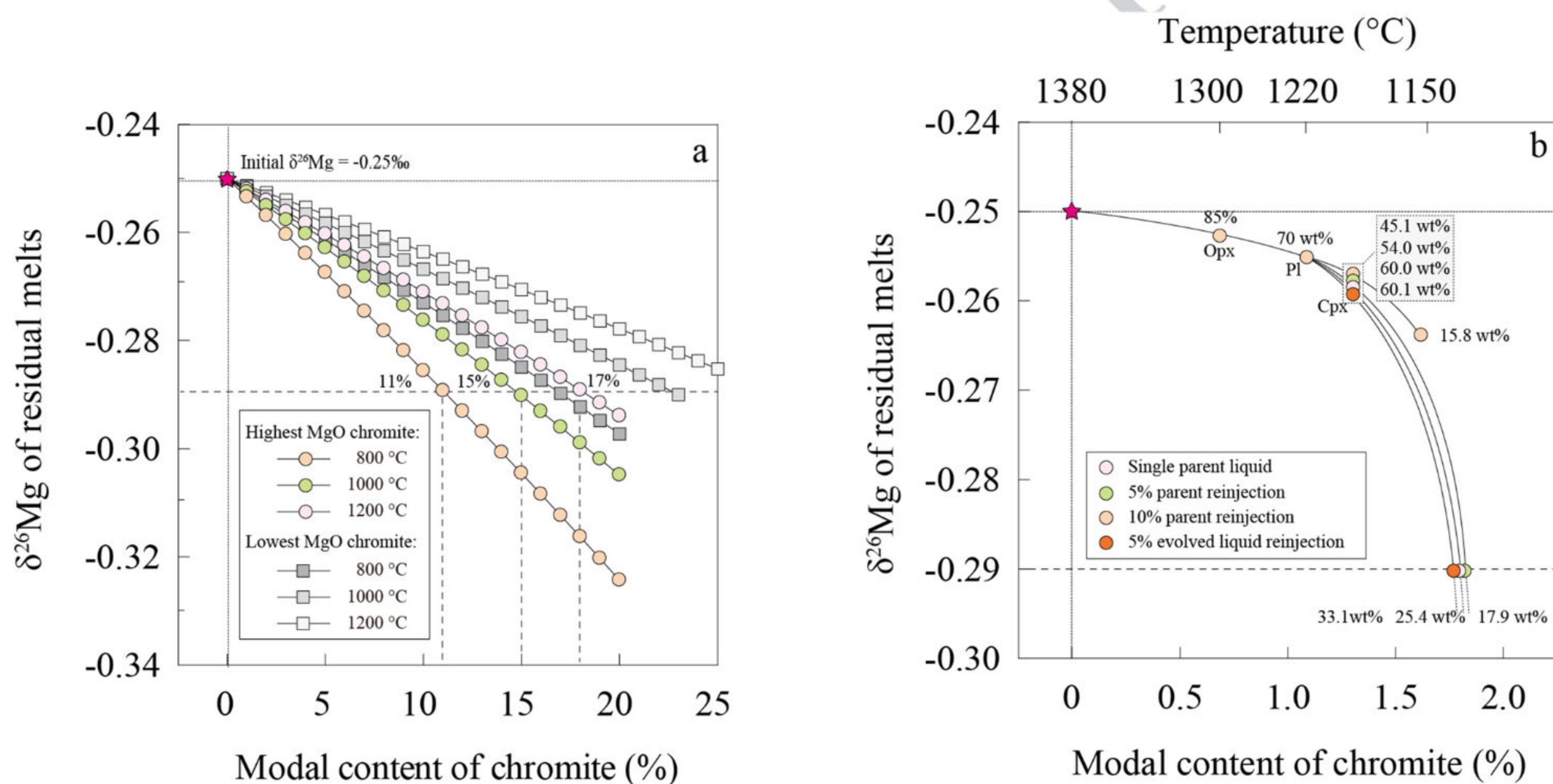


Fig. 3 (a) Model of Mg isotopic variations of melts that is only saturated in chromite as the liquidus phase. Solid lines represent calculated Mg isotopic compositions of residual melts during chromite fractional crystallization by assuming a Rayleigh distillation process at different temperatures. Dashed horizontal lines represent average Mg isotopic compositions in samples from the Banded Series of the Stillwater Complex. The colored stars represent the initial compositions of melts with the $\delta^{26}\text{Mg}$ values and MgO contents assumed as -0.25 ‰ (Dauphas et al. 2010) and 15.37wt% (Jenkins and Mungall 2018). The variable chromite–melt fractionation factors ($\Delta^{26}\text{Mg}_{\text{chromite-melt}} = \delta^{26}\text{Mg}_{\text{chromite}} - \delta^{26}\text{Mg}_{\text{melt}}$) at different temperature are showed in Table S2 and S3. (b) Model of Mg isotopic variations of melts that chromite and other minerals are cotectic in proportions. The mineral assemblages

and fractional crystallization process are simulated by Alpha-MELTS experiments designed by Benson and Boudreau (2023; see more details in the Table S4). Solid lines represent calculated Mg isotopic compositions of residual melts induced by chromite fractional crystallization at closed or open systems. The point represents the initial appearance of different minerals and the liquid proportions at that time. The colored stars represent the initial compositions of melts with the $\delta^{26}\text{Mg}$ values and MgO contents assumed as -0.25 ‰ (Dauphas et al. 2010) and 15.37wt% (Jenkins and Mungall 2018). To simplify the calculations, we only utilize the chromite–melt fractionation factor at 1200 °C ($\Delta^{26}\text{Mg}_{\text{chromite-melt}} = \delta^{26}\text{Mg}_{\text{chromite}} - \delta^{26}\text{Mg}_{\text{melt}}$) at different temperature are shown in Table S2 and S3

necessary Cr for chromite crystallization, the Mg isotopes of this additional melt would not show obvious variations.

Model II: chromite reaches a Cotectic line with other silicate minerals

Another scenario to consider is that the onset of chromite crystallization is cotectic with other silicate minerals with high Mg concentrations. This results in steeply decreased Mg contents in the residual melts with a more pronounced deviation in Mg isotopes due to chromite crystallization. This model is based on the Alpha-MELTS experiments designed by Benson and Boudreau (2023) for the Stillwater Complex, which track the fractional crystallization history and the proportions of segregated minerals. Our calculations utilize a pressure of 2.5 kbar and quartz-fayalite-magnetite (QFM) oxygen fugacity, along with starting compositions suggested by Jenkins and Mungall (2018) as the primitive magma of the Stillwater Complex.

The parental magma yields a liquidus temperature of 1420 °C and is fully solidified at 1100 °C (with over 99% solidification). To simplify the calculations, we set a 10 °C temperature interval to calculate the Mg isotopic deviations, using the following equation at each step:

$$\delta^{26} \text{Mg}_{\text{melt}, n+1} = f' (\delta^{26} \text{Mg}_{\text{melt}, n})$$

The fraction of MgO remaining in the melt is given by $f' = (F \times C_{n+1}) / C_n$, where C_{n+1} and C_n represent the MgO concentrations in the melts before and after each crystallization step. The fraction of remaining melt F is calculated from the MELTS model. In addition to calculating conventional Rayleigh fractionation in a closed system, we also incorporate calculations considering open-system, in which the parent melt mixes 5% and 10% reinjection of fresh pulses of primitive magmas, as well as 5% evolved liquids, at every 10% interval. Model results, including chromite modal abundances, residual melt compositions, and $\delta^{26}\text{Mg}$ values at each crystallization step, are reported in the Table S4.

The Rayleigh fractionation model based on the Alpha-MELTS calculation shows a more reasonable scenario. In both closed and open systems, chromite and olivine are the cotectic phases on the liquidus before temperature decreases to 1380 °C. Orthopyroxene and plagioclase join the crystallized assemblages at about 1380 °C and 1220 °C, respectively, corresponding to decreased liquid proportions of about 70% and melt MgO content of about 7.50 wt%, as well as 1.09% modal content of crystallized chromite. At these two stages, the segregation of chromite from the liquid has a limited effect on the Mg isotopic compositions of the residual liquid, with $\delta^{26}\text{Mg}$ values decreasing only from -0.250‰ to about -0.255‰ (Fig. 3b). As the temperature

continues to decrease to 1150 °C, the crystallization of clinopyroxene further decreases the melt MgO content. Although the chromite modal content increases from 1.09% to about 1.72% in this interval, this can significantly influence Mg isotopic variations in residual melts (Fig. 3b).

Furthermore, the addition of an evolved liquid can deplete the Mg budget of the melts compared to a closed system, leading to a more significant decrease in melt Mg isotopes. Conversely, the injection of fresh pulses of primitive magma moderates the decline in melt Mg contents and delays the isotopically light trend. Therefore, to reach similar Mg isotopic deviations, necessitates a melt with a lower fraction and separation of a higher proportion of chromite. If more primitive magmas are mixed, such as incorporating 10% into the initial melt in our model, the melt Mg isotopes would retain relatively constant even after all the Cr budget is exhausted.

Significant Mg isotope fractionation prefer to Cotectic model

Although multiple mechanisms can drive the magma into the chromite-only field in stratiform chromitite, our fractional crystallization model demonstrates that significant Mg isotopic fractionation can only be expected with an extremely high proportion of chromite removal. The predicted chromite proportion of less than 2% from the chromite-silicate cotectic model results in Mg isotopes deviating from -0.25‰ to -0.29‰ (Fig. 3b), aligning more closely with theoretical predictions and petrological evidence for komatiitic melts (Barnes 1998; Barnes and Brand 1999). In particular, the Alpha-MELTS model used here simplifies the magmatic process. Low MgO backgrounds in the initial melt (such as the suggested MgO contents of 10.2 wt% and 13.0 wt% by Longhi et al. (1983) and Helz (1985), respectively) and the increasing Al in chromite (which raises the chromite-melt fractionation factor $\Delta^{26}\text{Mg}_{\text{Chr-melt}}$) can further facilitate Mg isotopic fractionation. Consequently, a higher proportion of residual melts or less chromite segregation might be expected in natural samples. More importantly, the Mg isotopic results support the assumption of the chromite-silicate cotectic model, indicating the mechanical sorting of chromite from crystal mushes into monomineralic layers, which corresponds to a low cotectic proportion of chromite relative to silicate minerals (Roeder et al. 2006; Eales and Costin 2012; Mungall et al. 2016).

On the other hand, it should be noted that light Mg isotopes are only expected to become pronounced following the crystallization of two types of pyroxenes. However, the measured Mg isotopic compositions for samples in the Ultramafic Series do not conform to the MELTS model due to the mechanical sorting. The poikilitic harzburgite samples

exhibit lighter $\delta^{26}\text{Mg}$ than komatiitic melts, while the granular harzburgite and olivine cumulate, in contrast, show isotopically heavier characteristics. Therefore, in addition to altering the initially crystallized mineral modal proportions, the chemical information of fractional crystallization would also be obscured by mineral sorting. In the following sections, we will consider the sorting mechanism in studies of Mg isotopes in samples of the Ultramafic Series.

Kinetic Sieving of chromite during fluid migration

The earliest proposed sorting mechanisms primarily focused on density (Eales 2000) or size (Mondal and Mathez 2007) distinctions between chromite and other silicate minerals. Small and dense chromite grains can segregate from the chromite-olivine-pyroxene slurries by percolating downward throughout interstitial spaces between cotectic silicate crystals, leading to the formation of a basal layer of chromite. Subsequently, researchers have gradually recognized that a kinetic sieving model may be more effective than the gravity settling model. In this context, spontaneous lateral transport allows chromite to migrate downward through interstitial spaces, resulting in the formation of monomineralic layers (Mungall et al. 2016; Jenkins and Mungall 2018). The dynamics of this process can be influenced by factors such as the forceful injection of slurries from deeper chamber (Eales 2000), the downward slumping of semi-consolidated cumulates (Maier et al. 2018; Barnes et al. 2023), and earthquakes accompanying magma injection (Elsworth and Voight 1995; Nex 2004). Although these mechanisms present different scenarios, they mainly emphasize physical processes (e.g. Roeder et al. 2006). In other words, the Mg isotopic variations in samples from the Ultramafic Series merely corresponds to the varying mixing proportions of chromite. The model presented here aims to replicate these Mg isotopic variations.

Mg isotopic fractionation induced by variable chromite mineral modal

Chromite grains serve as the representative oxide with heavier $\delta^{26}\text{Mg}$ (Bai et al. 2021; Su et al. 2021; Xiao et al. 2023). Assuming that variable chromite modal content controls the Mg isotopic compositions for whole rocks, the theoretical $\delta^{26}\text{Mg}$ value can be calculated using the following equation:

$$\delta^{26}\text{Mg}_{\text{rocks}} = \delta^{26}\text{Mg}_{\text{chromite}} \times \text{achromite} \times \text{MgO}_{\text{chromite}}/\text{MgO}_{\text{rocks}} + \delta^{26}\text{Mg}_{\text{silicates}} \times (1 - \text{achromite} \times \text{MgO}_{\text{chromite}}/\text{MgO}_{\text{rocks}})$$

where $\delta^{26}\text{Mg}_{\text{rocks}}$ is the calculated Mg isotope composition of the sample and α_{chromite} represents the normalized chromite modal content (Table S5). The $\delta^{26}\text{Mg}_{\text{chromite}}$ and $\delta^{26}\text{Mg}_{\text{silicates}}$ represent the Mg isotopic compositions of chromite and the sum of other Mg-bearing silicate cumulates, respectively. The $\delta^{26}\text{Mg}_{\text{chromite}}$ has been theoretically calculated in *Model I* of Sect. 5.1.1, and $\delta^{26}\text{Mg}_{\text{silicates}}$ is assumed to be equal to that of mean komatiites (Dauphas et al. 2010). The calculated trend of Mg isotopes for whole rocks driven by variations in chromite modal content is shown in Table S6.

The model results indicate that a declining chromite modal content correlates with an isotopically light whole-rock compositions (Fig. 4a). Measured $\delta^{26}\text{Mg}$ values of samples derived from the B and G cyclic units exhibit similar trend compared to model, although they generally show lighter $\delta^{26}\text{Mg}$ in poikilitic harzburgite and chromitite samples, excluding isotopically heavy outliers in the B cyclic unit induced by alteration. For olivine cumulate and granular harzburgite with minimal chromite in the B and A cyclic units, the Mg isotopes for whole rocks are heavier than model results. In brief, the deviation between measured and model Mg isotopic results suggests that physical factors alone cannot fully explain the observed variations. While analogous experiments have validated physical sorting processes (Forien et al. 2015), the precise mechanisms are complex and debated. For instance, Forien et al. (2015) used the term “viscous particle sorting” specifically to include processes beyond simple kinetic sieving. This complexity arises because chromite crystals may not act as independent particles (Glazner 2014); they can be bound by viscous forces and form chains or clusters through particle interactions (Hammer et al. 2010). Therefore, applying a simple kinetic sieving model to stratiform chromitites is likely an oversimplification and introduces uncertainty.

Mg isotopic fractionation induced by fluid migration

A key model for the physical sorting of chromite involves the role of an exsolved volatile phase (Tang et al. 2023; Bai et al. 2024). Specifically, the model proposed by Bai et al. (2024) suggests that as magma ascends and is emplaced into a shallow crustal chamber, the rapid decompression causes an H_2O -rich volatile phase to exsolve from the melt, forming bubbles. These low-density bubbles are thought to preferentially attach to chromite microlites, creating buoyant chromite-bubble aggregates. This flotation mechanism can efficiently separate chromite from the surrounding silicate minerals, leading to the formation of nearly monomineralic chromitite layers. The redistribution of this H_2O -rich fluid phase could contribute to the isotopic heterogeneity between chromitite and silicate cumulates in each cyclic

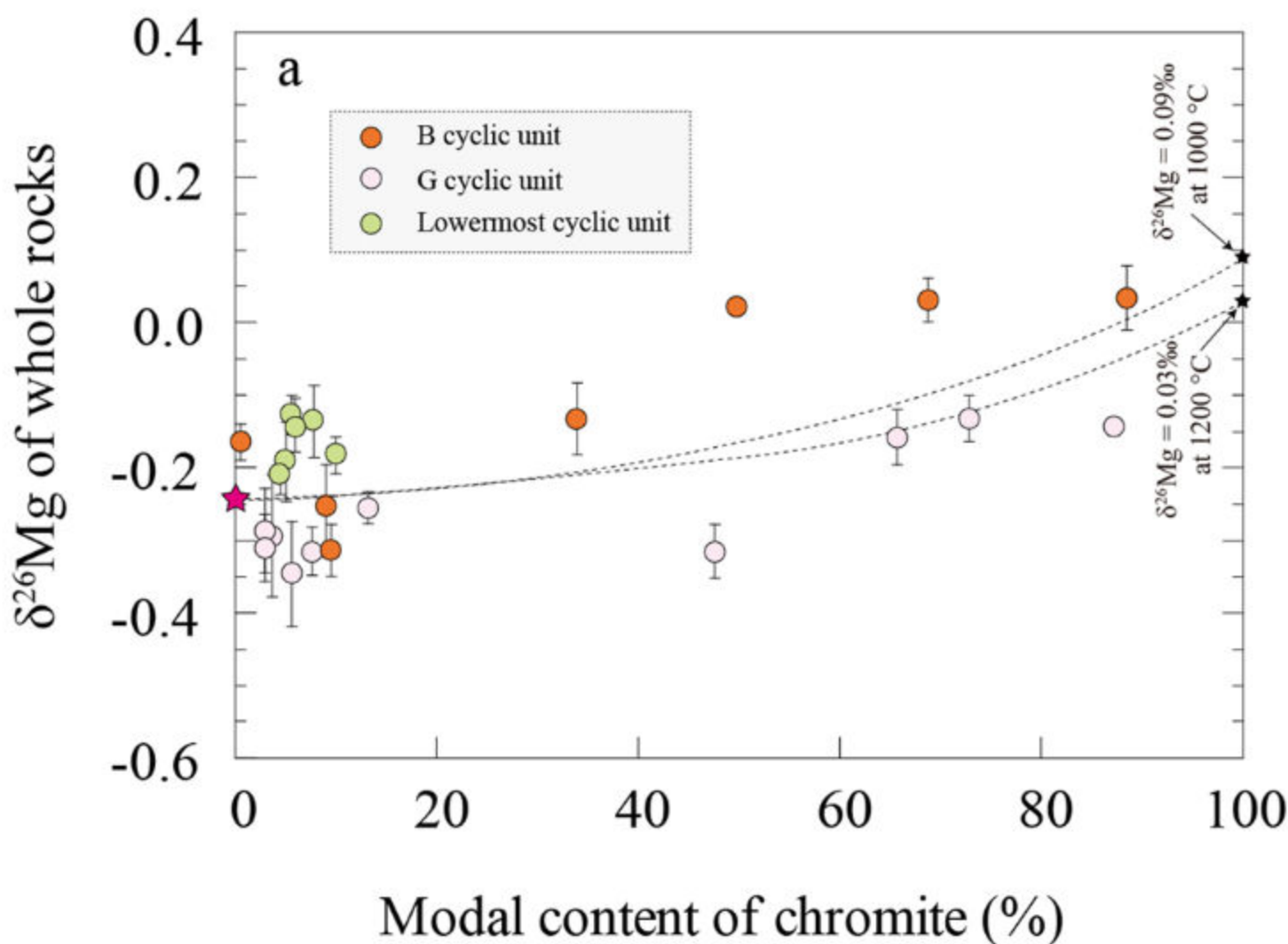
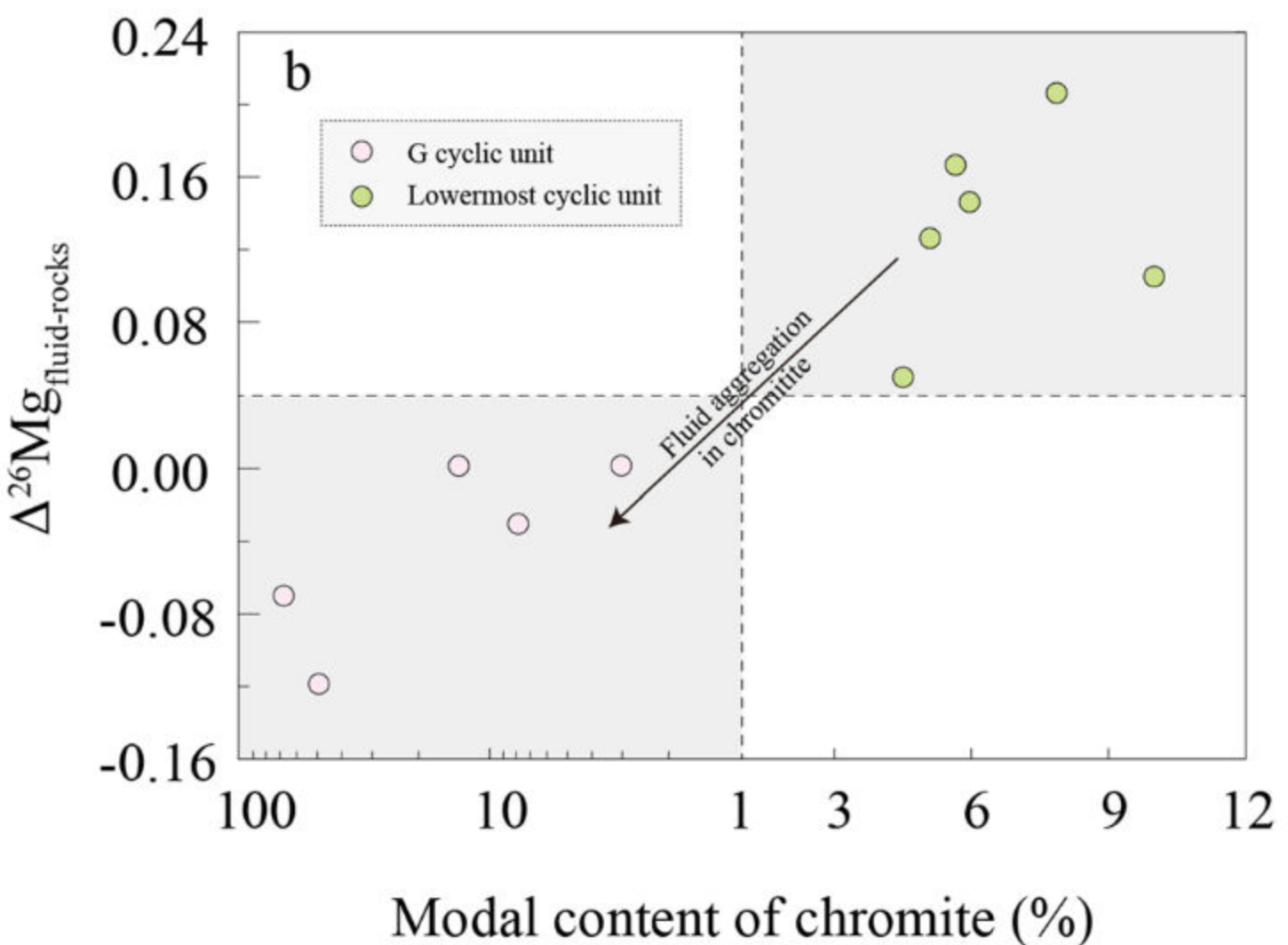


Fig. 4 (a) Correlation diagrams of Mg isotopes for whole rocks and chromite modal content from the Ultramafic Series. Dashed lines represent calculated Mg isotopic compositions induced by variable chromite modal content at different temperatures. Only the chromite–melt fractionation factor at 1000 °C and 1200 °C are used in this model. The colored stars represent the initial compositions of samples without chromite and the $\delta^{26}\text{Mg}$ value assumed as -0.25‰ (Dauphas et al. 2010). Normalized modal contents of chromite are estimated using least-squares linear regression from the major element composition of the cumulus phases and whole rocks (Table S6). Such algorithms were designed by Stormer and Nicholls (1978) and applied in studies for other layered intrusions (e.g., Namur et al. 2010). Sums of squares of residuals ($\sum r^2$) are usually below than 2.0. For seven samples where mineral analyses are not available (such as the highly altered silicate

unit (e.g., Bai et al. 2021; Su et al. 2020, 2021). This Mg isotopic heterogeneity can be anticipated in the corresponding layers, with lighter Mg isotopes preferentially partitioning into the H_2O -rich fluid. This would result in a lighter bulk $\delta^{26}\text{Mg}$ signature in the chromitite layers where the fluid phase is concentrated, while the residual silicate cumulates would exhibit heavier Mg isotopes. However, effective means to directly analyze the isotopic compositions of fluids are still lacking. The $\delta^{26}\text{Mg}$ values of chromite, olivine, and orthopyroxene reported in previous study have demonstrated isotopic fractionation caused by the re-equilibration with trapped fluids and inter-mineral subsolidus diffusion (Bai et al. 2021). Thus, probing the Mg isotopic discrepancies recorded in whole rocks and mineral aggregates offers another approach. The deviation of Mg isotopes caused by fluids for Mg-rich minerals can be expressed by the following formula:

$$\delta^{26}\text{Mg}_{\text{fluid} - \text{mineral}} = \delta^{26}\text{Mg}_{\text{measured}} - \delta^{26}\text{Mg}_{\text{diffusion}} - \delta^{26}\text{Mg}_{\text{initial}}$$

The Mg isotopic variations in minerals caused by inter-mineral subsolidus diffusion can be offset by calculating the weighted average of the Mg isotopes in all Mg-rich minerals.



minerals in B cyclic unit), the mineral chemistry from same rock types in G cyclic unit is used because of their homogeneous and limited compositions. In addition, point counting are carried out on their polished thin-sections. At least 1000 points are counted for each sample on a grid of $0.4\text{ mm} \times 0.4\text{ mm}$. The data obtained in volume percent are converted to weight% using the average density of minerals. All of the samples in B cyclic unit are determined by point counting, and give consistent results with the calculations from least-squares linear regression. **(b)** Calculated Mg isotopic variations for whole rocks induced by fluid migration versus the chromite modal content in samples from the G cyclic unit and the A cyclic unit. The arrows represent the direction of fluid migration from olivine cumulate and granular harzburgite to chromitite and poikilitic harzburgite

The theoretical $\delta^{26}\text{Mg}$ values induced by fluid migration for whole rocks can be represented by the following equation:

$$\delta^{26}\text{Mg}_{\text{fluid-rocks}} = \delta^{26}\text{Mg}_{\text{measured-rocks}} - \Delta^{26}\text{Mg}_{\text{phase 1}} \times \alpha_{\text{phase 1}} \times \text{MgO}_{\text{phase 1}} / \text{MgO}_{\text{rocks}} - \Delta^{26}\text{Mg}_{\text{phase 2}} \times \alpha_{\text{phase 2}} \times \text{MgO}_{\text{phase 2}} / \text{MgO}_{\text{rocks}} - \dots - \Delta^{26}\text{Mg}_{\text{phase X}} \times \alpha_{\text{phase X}} \times \text{MgO}_{\text{phase X}} / \text{MgO}_{\text{rocks}}$$

The $\alpha_{\text{phase X}}$ represents the normalized mineral modal contents of chromite or silicate minerals within the whole rocks, and $\Delta^{26}\text{Mg}_{\text{phase X}}$ is the difference between measured and initial Mg isotopes. To simplify calculations, we assume that the Mg isotopes of clinopyroxene are similar to those of orthopyroxene. Additionally, for silicate minerals (olivine, clinopyroxene, and orthopyroxene), we consider their initial Mg isotopes to be -0.25‰ , ignoring the Mg isotopic variations induced by fractional crystallization (see the Rayleigh fractionation *Model I* in Sect. 5.1.1). The initial Mg isotopes of chromite are also cited from Section of 5.1.1. Representative samples are selected for calculations, and the results are shown in Table S7 and plotted in Fig. 4b.

The calculations based on the fluid migration model show that the weighted $\delta^{26}\text{Mg}$ values of Mg-bearing mineral assemblages in the chromitite and poikilitic harzburgite

from the G cyclic unit are commonly lighter than the measured Mg isotopes for whole rocks, with $\Delta^{26}\text{Mg}$ ranging from 0.00‰ to −0.12‰ (Fig. 4b). Increased chromite modal content enhances the degree of fluid aggregation, exhibiting a negative correlation with $\Delta^{26}\text{Mg}$ values. Conversely, the weighted $\delta^{26}\text{Mg}$ values of Mg-bearing mineral assemblages in olivine cumulate and granular harzburgite from the A cyclic unit are typically heavier than the measured Mg isotopes for whole rocks, with $\Delta^{26}\text{Mg}$ values ranging from 0.05‰ to 0.21‰. Therefore, extraction of isotopically light fluids by chromite accumulation, leaving heavier partner in residual melts, may contribute to the Mg isotopic fractionation in samples of the Ultramafic Series. This result also supports the kinetic sieving model related to the aggregation between fluids and chromite (Bai et al. 2024), reflecting the process of chromite separation from the slurries.

Genesis of the stratiform chromitite

Based on the discussion above, we propose a model that combines the cotectic model and mechanical sorting processes to summarize the formation of chromite layer in the Stillwater Complex:

Partial melting in the mantle led to the ascent of komatiitic melts through conduits in the crust. As the primitive magma ascended, it assimilated crustal rocks and cooled. We argue that the dominating chromite and olivine as well as some orthopyroxene and plagioclase are precipitated before reaching the shallow-level chamber. This is because the chromite cumulation load in each batch had no memory in Mg isotopes from previous cycles. The trans-crustal melt conduits likely provide both lithological materials and sufficient time for wholesale rock assimilation (e.g., Cashman et al. 2017). Mungall et al. (2016) suggested that highly contaminated batches of komatiite magma could carry up to 20% suspended crystals. Yao et al. (2021) proposed that contaminated magma comprises approximately one-third ultramafic solids and two-thirds liquids. At this stage, the residual melts have reached a relatively evolved state with low MgO contents and light Mg isotopes. The crystal-rich slurries are subsequently transported by the low-viscosity melt up the conduit (e.g., Eales and Costin 2012). From this perspective, our results align with the current “crystal mush model” (e.g., Yang et al. 2019; Yao et al. 2021; Xing et al. 2022), which posits that layered intrusions, particularly the cyclic units in the Ultramafic Series, form through multiple injections and emplacements of crystal mushes.

In contrast, heavily contaminated komatiitic melts and high degrees of evolution contribute to the aggregation and immiscibility of fluids, which have a stronger attraction for chromite (Barnes 1998; Barnes and Brand 1999). Concurrently, fluids exsolving from melts with low $\delta^{26}\text{Mg}$ values

account for the isotopic fractionation observed between silicate cumulates and chromitite (Fig. 4b). In this dynamic setting, the continuous intrusion of new magma spreads laterally, resulting in local oscillating conditions and turbulence. Chromite is segregated from coexisting silicate crystals through kinetic sieving and accumulates at the base of the slurry while still flowing along the sill floor (Maier et al. 2018; Forien et al. 2015; Mungall et al. 2016), or it may even be floated by these fluids, causing it to migrate away from the silicate minerals in the magma channel (review in Bai et al. 2024). Therefore, the stratiform chromitite forms from crystal-rich mushes through the separation of chromite from coexisting silicate minerals. Fluid migration plays a critical role as a medium for these mechanical sorting processes.

Conclusions

We presented high-precision Mg isotopic analyses for whole rocks from the Stillwater Complex, one of the oldest and largest known layered intrusions. From our results, we draw the following conclusions:

- 1) The samples in the Banded Series exhibit limited $\delta^{26}\text{Mg}$ variations and light Mg isotopic compositions ranging from $-0.32 \pm 0.04\text{‰}$ to $-0.25 \pm 0.07\text{‰}$, with an average $\delta^{26}\text{Mg}$ of $-0.29 \pm 0.05\text{‰}$ ($n=10$). Cotectic chromite-silicates can rapidly deplete the Mg content in residual melts, resulting in significant Mg isotopic fractionation.
- 2) Chromitite and poikilitic harzburgite samples within the Ultramafic Series display isotopically light signatures, varying from $-0.35 \pm 0.07\text{‰}$ to $+0.03 \pm 0.04\text{‰}$. In contrast, the olivine cumulate and granular harzburgite samples show heavier Mg isotopic compositions from $-0.21 \pm 0.03\text{‰}$ to $-0.13 \pm 0.03\text{‰}$. We suggest that the isotopically lighter fluids, which accumulate on the surfaces of chromite, lead to isotopically heavier melts, resulting in the preferential partitioning of these heavier Mg isotopes into silicate cumulates.

Appendix 1 Supplementary material

Supplementary information for this article includes: **Table S1** Whole-rock major element compositions of samples from the Banded Series of the Stillwater Complex; **Table S2** Representative composition and chemical formula of chromite in the Ultramafic Series of the Stillwater Complex as well as the theoretically equilibrated $\delta^{26}\text{Mg}$ values with komatiitic magmas at variable temperature; **Table S3** Calculated Mg isotopic compositions of residual melts during the fractional crystallization of variable chromite modal contents by assuming a Rayleigh distillation process at different temperatures; **Table S4** Alpha-MELTS experimental

results and calculated Mg isotopic compositions of residual melts; **Table S5** Normalized mineral modal and the major element ratios of whole rock and mineral of samples from the Ultramafic Series of the Stillwater Complex; **Table S6** Calculated Mg isotopic compositions induced by variable chromite modal content at different temperatures; **Table S7** Calculated Mg isotopic variations in whole rock induced by fluid migration. **Figure S1** Geologic map of the Stillwater Complex and sample locations; **Figure S2** Photomicrographs under cross-polarized light of representative samples in the Stillwater Complex; **Figure S3** Major elements and MgO/(MgO + FeO_t) for whole rocks in the stratigraphic section of the Stillwater Complex. Supplementary material to this article can be found online at <https://doi.org/10.17632/7rmc2wgfwc.2>.

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Author contributions Yang Bai and Ben-Xun Su conceived and designed the study. Yang Bai performed the experiments, analyzed the data, and wrote the first draft of the manuscript. Yan Xiao and Yong-Sheng He contributed to the sample preparation and data collection. Yishen Zhang and Bernard Charlier provided critical resources and feedback. The entire research was supervised by Ben-Xun Su and Bernard Charlier. All authors contributed to the discussion and manuscript revision, and have read and approved the final submitted version.

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Data availability Data are available through Mendeley Data at <https://doi.org/10.17632/7rmc2wgfwc.2>.

Declarations

Ethics approval and consent to participate Not applicable.

Consent for publication Not applicable.

Competing interests The authors declare that they have no competing interests.

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