



Boron isotopes in Archean–Proterozoic marine deposits trace continental emergence

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Supplementary Information

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1. Rationale, sampling and methods

1.1 Research concept:

Boron isotopes are a powerful tracer of a large variety of geological processes (e.g. Foster *et al.*, 2018; Marschall and Foster, 2018 and references therein). As an incompatible lithophile element, boron is readily concentrated in the continental crust and depleted in the mantle (Marschall *et al.*, 2017). Processes of erosion, dissolution and deposition in the terrestrial and the marine environment fractionate boron isotopes and consequently drive modern seawater towards a high $\delta^{11}\text{B}$ value over time (modern seawater $\delta^{11}\text{B} = +39.6\text{‰}$; Foster *et al.*, 2010). Yet, the B isotopic composition of the ocean varied over geologic time (Joachimski *et al.*, 2005; Kasemann *et al.*, 2005, 2009; Lemarchand *et al.*, 2000; Legett *et al.*, 2020; Jurikova *et al.*, 2025). These long-term changes in oceanic B concentration and $\delta^{11}\text{B}_{\text{sw}}$ are controlled by the balance between input and output fluxes of boron to and from the ocean, which are relatively well characterized for the modern Earth (Marschall, 2018; Lemarchand *et al.*, 2002; Joachimski *et al.*, 2005; Simon *et al.*, 2006; Legett *et al.*, 2020). The largest in-flux of boron into the modern ocean is through continental erosion, together with smaller fluxes of hydrothermal venting and fluid expelled from accretionary prisms at convergent margins (Lemarchand *et al.*, 2002). The most dominant boron sink from the ocean is low-temperature alteration of oceanic crust (AOC), with additional

sinks of boron adsorption on oceanic clastic sediments and clays, as well as coprecipitation of boron in carbonate rocks (Fig. 2a; Lemarchand *et al.*, 2000; Simon *et al.*, 2006). For the modern setting, the isotopic composition, isotopic offset values (Δ ; see below), and flux magnitude of each of the boron sources and sinks to and from the ocean are relatively well constrained (Fig. 2a).

On the Archean Earth, in the absence of emerged continents, many of the boron fluxes to and from the oceans were not operating, and available constraints on the fluxes that likely operated are either poor or absent (Fig. 2b). The only large reservoir which certainly interacted with the ocean water at that time was the oceanic crust. Thus, the only major operating flux of boron in the absence of emerged continents is the removal of boron from the ocean by AOC. Therefore, prior to the emergence of continents, the Archean ocean likely reached an equilibrium with the oceanic crust (or mantle rocks exposed to the seafloor) and was low in its boron concentration, as the sources of boron were small to negligible. From this starting point, the emergence of continental crust will have ensued the operation of additional fluxes of B into and out of the ocean through the development of terrestrial erosion systems and rivers. Over time, the enrichment of ocean water with nutrients from terrestrial erosion also enabled the evolution and flourishing of biomineralizing marine microorganisms and the development of boron sinks out of the ocean through co-precipitation during carbonate deposition. Terrestrial erosion is also responsible for the introduction of clay material into the ocean by riverine and aeolian transport. These clay particles remove boron from the ocean water via adsorption as they sink through the water column. Thus, the emergence of continental crust above sea-level, together with the ensuing development of Earth's tectonic and physical and geochemical mechanisms, eventually drove the oceanic B cycle closer to its modern state. Chemical marine deposits, such as cherts, IFs and shales preserved from ages across the suggested continental emergence transition (3.5–2.5 Ga; Fig. S-1) will record this change in $\delta^{11}\text{B}_{\text{sw}}$, in the same way they record Archean oceanic chemistry (Bekker *et al.*, 2010, 2014; Konhauser *et al.*, 2017); accordingly, we targeted such rock units in the present study.

For the evaluation of the oceanic composition, any measured $\delta^{11}\text{B}$ value from a given marine deposit needs to be translated to the contemporaneous oceanic value ($\delta^{11}\text{B}_{\text{sw}}$) by the application of a fractionation factor (α ; expressed in equilibrium as $1000\ln\alpha \approx \Delta^{11}\text{B}_{\text{a-b}} = \delta^{11}\text{B}_{\text{phase a}} - \delta^{11}\text{B}_{\text{phase b}}$). Boron isotope fractionation between seawater and silica and/or clays have been constrained both empirically (Spivack *et al.*, 1987; Kolodny and Chaussidon, 2004) as well as by experimental work (Saldi *et al.*, 2021; Ring *et al.*, 2025), usually testing the effects of pH on fractionation. The effect of variation in temperature on boron isotope fractionation is also relatively well understood from experiments and ab-initio modelling (Kowalski and Wunder, 2018). The major ion composition and ionic strength of the seawater solution also have the potential to affect fractionation. While it may seem that all these variables, as well as the possible post depositional alteration of the boron signal (see below) pose difficulty for the reconstruction of $\delta^{11}\text{B}_{\text{sw}}$ from marine deposits, a critical evaluation of each of these elements can allow us to assign fractionation factors and evaluate past $\delta^{11}\text{B}_{\text{sw}}$ with meaningful confidence, given the available constraints on the pH and T of Archean–Proterozoic seawater (Krissansen-Totton *et al.*, 2018).

In silicate materials, the magnitude of boron isotope fractionation increases at lower temperatures (Kowalski and Wunder, 2018), and while an increase in SW temperature may reduce the fractionation factor, a 10 ‰ shift in $\Delta^{11}\text{B}$ would require a temperature increase of ca. 100 °C. Modelling envelopes of Archean seawater temperature range between 0–45 °C (Krissansen-Totton *et al.*, 2018), while empirical proxies suggest a range between 32–65 °C (Hren *et al.*, 2009). The suggested model mean predicts fluctuations of no more than 10 °C in seawater temperatures across all of Earth history (Krissansen-Totton *et al.*, 2018). These considerations render the plausibility that large variations in reconstructed $\delta^{11}\text{B}_{\text{sw}}$ come from temperature fluctuations very low, as these would realistically result in relatively minor changes.

The effect of pH on fractionation is the one best tested experimentally and well evaluated both for SW–silica (Saldi *et al.*, 2021) and SW–clay (Spivack *et al.*, 1987; Ring *et al.*, 2025). Both materials show a large deviation and pH dependence of the fractionation factor between pH = 7.5–8.5, at which boron speciation is approximately equally divided between boric acid ($\text{B}(\text{OH})_3$) and borate ion ($\text{B}(\text{OH})_4^-$) and mineral–seawater fractionation is considered as a combination of boric acid/mineral and borate/mineral fractionations. In contrast, under acidic to circumneutral conditions, boron speciation is vastly dominated by boric acid and there is thus little to no variation of the fractionation factor with pH. Hence at low pH, boron isotope fractionation can be expressed through a single mineral/bulk seawater isotope fractionation factor. Considering the modelled pH of Archean–Proterozoic seawater is not predicted to reach 7.5 until the Phanerozoic (Krissansen-Totton *et al.*, 2018), changes in the recorded $\delta^{11}\text{B}_{\text{sw}}$ due to pH fluctuations are unlikely, and the fractionation factor is expected to remain constant. Moreover and importantly, the SW–clay fractionation factor (Spivack *et al.*, 1987) and the SW–silica fractionation factor (Saldi *et al.*, 2021), evaluated at pH values modelled for Archean seawater (Krissansen-Totton *et al.*, 2018), converge to the same value of $\Delta^{11}\text{B} = -25$ ‰. This allows us to use a single value for the reconstruction of $\delta^{11}\text{B}_{\text{sw}}$ from both archives.

The ion composition and salinity of Archean seawater are less firmly constrained (Albarede *et al.*, 2020). It has been shown though that the strong variation in the SW–clay fractionation factor between pH = 7.5–8.5 is largely removed in experiments conducted with pure water, and that B fractionation and adsorption on clay surfaces is mostly modulated by fluid composition (Ring *et al.*, 2025). Therefore, if Archean seawater was less saline than today due to reduced continental weathering fluxes, this would reduce rather than amplify departures from the experimentally determined fractionation factors we assigned for the purpose of $\delta^{11}\text{B}_{\text{sw}}$ reconstruction.

While we acknowledge the inherent uncertainties when aiming to reconstruct seawater isotopic compositions from periods in Earth history for which little constraints exist, available theoretic and experimental data allow us to assign and implement the aforementioned fractionation factors to our data with reasonable confidence. Moreover, the magnitude of variation we observe in $\delta^{11}\text{B}_{\text{sw}}$ is such that is robust to any plausible variation in these parameters (T , pH, salinity), given the magnitude of the observed signal (see main text).

1.2 Sampling and analytical methods:

Rock samples of IFs, chert and marine shales, with constrained ages between 3.70–1.88 Ga were selected for analysis in the present study (Fig. 1; Table S-1). Sample material for analysis was carefully selected preferably from previously well characterized localities known for their IFs, and which their available age constrains place them in the period of interest for the present study. We attempted to include in our sample suite material that i) comes from multiple Archean cratons, ii) covers the time interval of interest as densely as possible, and iii) comprises units which experienced no metamorphic overprint or conditions of greenschist facies or below in order to minimize the possibility of post depositional and/or diagenetic alteration of its boron inventory. Where available, drill core material was preferred over surface samples to minimize the effects of surface and post-depositional alteration. We also benefited from the possibility to directly collect sample material from ICDP drill core projects BARB and BASE.

The assembled suite of IFs as well as cherts and marine shales associated with them includes material from the Isua belt in Greenland (Frei *et al.*, 1999; Polat and Frei, 2005; Nutman and Friend, 2009; Czaja *et al.*, 2013), the Pilbara and Yilgarn cratons in Australia (Chen *et al.*, 2003; Czaja *et al.*, 2018; Duuring *et al.*, 2020; Angerer *et al.*, 2022), the Abitibi Greenstone Belt in NE North America (Delvigne, 2012; Delvigne *et al.*, 2012), the Barberton Greenstone Belt (Hofmann, 2005; Bontognali *et al.*, 2013; Heubeck *et al.*, 2013; Satkoski *et al.*, 2015; Ledevin *et al.*, 2019;) and other localities within the Kaapvaal craton in Southern Africa (Hofmann *et al.*, 2003; Smith *et al.*, 2013; Smith and Beukes, 2023), as well as one sample from the Gogebic Range in the Superior province (Eyster *et al.*, 2021). Sample origins, locations, geological formations and age constraints are given in Table S-1 and Fig. 1.

Given that boron is a fluid-mobile element and that the response of its isotopic signal to post depositional alteration in IFs is not well constrained, the possibility of alteration hampering the original seawater signal exists, especially in Archean-Proterozoic rocks. As IFs are typically comprised of alternations of siliceous and Fe-oxide phases, we measured $\delta^{11}\text{B}$ in both mineral domains in the examined samples to test whether the two phases do indeed preserve an original signal manifested by distinct isotopic values. Such a sedimentary signal results from the different fractionation factors of the two phases relative to seawater. Only samples preserving such distinct values were considered for the reconstruction of the seawater B composition from the $\delta^{11}\text{B}$ value of the siliceous phase. Our data below shows that this approach was constructive, as we can show that IF samples exhibit a pronounced alternation between two distinct B isotopic values depending on the mineral phase (Fig. S-2). Additional support comes from the fact that shale samples from the Moodies group align in their B isotopic composition with silica in IF and chert from the same section (Fig. 3). This serves as independent evidence for the legibility of using a single value for the fractionation factor of both SW–silica and SW–clay during the Archean, as the measured values of both mineral phases precipitated from the same seawater in the same locality at the same time converge to the same $\delta^{11}\text{B}$. This also provides evidence for the lack of matrix effects on our measurements.

From each sample a thin section was cut for petrographic examination, as well as a matching 100 μm (or thicker) thick section for the purpose of in situ $\delta^{11}\text{B}$ analysis by Laser-Ablation Multi-Collector Inductively-Coupled-Plasma Mass-Spectrometry (LA-MC-ICPMS). Prior to analysis, samples were cleaned and then desiccated overnight under inert Ar in order to minimize background caused by degassing into the analytical system. For B concentration and isotopic measurement, samples were ablated using a RESolution LR 193 nm ArF laser ablation system equipped with a large, two-volume Laurin Technic S155 ablation cell. Ablation was performed in a He atmosphere while admixing Ar and N_2 sample gasses into the upper funnel of the ablation cell. Depending on the B concentration in the different samples and the size of target mineral phases, we used laser spot sizes between 60 μm and 235 μm diameter, repetition rates of 8–12 Hz, and laser fluences of ca. 2.0–3.5 J/cm^2 . Ablated material was then directed to the Neptune Plus MC-ICPMS. A primary reference material of NIST-SRM612 glass or GSC-1G basaltic glass was used for tuning the machines for maximum B signal and low background, as well as low oxide formation and low Ar^{4+} production, and later to correct for drift over the analytical sequence. This was achieved by optimizing the MS lenses, as well as the flow of N_2 , Ar sample gas and RF power. All analyses were run in automatic mode in sequences of up to 500 analyses following a standard-sample-standard bracketing procedure, involving duplicate measurements of reference materials following fifty unknown analyses. Secondary standards used for matrix match correction were BIR, BHVO, and BCR basaltic glasses, as well as two in-house hematite reference materials (HEM, Elba; See Table S-4, Fig. S-3). Each spot analysis took 45 s, with 0.25 s integration time, 25 s of ablation, and 20 s of gas blank analysis. Pre-ablation (2 s) followed by 20 s wash out was conducted to minimize potential surface boron contamination. B isotopic measurement on the MC-ICP-MS was performed either using two Faraday cups (equipped with $10^{13} \Omega$ amplifiers) for the m/z 11 and 10 in material with high B concentration, two ion counters in low B concentration, or a combined method where ^{11}B was measured using a Faraday cup and ^{10}B was measured using an ion counter. Due to observed elevated background in the m/z range between 9.7 and 11 (Coenen *et al.*, 2024; “bulge effect”; Standish *et al.*, 2019; Evans *et al.*, 2021; Fietzke and Anagnostou, 2023; Xu *et al.*, 2024), we monitored the intermediary mass range between $m/z = 9.9$ – 10.9 in order to quantify and correct for it. Detailed elaboration on the method developed at FIERCE for measurement of B isotopes in low concentration silicate materials, as well as for the possible causes and means of correction for the ‘bulge effect’ are detailed in Xu *et al.*, (2024). Data processing was performed using an in-house spreadsheet developed by Axel Gerdes. Processing steps include resolving signal from the background, rejection of outliers from the analysis, correction for the bulge interference effect on the ^{10}B and ^{11}B masses (see Standish *et al.*, 2019; Evans *et al.*, 2021; Fietzke and Anagnostou, 2023; Xu *et al.*, 2024), correction for instrumental drift and matrix matching using primary and secondary reference materials respectively, and evaluating boron concentration from the total signal based on post analytical examination of the laser pit volume. Consequently, we are able to attain a $<1\%$ uncertainty (2SE) on replicate analysis ($n > 10$) in samples with $<1 \mu\text{g/g}$ [B].

On individual analyses, the reported uncertainties on the $\delta^{11}\text{B}$ value are quoted at the 2-sigma level and are quadratic additions of the internal measurement uncertainty, the excess scatter of the primary reference material and the excess

scatter of the secondary reference material. Reported values of each of the measured samples are the product of replicate analysis ($n = 30\text{--}100$) of a mineral phase within a thin section or slab. To produce these, we calculate a weighted average which considers both the measured value as well as the associated analytical error of each considered datapoint. We then report the weighted mean $\pm 2\text{SE}$, as the standard error quantifies well the confidence in the mean value, despite the possibility of natural internal scatter. For sample populations (like pre- and post-3.0 Ga, see main text), a simple mean with 2SD and 2SE are reported, as well as the results of standard Welch's t-test, to quantify the distinction of the two populations. All data reduction calculations, including error propagations and statistical analysis were performed by using code-supported Microsoft Excel®-based spreadsheets developed by Axel Gerdes.

1.2.1 Post hoc processing of sample data:

The majority of measured samples (43 out of 48) does exhibit rather homogeneous compositions of boron concentration as well as $\delta^{11}\text{B}$ values (Fig. S-4). Some of the samples do show the expected B partitioning and isotope fractionation between points measured in silica and Fe-oxide zones. However, it is also expected that due to either detrital contamination or inhomogeneity of silica and Fe-oxide zones (e.g. small fragments of Fe-oxides within silica bands and vice versa), measured values will show some spread between the pure silica endmember and the detrital and Fe-oxide endmembers. In order to control these effects, we examined our results in Fig. S-4, and identified samples which required further screening to make sure the quoted values indeed represent silica endmember from which we can reliably reconstruct the seawater composition. A prime example of such treatment is seen in sample P7.3f (Fig. S-4), in which the pure Fe-oxide endmember is evident, but also values from the measured silica points that show a large spread between a group of low-B, high- $\delta^{11}\text{B}$ spots and the high-[B], low- $\delta^{11}\text{B}$ of the Fe oxide. This spread between points of low concentration with heavy isotopic composition and points of high boron concentration with light isotopic composition represent mixing of a Fe oxide (and possibly a detrital component) within the silica. As continent-derived detrital material would consist mostly of clay, it would also contain high boron concentration and show an isotopically light composition (Ishikawa and Nakamura, 1993). In samples showing a trend like this we applied a filter to consider only the silica endmember for seawater reconstruction. We selected a subset of spot from within each such sample which show low boron concentration and high $\delta^{11}\text{B}$. Such a filter was applied to 5 samples in total (Fig. S-4).

1.3 Reconstruction of crustal emergence from seawater $\delta^{11}\text{B}$:

The reconstruction of crustal emergence is based on a number of assumptions. These include: (1) a near-equilibrium state between seawater and its boron sources and sinks; (2) uniform erosion of the continents that is, on the timescale of tens of millions of years, independent of topography and exposed rock types; (3) that other fluxes, such as seafloor-alteration and adsorption to sediments remained constant.

Elevated activity of oceanic spreading centres at 3.5–2.9 Ga relative to modern Earth, as predicted by higher mantle potential temperatures and estimates of elevated degassing (Herzberg *et al.*, 2010; Aulbach and Arndt, 2019) would have removed increased amounts of boron from the ocean. However, increased boron removal by oceanic crust alteration drives seawater towards heavier $\delta^{11}\text{B}$ (due to the large negative fractionation factor; Fig. 2a in main text). To balance

this effect and keep seawater $\delta^{11}\text{B}$ at a lighter value, more continental crust would need to emerge and erode. This renders our estimates of continental exposure a minimum one. Our calculation assumes boron is spread and erodes uniformly within and from the continental crust. Rapid buildup of orogens in a condensed period causing enhanced emergence, the complex behavior of boron in the continental weathering environment (Gaillardet and Lemarchand, 2018), rapid eruption and erosion of flood basalts with near mantle $\delta^{11}\text{B}$ composition, the difference in boron concentration between dominantly mafic and felsic crust – all these may introduce complexities to this estimate, and are expected to guide future models building on our data. Nevertheless, it is unlikely that a process other than continental erosion could introduce large quantities of boron into the ocean and drive the oceanic composition towards a heavier value over extended periods of Earth history to change the first-order results of our model.

The model is further based on a number of fix points that are reasonable to assume or were established in this study:

- 1) In the Hadean (i.e. at 4 Ga), there was no continental crust emerged above sea level
- 2) In the late Neoproterozoic (i.e. at 635 Ma) 100 % emergence was reached (relative to the modern amount; but in a world without land plants)
- 3) The $\delta^{11}\text{B}$ of seawater at 3.70–3.21 Ga was +1.9 ‰
- 4) The $\delta^{11}\text{B}$ of seawater at 2.98–2.40 Ga was +16.1 ‰
- 5) The $\delta^{11}\text{B}$ of seawater at 1.88 Ga was +18.6 ‰
- 6) The $\delta^{11}\text{B}$ of seawater at 635 Ma was +25.5 ‰

Continental emergence in the Archean and throughout most of the Proterozoic (3.70–0.64 Ga) is unknown and its first-order reconstruction is the goal of this study. There may have been some emergence prior to the deposition of our oldest sample (3.7 Ga); hence, the model can at this stage only provide brackets for a possible evolution instead of an accurate solution. For this, we choose the following solution:

We first assume that seawater $\delta^{11}\text{B}$ can never be lower than the $\delta^{11}\text{B}$ of the primitive mantle (-7.1 ‰), because the differentiation of the silicate Earth and the hydrosphere started with the mantle and there is no mechanism that would fractionate the lighter isotope into water. Hence, the lowest possible $\delta^{11}\text{B}$ value of seawater in the Hadean in a world free of any (emerged) continents is -7.1 ‰. The highest possible value is +1.9 ‰, assuming that our reconstruction of Eoarchean and Paleoarchean (3.70–3.21 Ga) seawater reflects a world that still had no (or only negligible) emerged continents and thus seawater $\delta^{11}\text{B}$ had not changed at all until the end of the Paleoarchean.

In the first case, with the lowermost possible Hadean seawater $\delta^{11}\text{B}$ value (-7.1 ‰), seawater would have already risen by 9 ‰ (from -7.1 to +1.9) between the Hadean and the deposition of our oldest sample (3.7 Ga). Continental emergence would have driven seawater from -7.1 ‰ at 4 Ga to +1.9 ‰ at 3.7 Ga to eventually +25.5 ‰ in the Neoproterozoic. Hence, in this model, 100 % emergence is equivalent to a 32.6 ‰ rise in seawater $\delta^{11}\text{B}$, and a change of 1 ‰ is equivalent to 3.1 % increase in crustal emergence (hence, 1 % of crustal emergence causes a shift of 0.33 ‰ in seawater $\delta^{11}\text{B}$). The 9 ‰ increase to +1.9 ‰ by 3.7 Ga thus means 27.6 % crustal emergence in the earliest Archean, and the total of 23.2 ‰ increase to +16.1 ‰ by 2.98 Ga means 71.2 % emergence at that time.

The upper-boundary model (second case), which suggests that no continental crust had emerged prior to the deposition of our oldest sample, means that Hadean seawater already had a $\delta^{11}\text{B}$ of +1.9 ‰, and this value represents an ocean without (significant) emerged continental crust. In this case, 100 % emergence led to a total change in seawater $\delta^{11}\text{B}$ from +1.9 to +25.5 ‰ (a change of 23.6 ‰) between 3.21 Ga and the Neoproterozoic. Hence, a change of 1 ‰ would be equivalent to 4.2 % increase in crustal emergence (or 1 % of crustal emergence causes a shift of 0.24 ‰ in seawater $\delta^{11}\text{B}$). At 3.21 Ga with a seawater $\delta^{11}\text{B}$ of +1.9 ‰ there was 0 % crustal emergence, and at 2.98 Ga with seawater $\delta^{11}\text{B}$ = +16.1 ‰ (a 14.2 ‰ increase) there was 60.2 % emergence.

Based on the currently available data, the amount of emerged continental crust in the Eo-/Paleoarchean is bracketed to between zero and approximately 1/4 of the amount of crust emerged that we see today (0–27.6 %). The value of emergence in the late Archean to early Proterozoic (2.98–2.40 Ga) is not very sensitive to the choice of model: it only changes from 60.2 % continental emergence for one extreme to 71.2 % for the other extreme.

Supplementary Tables

Tables S-1 to S-5 (.xlsx) are available for downloading from the online version of this article at <https://doi.org/10.7185/geochemlet.2635>

Table S-1	list of samples and their origins used in the present study
Table S-2	LA-MC-ICP-MS B isotope data from all samples measured in this study:
Table S-3	LA-MC-ICP-MS B isotope data from all standards measured in this study:
Table S-4	LA-MC-ICP-MS B isotope data report
Table S-5	Crustal emergence model calculation table

Supplementary Figures

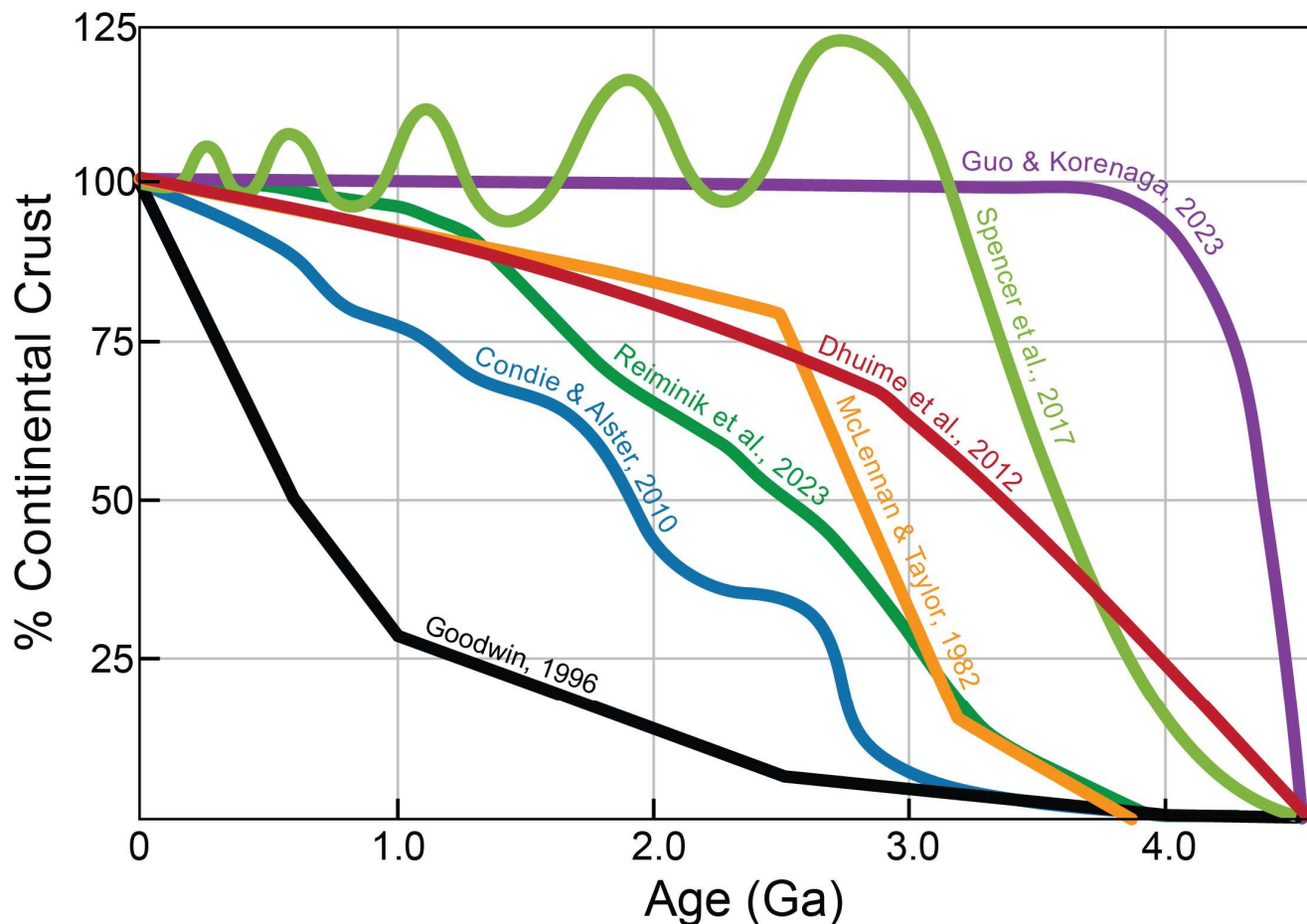


Figure S-1 Selected crustal growth models showing different modes of continental crust accumulation throughout geologic time, founded in different considerations: ^{142}Nd anomalies in Archean Mantle rocks (Guo and Korenaga, 2023), O and Hf isotopes in igneous and detrital zircons (Dhuime *et al.*, 2012), adjustment of growth curves to account for biased preservation (Spencer *et al.*, 2017), Eu anomalies and Nd- T_{DM} ages in sediments (McLennan and Taylor, 1982), major element indexes identifying reworked crustal rocks (Reimink *et al.*, 2023), present day distribution of juvenile continental crust (Condie and Aster, 2010), and present day distribution of all exposed continental crust (Goodwin, 1996).

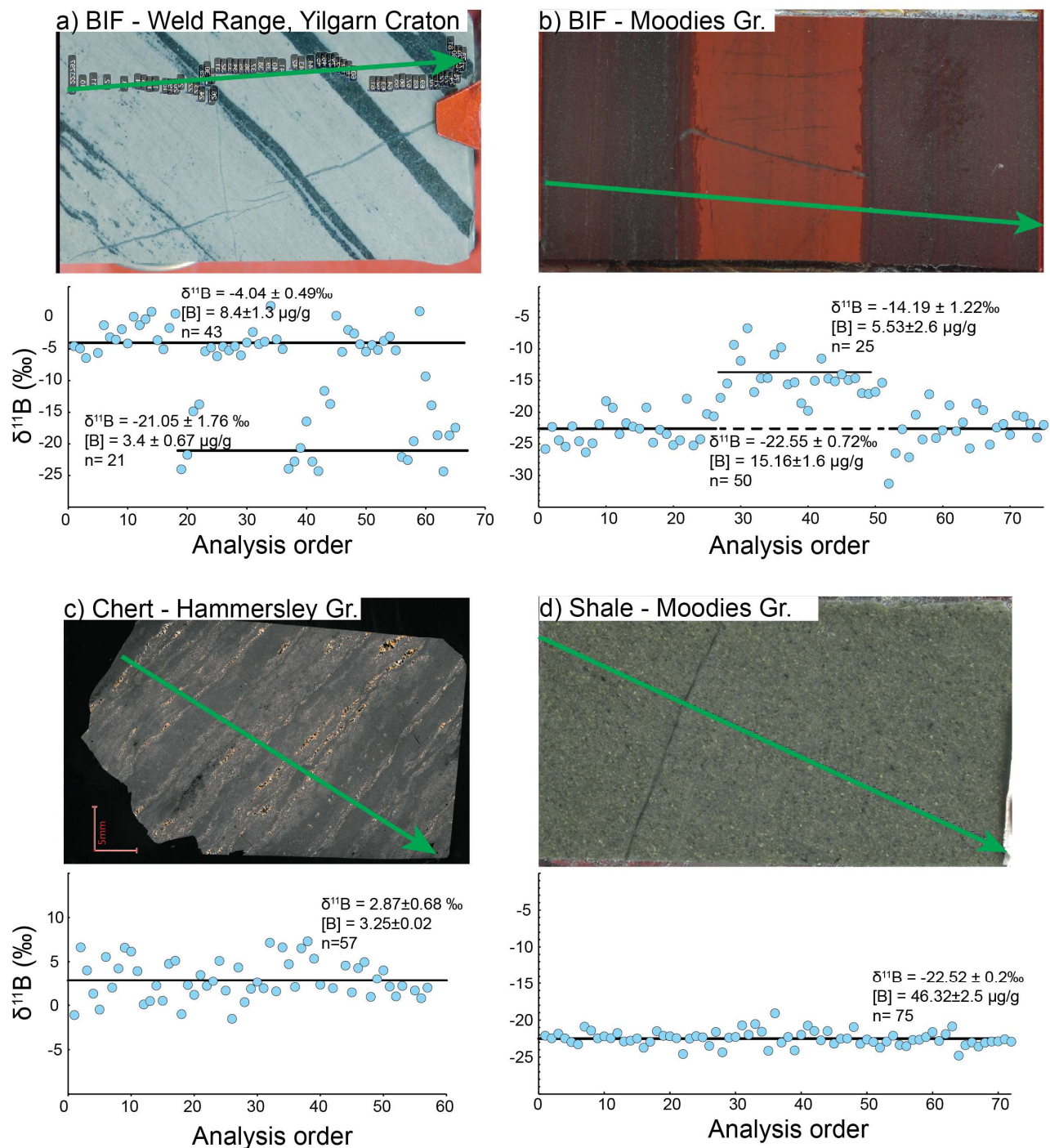


Figure S-2 Representative images of analyzed samples showing bi-modal variability of $\delta^{11}\text{B}$ in Iron Formations, or lack thereof in single phase samples like chert and shale. Green arrow denotes the direction of spot analysis order across the thin-section or slab in each sample.

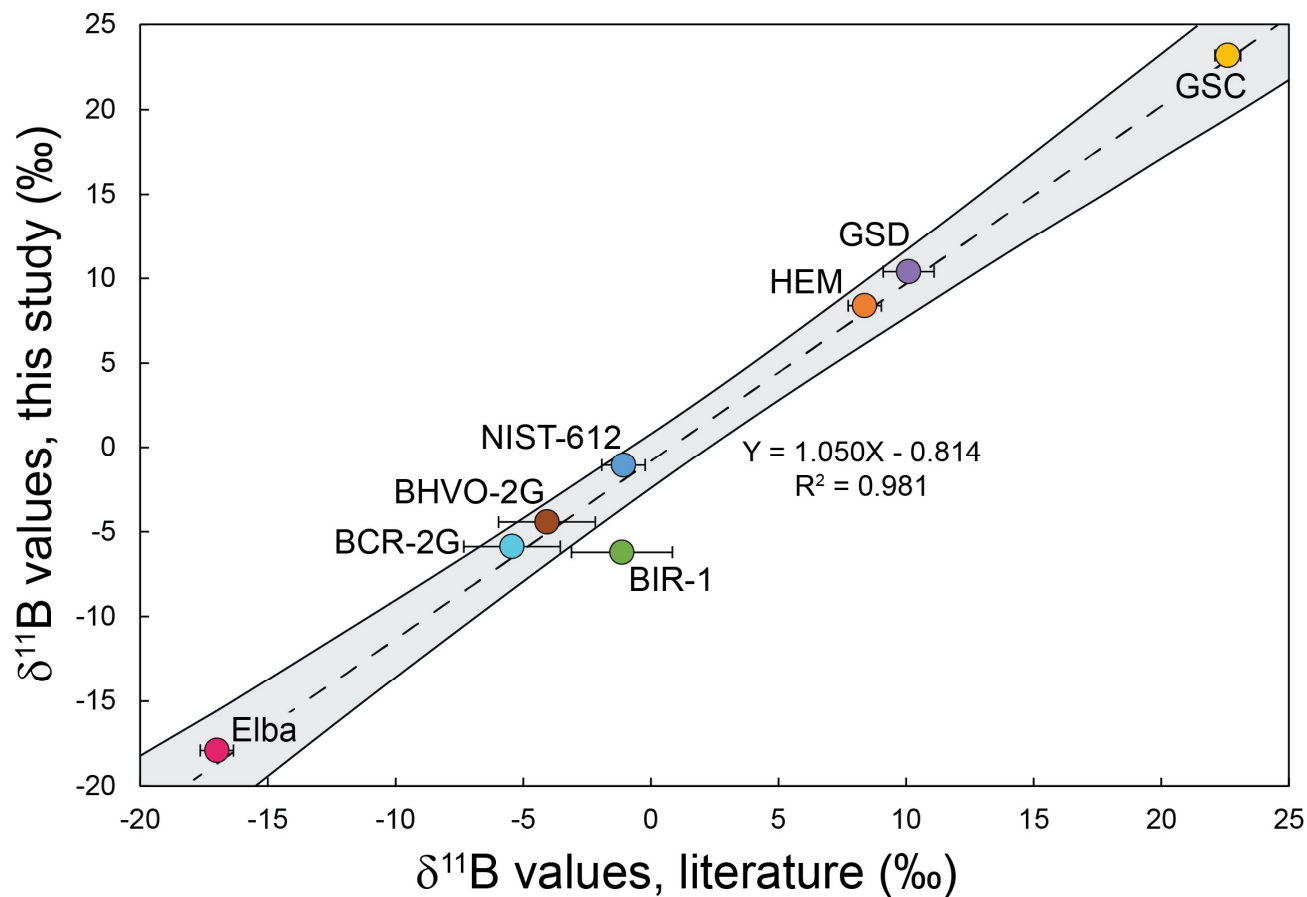


Figure S-3 Comparison of $\delta^{11}\text{B}$ values of standard reference materials determined in this study with weighted mean value of literature values (He *et al.*, 2020; Jochum *et al.*, 2011; Kasemann *et al.*, 2010; Xu *et al.*, 2024). Horizontal bars represent 2SE of the compiled literature values, depicting the variability of published data. Vertical bars are 2SE of values measured in this study, most of which are smaller than the symbol size (Table S-3). Dashed line shows a linear regression of the data, with the shaded area depicting 95% confidence level around it.

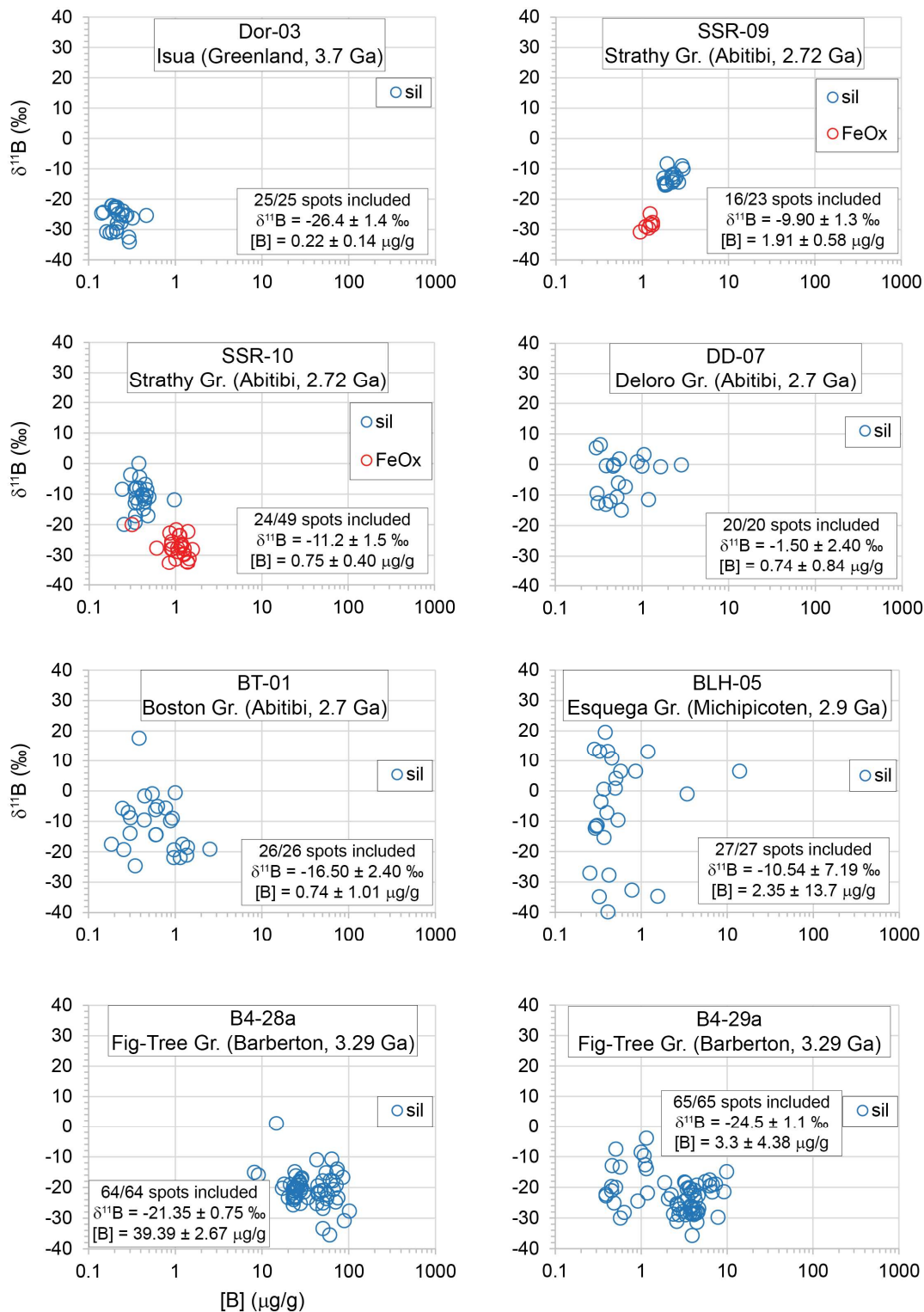


Figure S-4 Measured $\delta^{11}\text{B}$ vs. [B] values of every analytical point within each sample, from both silica and Fe-oxide phases. Samples showing a trend of mixing between silica and Fe-oxide or a detrital component were further filtered and their silica value was evaluated from a subset of points of the silica end-member. See text for explanation.

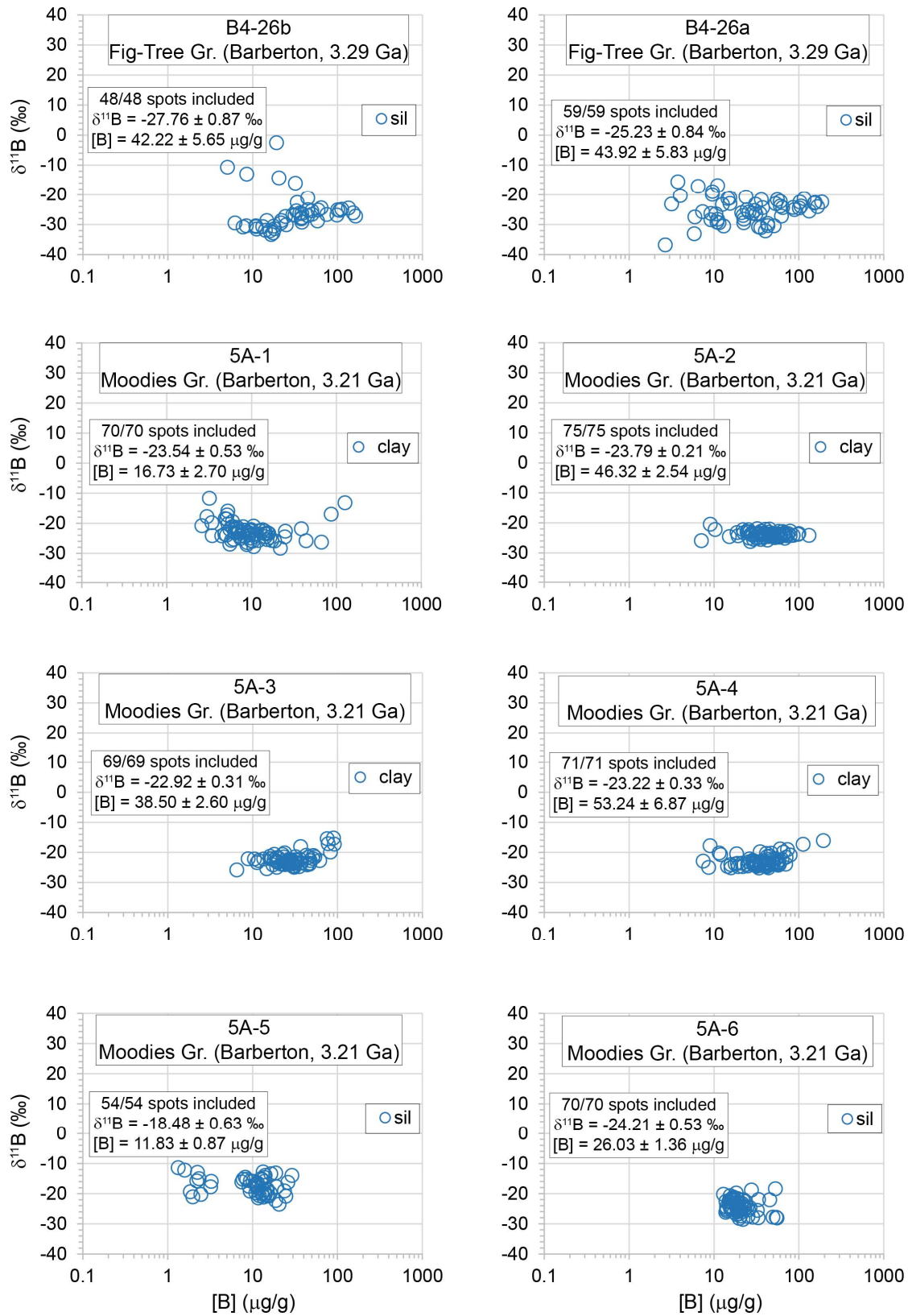


Figure S-4 cont.

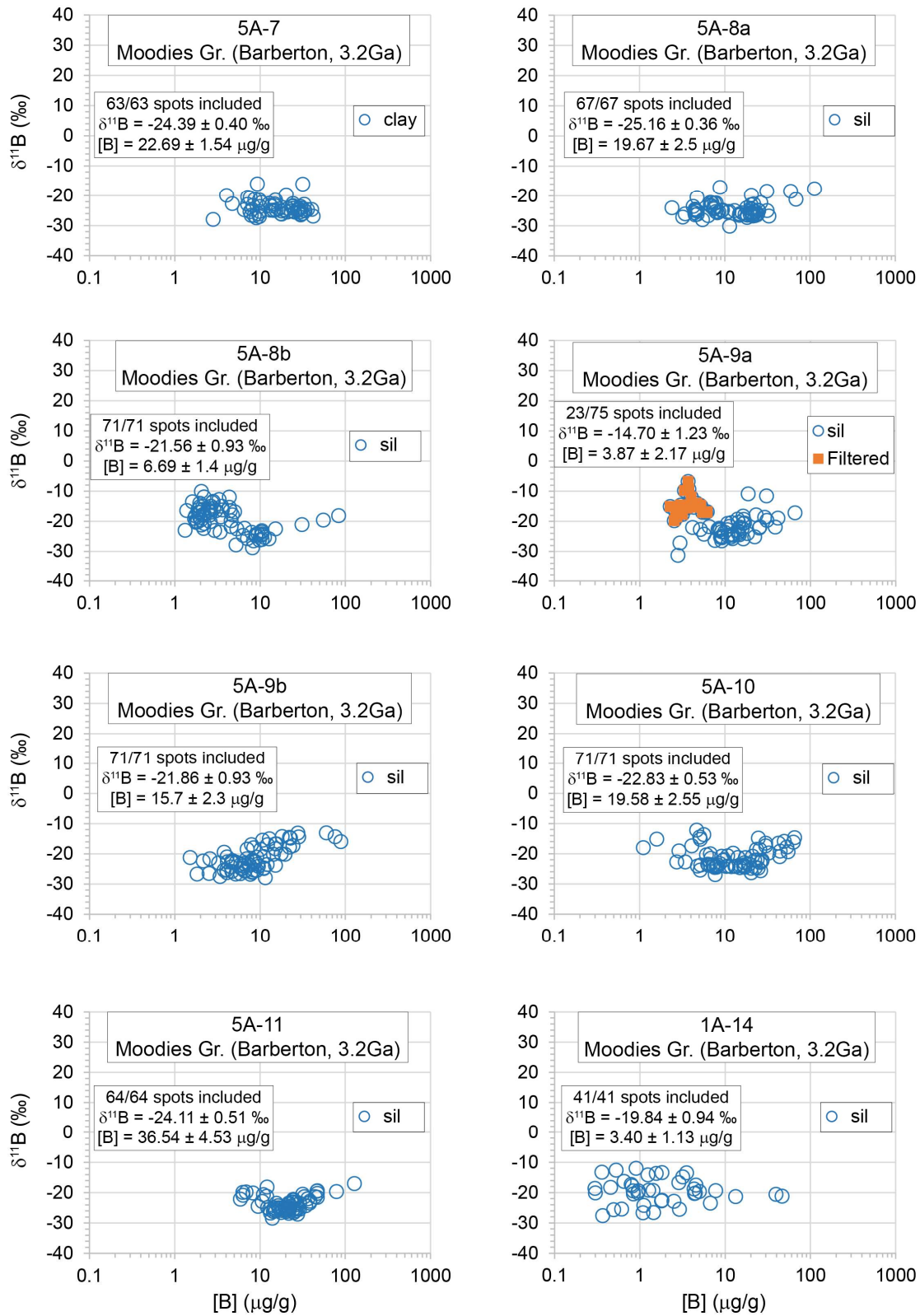


Figure S-4 cont.

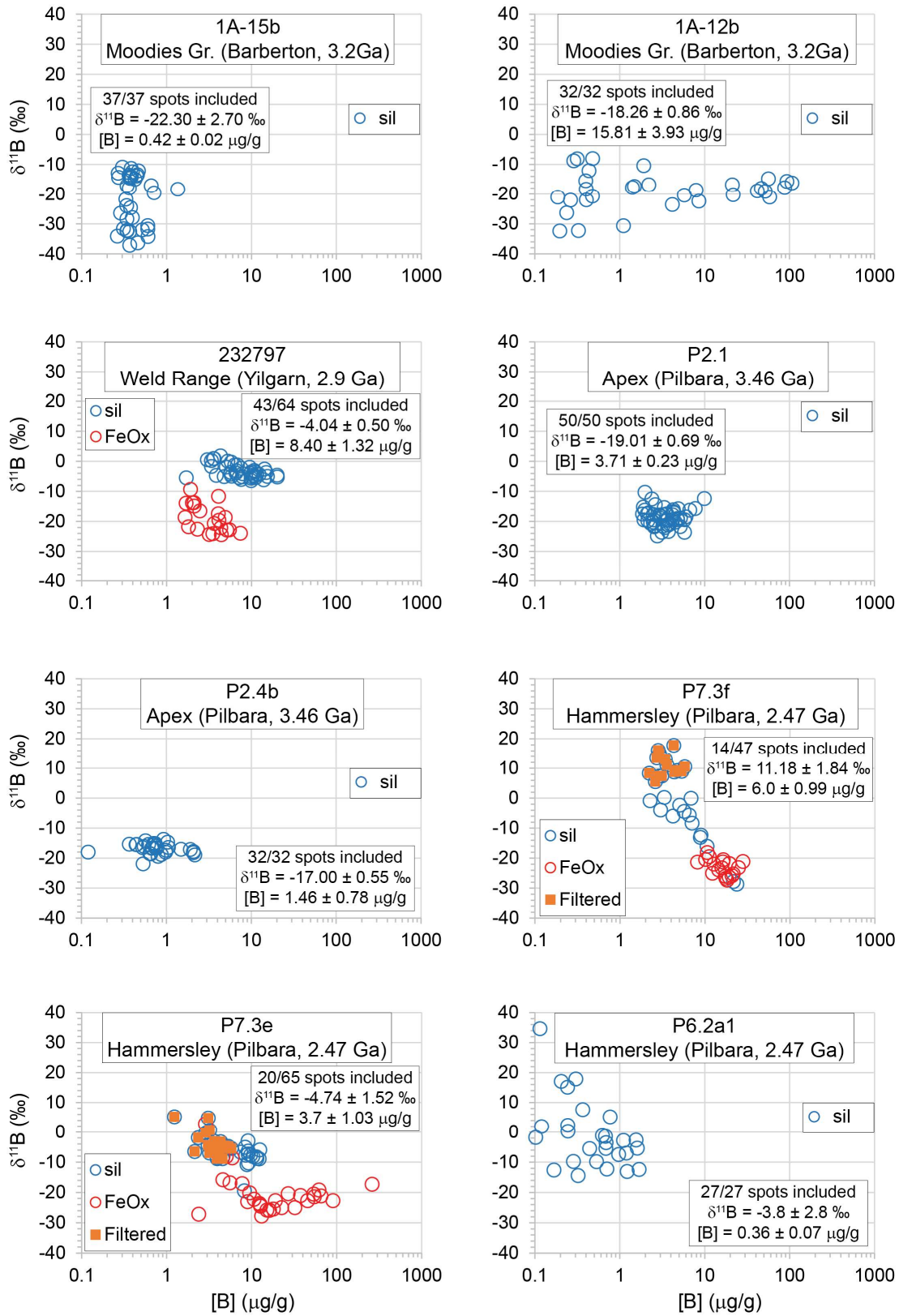


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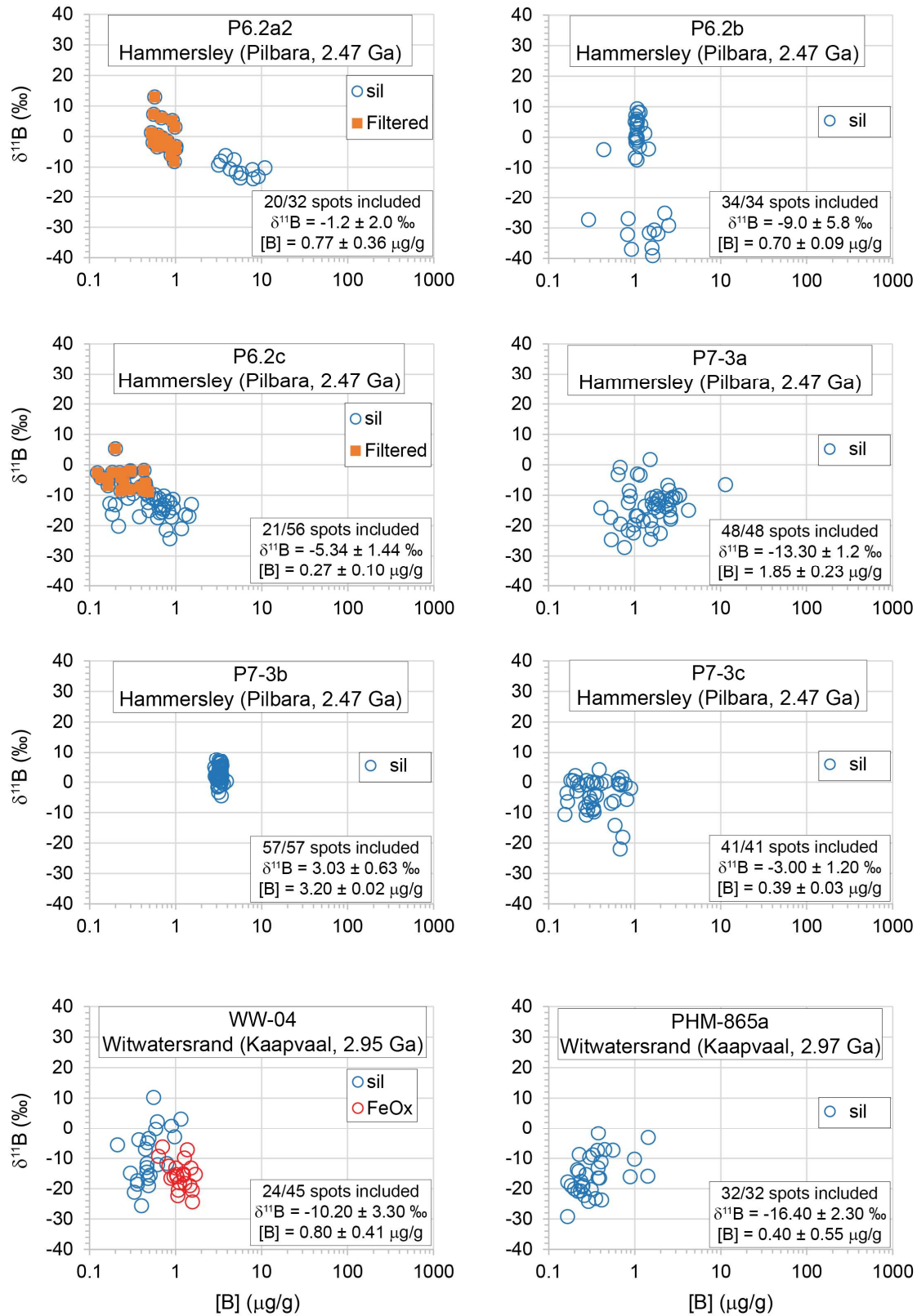


Figure S-4 cont.

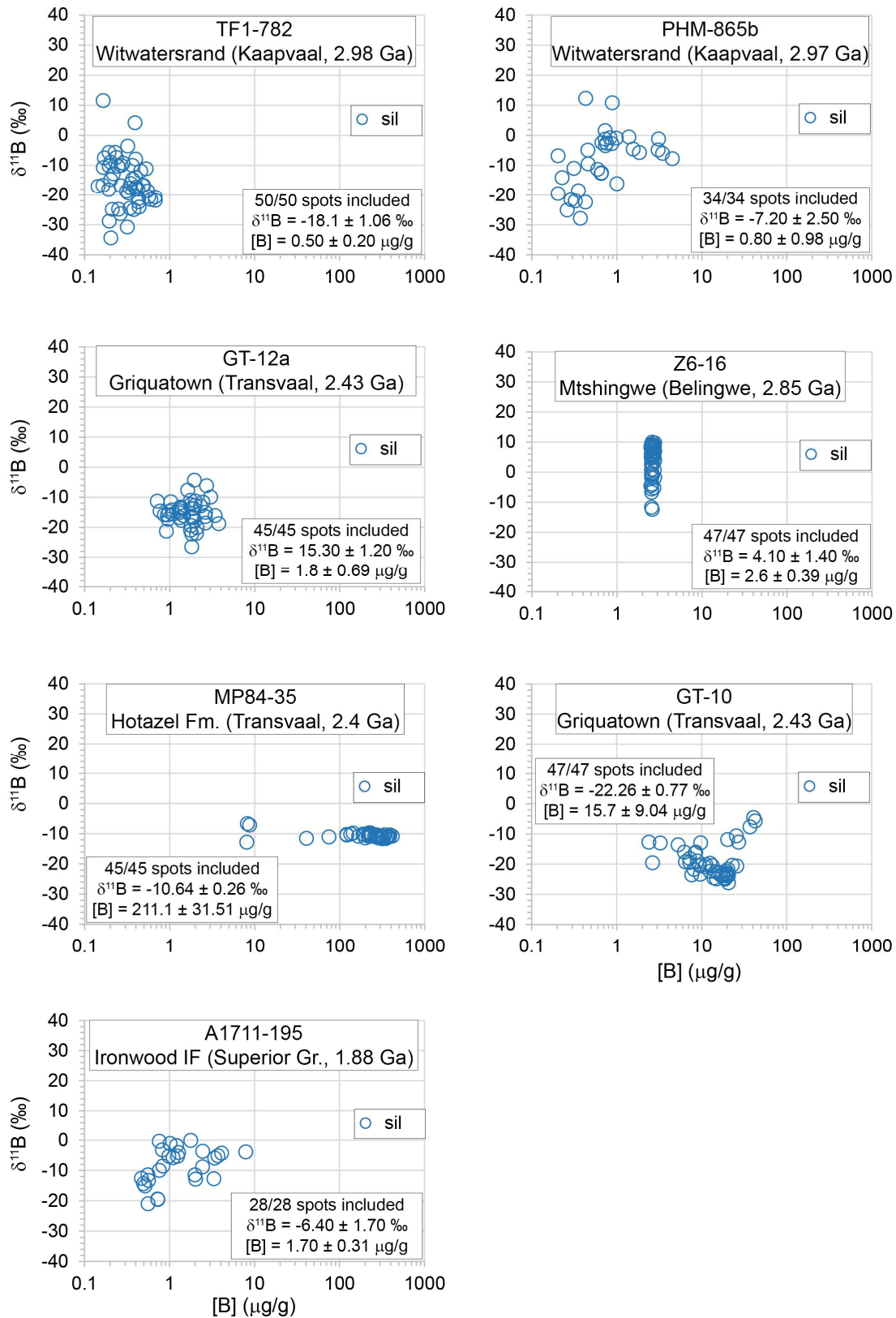


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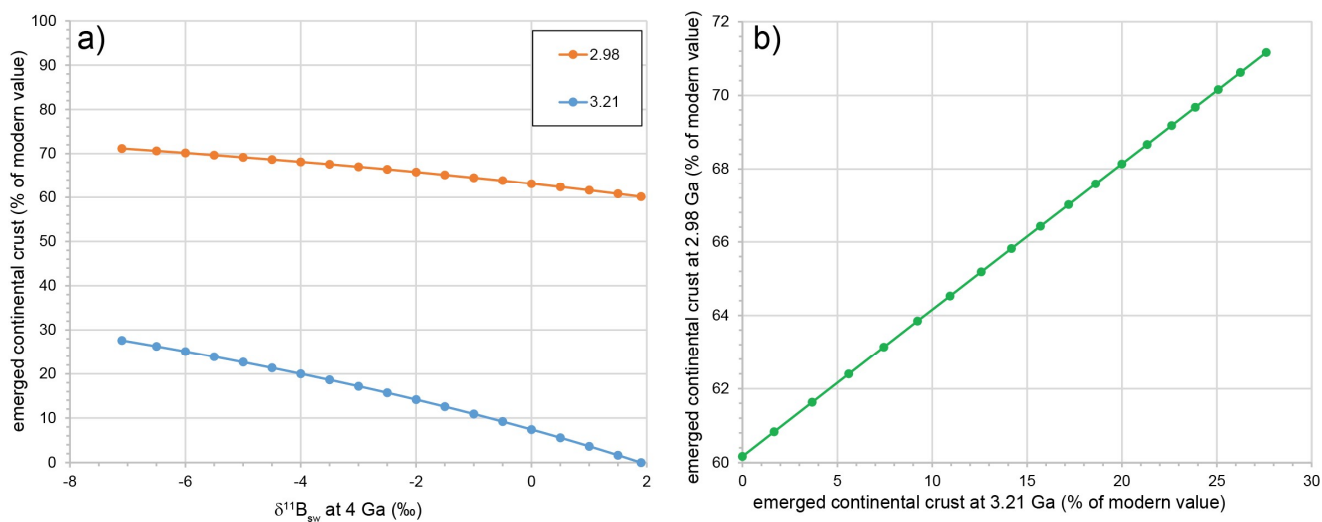


Figure S-5 Results of the mass balance model calculation. a) Comparison of emerged continental crust (expressed as % of the modern value) as a function of $\delta^{11}\text{B}_{\text{sw}}$. b) The variation of emerged continental crust at 2.98 Ga as a function of the amount of emerged crust at 3.21 Ga. See text and supplementary data for discussion.

Supplementary Information References

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