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Key Points:
- A subset of iron formations, known as Algoma-type, can be interpreted as deeper marine sediment accreted atop oceanic crustal rocks
- We present a preservation-corrected iron formation temporal record
- The scaled iron formation record suggests widespread, potentially ocean-wide, and continued deposition of iron formations from 3.8 to 1.7 Ga

Supporting Information:
- Supporting Information S1

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Widespread and Persistent Deposition of Iron Formations for Two Billion Years

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Abstract Composed of chemical precipitates rich in iron and silica, Precambrian iron formations from marine sedimentary records may reveal biogeochemical processes over the first half of Earth history. The limited record of early Archean rock suggests that preservation biases the iron formation record. Like ophiolites, which provide a sparse record of past ocean floor, iron formations deposited on oceanic crust ought to also be rare and preserved only when accreted onto cratons. To correct for potential preservation bias, we scaled masses of iron formations to the areal extent of basement rock of similar age and found that the resultant record is consistent with persistent deposition of iron formations across much of the deep ocean for two billion years. Widespread and long-term iron formations imply that ferrous iron was available in the deep ocean for billions of years and that the requisite (bio)geochemical mechanisms to produce iron formations were present by 3.8 Ga.

Plain Language Summary Why did the ancient ocean deposit an increasing volume of iron over time, until these “Iron Formations” largely disappeared? Does this pattern reflect a growing titration of oxygen, a surge in iron from mantle superplumes, or the prerequisite of developing continental crust? Or is this distribution simply from the diminishing record of ancient rocks? We normalize the iron formation record over time by scaling it to crustal preservation, and the preservation-scaled record suggests that there may have been widespread and unchanging deposition of these iron formations across the ancient oceans for two billion years—implying that the (bio)chemical process responsible for making iron formations operated from ~3.8 to ~1.8 billion years ago.

1. Introduction

With no modern analogues for the early ocean, the study of ancient marine iron- and silica-rich chemical precipitates offers insights into how this alien sea operated (e.g., Klein, 2005; Konhauser et al., 2017). These peculiar precipitates, containing ~30% iron and ~50% silica, were deposited primarily between ~3.8 and ~1.7 billion years ago (Ga) in sediment-starved marine settings (Klein, 2005; Trendall, 2002). Known as “iron formations,” here abbreviated IFs, they are studied both as records of the Archean and Proterozoic ocean chemistry and as sources of economically important iron ores (Clout & Simonson, 2005). Although the mechanism that formed the original IFs remains enigmatic, the transport and concentration of reduced iron in the ocean, away from continental weathering and hydrothermal sources, indicate that environmental oxygen levels were much lower than today (e.g., Bekker et al., 2010; Cloud, 1968).

The uneven IF record, with large volumes near the Archean-Proterozoic boundary at 2.5 Ga, has prompted many hypotheses suggesting various controls on IF deposition. Cloud (1973) proposed that early microbial oxygen producers originally were in balance with reduced iron, suppressing the release of oxygen, but then these microbes overcame physiological barriers to completely titrate the iron and allow a rise in oxygen. Recent dating, however, shows that the apparent ~2.5-Ga IF peak precedes the ca. 2.4–2.3 Ga major rise in oxygen (“Great Oxidation Event” or GOE; Gumsley et al., 2017; Luo et al., 2016). More recently, others have considered the IF temporal distribution as regulated by either limited stable depositional repositories (James, 1983; Simonson, 2003; Trendall, 2002) or insufficient ferrous iron, with influxes of ferrous iron—perhaps from upwelling hydrothermal fluids or mantle plume events—stimulating major IF deposition (Barley et al., 1997; Holland, 1973; Isley & Abbott, 1999; Konhauser et al., 2017; Morris, 1993). Others have suggested the confluence of both depositional basins and fluxes of iron together modulated the IF record (Bekker et al., 2010, 2014; Clout & Simonson, 2005).

An obstacle to evaluating hypotheses for the temporal distribution of IFs is the long-recognized underestimation of Archean IFs (Gole & Klein, 1981; James, 1983). Both subduction and deformation leading to "dismemberment," combined with erosion and the dwindling exposure of the older rock record on modern Earth, have greatly diminished the original extent of the earliest IFs (Gole & Klein, 1981). Additionally, the growth of continental crust and continental shelves over time not only affected the volume of IF that might have been deposited, but also provided environments that can preserve IFs (Simonson, 2003; Trendall, 2002). Although previous publications and compilations acknowledged these biases, we are not aware of any attempt to quantify and correct them, apart from a recent illustrative analysis using just the record from North America, New Zealand, and the Caribbean (Peters & Husson, 2018).

Another complicating factor is the exclusion of small, igneous rock-associated IFs from compilations of IFs over geologic time (Bekker et al., 2010, 2014). Because lithological hosts of IFs vary, differing categorizations of IFs have been proposed (Beukes, 1980; Gross, 1965; James, 1954; Kimberley, 1978). In the most widely used of these, Gross (1965, 1980) defined Algoma-type IFs as associated with volcanic rock in deep water settings, at least below wave base if not at depths of kilometers, and Superior-type IFs as affiliated with strata deposited on continental shelves and slopes. Most Algoma-type IFs consist of relatively thin (centimeters to tens of meter) layers associated with greenstone belts of igneous and sedimentary rock assemblages, whereas many of the thicker (often hundreds of meters thick) and more extensive Superior-type IFs accumulated on cratonic margins (Bekker et al., 2014; Konhauser et al., 2017). Despite the Algoma-type IFs having similar geochemistry to the Superior-type IFs (Beukes & Gutzmer, 2008; Gole & Klein, 1981; Trendall, 2002), Bekker et al. (2010, 2014) excluded Algoma-type IFs in recent IF compilations, with the argument that these record local environments associated with volcanic or hydrothermal inputs rather than representing large-scale ocean chemistry.

We contend that Algoma-type IFs must be included in IF compilations to understand the significance of IFs as recorders of early Earth marine biogeochemistry. Outcrops of Algoma-type IFs, often deposited on basalt in ocean basins, share features associated with ophiolites in more modern tectonic settings. Although we do not equate them with ophiolites, we suggest many Algoma-type IFs to be remnants of ocean floor sediment, later accreted onto continents along with the underlying basalt rock, as Rosing et al. (1996) interpreted for the earliest known IF (Isua). When smoothly functioning, subduction removes most oceanic crust, so that many obducted ophiolites can be seen as accidents (Dilek & Furnes, 2014; Moores, 1982). The ocean crust and uppermost mantle comprising many ophiolites formed in back arc settings, in some cases with island arcs built on them, and were later emplaced on continental crust as that crust was subducted beneath the arc (Dilek & Furnes, 2014; Moores, 1982). Regardless of how ophiolites have been emplaced, however, they offer an incomplete record of past seafloor. Whether or not basaltic crust capped rigid plates of lithosphere, as in classic plate tectonics, convective overturn in the Archean mantle likely carried oceanic crust back into the mantle. Accordingly, preserved oceanic crust with Algoma-type IFs should similarly be viewed as an incomplete record of past ocean floor, including volcanic arcs in some cases. From this perspective, their abundance in the geologic record today places only a qualitative constraint on the extent to which iron and silica were deposited in Archean time. Despite these limitations, Algoma-type IFs may actually provide our best record of the sediment, and the geochemical record that it contains, deposited on the Archean ocean floor.

The Superior-type IFs, associated with voluminous deposition on the margins of cratons, have a clear relationship to the preservation of continental crust (e.g., Trendall, 2002). Cratonic margins, including convergent settings where flexure of the lithosphere may have dropped strata below wave base (e.g., Fedo & Eriksson, 1996; Hoffman, 1987, 1988), provide preservational advantages for these IFs, even when buried or folded in collisional orogenies. Continental crust is predicted to escape subduction (e.g., McKenzie, 1969; Molnar & Gray, 1979), which efficiently banishes thinner, denser oceanic crust back into the mantle. Moreover, thick layers of buoyant Archean upper mantle caused its overlying continental crust to be less susceptible to both subduction and deformation than crust that caps younger lithosphere (Jordan, 1988). If Earth wanted to preserve intact sedimentary layers, it could choose no better place than Archean cratons, including their margins. The evidence for widespread development of continental margins near the Archean-Proterozoic boundary (Beukes, 1987; Bickle et al., 1975; Bradley, 2008; Campbell & Davies, 2017) may account for the exceptional examples of Superior-type IFs from ~2.7 to 2.4 Ga.

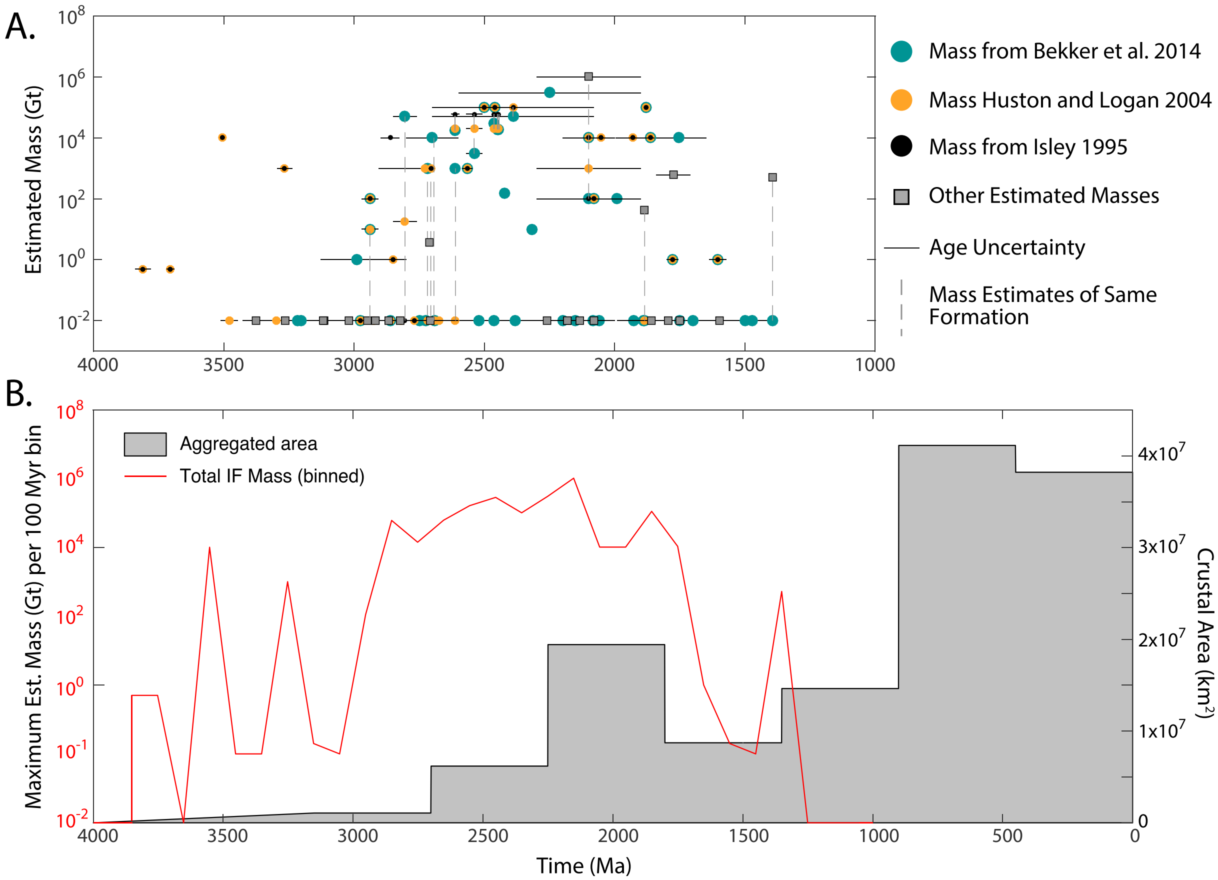

Figure 1. (a) Mass estimates for IFs reported in the literature, with age uncertainty plotted as error bars. IFs that were reported but not given a mass or described as small/unknown in previous compilations were assigned a small mass of 0.01 Gt arbitrarily to indicate their presence. Note the abundance of low-mass but prevalent IFs throughout the Archean and early Proterozoic. (b) Aggregated areal extent of continental basement rock (gray), after Hurley and Rand (1969) who binned areas into 450-million year periods, overplotted with the IF estimated masses (red) binned in 100 Myr intervals and using the maximum estimates for each formation solely to compare the general trend of the IF record against the temporal increase in preserved and exposed continental crust. IF = iron formation.

2. Removing Preservational Biases From the IF Record

To determine whether IF deposition underwent an increase at ~2.5 Ga or the major IF volumes around this time reflect the growth of appreciable and better preserved continental shelf area (Figure 1), the IFs need to be scaled to a measure of the preservational biases against the older rock record. Although Archean continental crust might have been nearly as extensive as crust today (e.g., Armstrong, 1981), most of that crust has been altered or buried, and can provide no record of early IFs. To estimate how biased the present-day record of IFs is, we need to weight estimates of the initial IF masses (estimated using the observed thicknesses and areal extents of each IF, in gigatons (Gt), Table S1) against the amount of preserved rock from that time period. Using K/Ar ages, adjusted for Ar loss using Rb/Sr ages and initial values of $^{87}Sr/^{86}Sr$, Hurley and Rand (1969) measured areas of exposed continental basement rock through time from the America, Africa, Europe, India, Australia, and Antarctica; they could not access adequate information for the former Soviet Union or China. Because of the dependence on K/Ar ages, some of their ages were likely incorrect, but the considerable amount of data used and the binning into 450-Myr windows compensates for such errors. Hurley and Rand's compilation therefore contains accurate dates for the regions and time bins that they considered. The areas of exposed continental basement are subject to the same biases that the ancient IFs have: limitations from subduction, tectonic reworking, and exposure. Therefore, using Hurley and Rand's measurements removes much of the preservational biases embedded in the raw temporal record of IF volume. We assigned each IF to the appropriate 450-Myr bin assigned by Hurley and Rand (Figure 1b) and divided the mass estimate by the total measured area of exposed continental crust from that age bin. For IFs older than 3.15 Ga, we assumed the preservation of crust increased linearly from nil at 4.0 Ga to

1,065,600 km^2 at 3.15 Ga, the value that Hurley and Rand gave for the interval from 3.15 to 2.7 Ga (Figure 1b). We assign no physical meaning to the dimensions, gigatons per square kilometer, of these ratios, and merely use them to compare accumulation amounts over a period of time during which preservation has drastically changed.

We applied this preservation scaling to previous compilations of IF tonnage estimates over time from 4 to 1 Ga (Figure 2) (Bekker et al., 2014; Huston & Logan, 2004; Isley, 1995; Isley & Abbott, 1999). The masses for each IF are order-of-magnitude estimates originally from James (1983), who acknowledged that the fragmentary nature of the geologic record suggests “that the time distribution pattern may be more related to accidents of preservation than to a time-controlled cycle of deposition.” Successive compilations, by Isley (1995), Isley and Abbott (1999), Huston and Logan (2004), and then Bekker et al. (2010, 2014), added IFs and refined mass estimates. For virtually every major IF, masses are estimated to only one significant figure, and therefore uncertainties are at least a factor of 3 of those estimated masses. Continually improving geochronology allowed us to refine some IF age constraints (Figure 1a and Table S1). Scaling these mass estimates to preserved continental area alters the impression of an IF peak at ~2.7–2.4 Ga, as Figures 2 and S1 show with histograms of the raw masses of IFs over time compared to preservation-scaled IF amounts.

The approach taken here necessarily ignores some major IFs. In particular, those in China and the former Soviet Union were excluded from Figure 2 as we cannot correct them for preservational bias using Hurley and Rand's areal compilation. From China, we excluded the small 1700 ± 100 Ma Chuanlinggou IF (Wan et al., 2003; Bekker et al., 2014), the ~100 Gt Liaohe Formation constrained to 1990 ± 60 Ma (Bekker et al., 2014; Luo et al., 2004), and the more important 2700 ± 100 Ma Anshan Liaoning deposit with 10,000 Gt of IF mass (Table S1; Bekker et al., 2014; Hao et al., 2017). Canfield et al. (2018) recently revised the mass of the ca. 1400 Ma Xiamaling Formation to 520 Gt. The imprecisely dated (between 2700 and 2080 Ma) Krivoy Rog IF in the former Soviet Union is estimated to have contained 50,000–100,000 Gt of IF (Bekker et al., 2014; Isley & Abbott, 1999; Kulik & Korzhnev, 1997). The Belozyorsky-Konsky IF in Ukraine, with an estimated age of 3267 ± 26 Ma (Isley, 1995), may have originally contained 1000 Gt of IF (Huston & Logan, 2004; Isley, 1995). The Kursk Magnetic Anomaly suggests that a large unexposed IF is in the Ukrainian shield, with estimated iron reserves at least 13 times larger than Krivoy Rog (Alexandrov, 1973). Similarly, Launay et al. (2018) inferred from a large magnetic anomaly that the total volume of the ca. 2.1 Ga West African Ijil IF is 350,000 km^3, and therefore ~1,000,000 Gt, more than three times larger than any other IF (Figure 1). Notably, this IF accumulated after the GOE. Since we correct for biases of preservation and exposure by normalizing to exposed continental crust, however, we use previous estimates of the Ijil deposit based on its exposed volume to scale for preservation.

The most recent estimates of Superior-type IF masses (Bekker et al., 2014) combined with those from Algoma-type IFs (Huston & Logan, 2004; Isley & Abbott, 1999) and other estimates (Table S1) yield a compilation of IF mass/area ratios from 4.0 to 1.0 Ga (Figure 3), evaluating the temporal record of reported IFs from the Eoarchean to the Mesoproterozoic Era. This preservation-scaled accumulation record is contextualized by additionally plotting ocean-wide deposition of IF. Measured thicknesses of Algoma-type IFs can be as thin as centimeters, but many are tens of meters (see supporting information). Therefore, an ocean floor covered by 25 m of IF is a reasonable estimate to use as a representative thickness of ocean-wide, Algoma-type IFs in Figure 3. We follow Labrosse and Jaupart (2007) and assume that the mean age of the oceanic crust has changed little since Archean time, and therefore equaled the present-day mean age of ~80 Myr. Figure 3 includes the well-dated Chinese and former Soviet Union IFs, corrected using the appropriate time bin from Hurley and Rand but shown with dashed lines to note that the preservation correction does not incorporate the areal record from these regions. The poorly dated Krivoy Rog IF, unexposed Kursk Magnetic Anomaly, and large buried portion of the Ijil IF were again excluded from this figure.

3. Sustained Ocean-Wide IF Deposition

The bias-corrected consensus record (Figure 3b) is more consistent with continual iron- and silica-rich precipitate deposition over the entire ocean from ~3.8 to ~1.7 Ga than with an abrupt increase around 2.5 Ga. The apparent ~2.5 Ga “peak” in IF deposition has previously been interpreted as an environmental signal. Correcting for preservation and exposure biases, however, indicates that this late Archean peak in IF may be an artifact due to the growth and increased preservation of continents and continental shelf areas

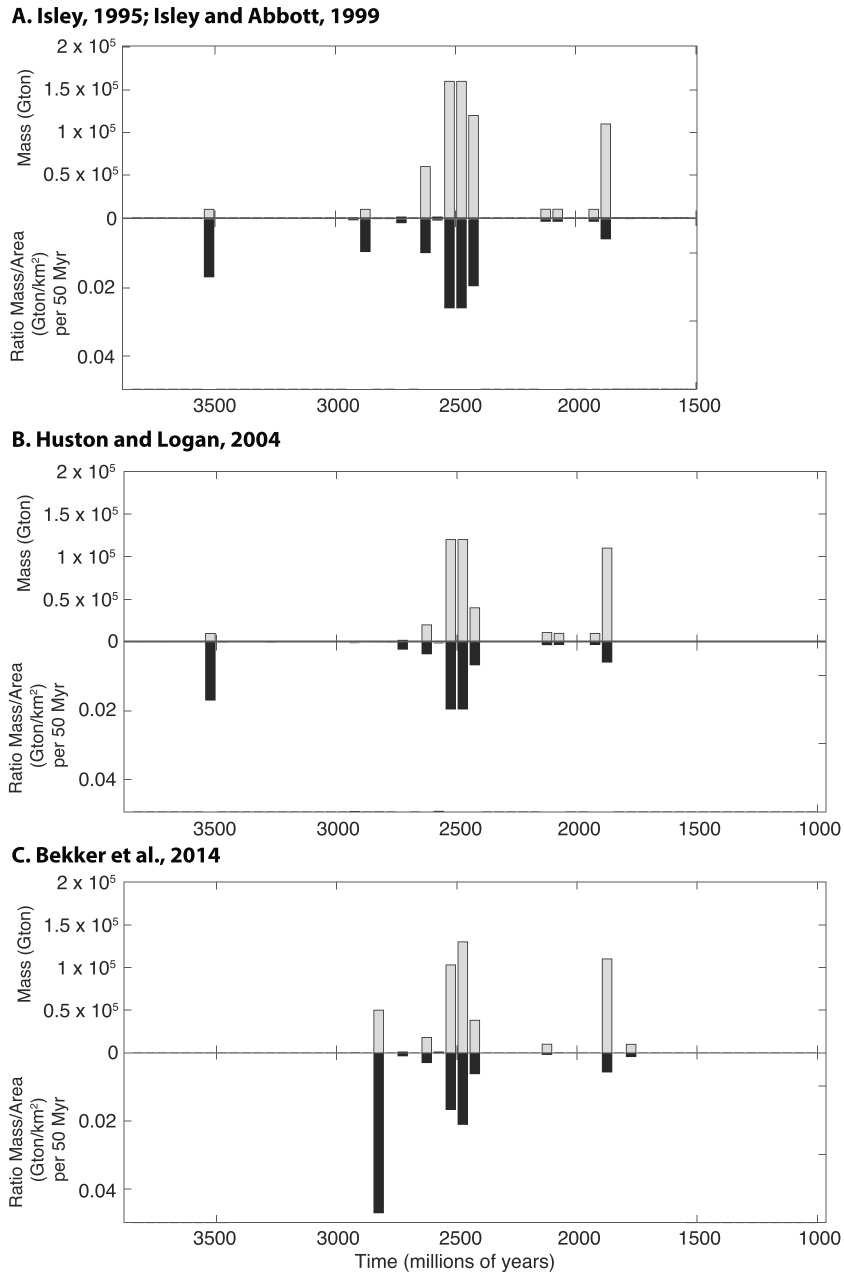

Figure 2. Histograms of iron formation estimated masses (Gt per 50 Myr bin) through time (above, gray), compared to histograms of masses scaled to preserved crustal area (below, black, Gt/km^2 per 50 Myr bin): (a) From Isley (1995) and Isley and Abbott (1999); only to 1.5 Ga because that was the interval considered by Isley and Abbott. (b) From Huston and Logan (2004). (c) From Bekker et al. (2014). Note how preservation-scaling increases the apparent importance of earlier iron formations.

(Figure 1). The normalized IF record in Figure 3b allows an equally plausible hypothesis that extensive long-term IF deposition occurred, with an imperfect geologic record capturing the deeper ocean record in instances of accreted oceanic crust (Algoma-type IFs) and at the margins of growing continental shelves (Superior-type IFs). This preservation-scaled temporal distribution of IF masses likewise suggests that influences of changing chemistry, for example, hydrothermal/mantle plumes of iron (Isley & Abbott, 1999), or biology, for example, related to the evolution of oxygenic photosynthesis and associated release of O_2 (Cloud, 1973), may have been overemphasized, if relevant at all, in IF deposition. The accumulation of iron-rich and silica-rich precipitates appears to have been an important oceanic process as far back as

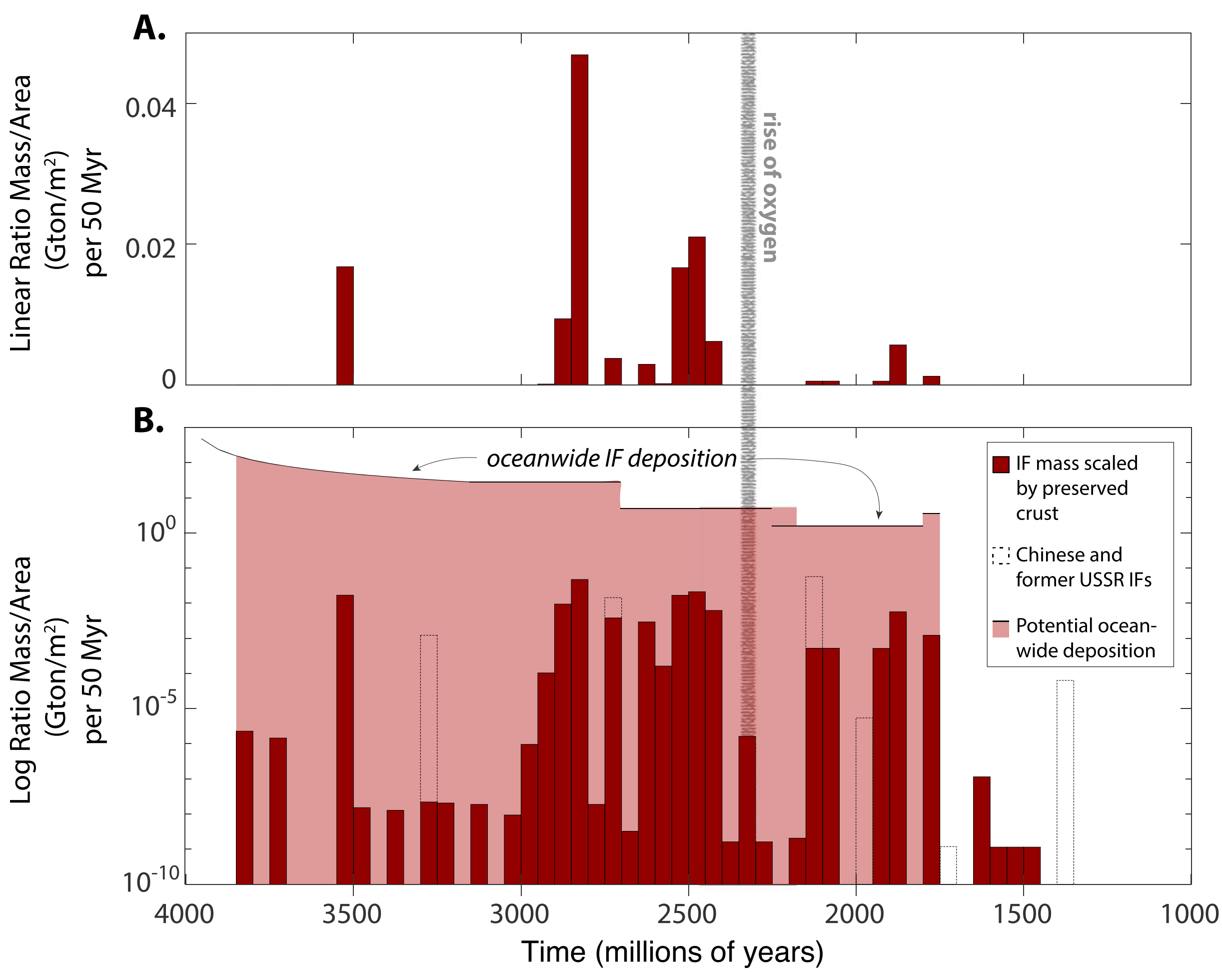

Figure 3. IF accumulation over time, in 50-Myr bins, using previous mass estimates of Isley (1995), Huston and Logan (2004), Bekker et al. (2014), and others (Table S1) and scaled by crustal preservation (see text). (a) Assembled record shown with a linear scale as in Figure 2. The prominent peak at ~2.8 Ga is an artifact of its slightly higher mass estimate and should be disregarded because mass estimates are uncertain by a factor of 3. (b) Log-transformed IF record, suggestive of persistent accumulation of IFs from ~3.8 to ~1.7 Ga. The mass/area ratio expected for 25 m of ocean-wide deposition through time shown for context. The IF record preserves considerably less than this calculation would expect, indicating that, if the IFs were ocean-wide, the entire record has been affected by subduction, deformation, and burial. IF = iron formation; GOE = Great Oxidation Event.

3.5 Ga and possibly 3.8 Ga (Figure 3). Rather than viewing the IF record as reflecting oceanic (bio)chemical changes or the availability of stable depositional basins, we suggest this record could be viewed as a tectonically dismembered and fragmented record of continual deeper water deposition of iron and silica precipitates.

If IF precipitation and deposition did occur across all deep ocean basins starting at 3.8 Ga, then the fluxes of iron and silica into seawater must equal or exceed the depositional fluxes of these elements associated with IF formation. It is generally accepted that marine silica levels were higher in the Precambrian ocean due to the lack of silica-secreting organisms (Maliva et al., 2005). Iron enters the ocean today through continental weathering and hydrothermal systems, such as along mid-ocean ridge vents. Today these vents have variable iron concentrations, from 9 nanomolar to ~25 millimolar (Douville et al., 2002; Seyfried et al., 1991), and the total hydrothermal iron input has been estimated at ~10^{10} kg (or 0.01 Gt) per year (Isley, 1995). In examining the mass balance of modern hydrothermal fluxes of iron, Isley (1995) determined that IF deposition would consume a large fraction of hydrothermal plume iron and that IF deposition near 2.5 Ga needed additional iron sources. Yet ancient hydrothermal vent fluids had a different balance of iron and sulfur, largely due to the lack of oceanic sulfate to reduce to hydrothermal sulfide (Kump & Seyfried, 2005; Walker & Brimblecombe, 1985). With circulating seawater containing low sulfate and little to no oxygen, hydrothermal fluids likely had Fe (II) concentrations as high as 80 mmol/kg (Kump & Seyfried, 2005). Continental weathering sources of iron could also have been higher than today without oxygen to "trap" iron terrestrially in paleosols (e.g., Hao et al., 2017; Rye & Holland, 1998), but it is challenging to estimate Archean continental weathering sources of iron due to the uncertainty surrounding the development and emergence of continents (e.g., Bradley, 2008; Campbell & Davies, 2017).

To determine hydrothermal inputs of ferrous iron into the Archean ocean, we use estimates of the flux of seawater undergoing high-temperature hydrothermal circulation at modern spreading centers combined with estimated concentrations of iron in high-temperature Archean vent fluids. The latent heat released by the formation of crust at mid-ocean ridge systems today is transported by circulating seawater, with approximately 80% of the heat transported by water circulating at low temperatures along ridge flanks and 20% of the heat transported by high-temperature water circulating close to the ridge axis (German & Seyfried, 2014). This high-temperature fluid flux is estimated to be $0.17–2.93 \times 10^{13}$ kg/year using thallium isotopes (Nielsen et al., 2006). With higher Archean mid-ocean ridge temperatures (e.g., Herzberg et al., 2010; Sleep & Windley, 1982), the high-temperature fluid flux was likely at least on the upper end of today's range, at ~3×10^{13} kg/year. Applying estimated pre-GOE hydrothermal iron concentrations of 80 mmol/kg (Kump & Seyfried, 2005), hydrothermal fluids would introduce 0.134 Gt of Fe (II) per year (see supporting information).

Could this hydrothermal source flux supply sufficient iron for ocean-wide precipitation of IF precursors? Commonly used depositional rates for IFs come from the two of the largest Superior-type IF sequences: the Hamersley Group of Australia and the Transvaal Supergroup in South Africa. Assuming that 0.2 to 1.6 mm-scale laminae in Hamersley Group IFs represented varve-like annual depositional cycles, Trendall and Blockley (1970) estimated a high depositional rate for IFs of 227 m/Myr. Ewers and Morris (1981) later suggested an even higher rate of 893 m/Myr. Geochronological constraints of the Hamersley Group are consistent with a depositional rate as low as 7 m/Myr (Trendall et al., 2004); Trendall et al. preferred 180 m/Myr, not accounting for any depositional hiatuses. Others have proposed that many sedimentary hiatuses are recorded by the bedded cherts in IFs (Krapež et al., 2003). Much lower IF depositional rates in the Transvaal Supergroup, at 2–6 m/Myr, have been inferred from geochronological constraints (Barton et al., 1994; Pickard, 2003). If IFs accumulated in deep water across Archean ocean basins as well as along continental margins, depositional rates of IF chemical precipitates may indeed have been quite low, similar to modern pelagic deposition accumulating ~1 to 6 m/Myr (Müller & Suess, 1979; Sugisaki, 1984). If average depositional rates of 1 m to 180 m/Myr characterized the entire ocean with a surface area of 4×10^{14} m^2 (78% of Earth's total surface area), endmember bounds on the IF iron sink would be ~0.36 to 65 Gt Fe per year.

The high depositional rates proposed by Trendall and others for continental shelf regions (Ewers & Morris, 1981; Trendall & Blockley, 1970) would not be consistent with extensive and long-term deposition, as Isley (1995) determined. Ocean-wide depositional rates similar to modern pelagic deposition, however, come closer to an equilibrium with the hydrothermal input of Fe (II) calculated above. The lack of detrital material in IFs is also consistent with slow sedimentation in abyssal plains. It is possible that limited areas such as the Hamersley Basin and other continental shelves had higher depositional rates, perhaps stimulated by Fe (II) upwelling, but those rapid rates could not be responsible for Algoma-type IFs spanning the Archean ocean. We can estimate a depositional rate for an average IF thickness, consisting of 30% Fe, covering the entire ocean basin in the period of time given by the mean age of the ocean floor. Ocean floor spanning 4×10^{14} m^2 and accumulating 25 m of IF sediment with a density of 3×10^3 kg/m^3 (Trendall & Blockley, 1970) in 80 Myr would remove Fe from the ocean at 0.11 Gt Fe/year (see supporting information), comparable to the hydrothermal Fe input of 0.134 Gt/year calculated above. With additional iron fluxes from continental weathering and higher hydrothermal fluxes due to shallower and hotter Archean spreading centers (Sleep & Windley, 1982), the potential ocean-wide IF accumulation rate could reach 1 m/Myr (0.36 Gt Fe/year) or even higher. Although depositional rates of 180–893 m/Myr are not feasible at this ocean-wide scale, a low IF sedimentation rate of ~1 m/Myr would also help explain the perplexing lack of organic matter in IFs (Klein & Beukes, 1989) by allowing extensive organic degradation to occur in slow-accumulating sediments during diagenesis. This post-depositional processing could also rerelease ferrous iron back into seawater and act as an additional iron flux to balance the Archean iron cycle (Konhauser et al., 2005).

Sustained and ocean-wide accumulation of IF precipitates would make IFs almost certainly the largest sink for iron, and perhaps also for silica. Therefore (bio)geochemical processes forming these iron and silica precipitates would have acted as major controls on elemental cycling for over a billion years, from ~3.8 to the ~2.3 Ga GOE, if not later. This apparent availability of iron throughout much of the Archean supports hypotheses of a long-term Archean ferruginous ocean (e.g., Poulton & Canfield, 2011). Persistent IF deposition, moreover, suggests that any organisms relying on iron for either enzymatic requirements or iron-based (iron-oxidizing or iron-reducing) metabolisms could have flourished over this entire ~1.5-Ga time period. With no obvious change in IF deposition throughout this pre-GOE time, the IF temporal record appears less like a growing titration of oxygen and more consistent with over a billion years of mineralization processes unrelated to oxygen. Likewise, pervasive and widespread IF deposition implies that either IFs formed independently of biological influences or the necessary IF-generating organisms were prevalent beginning at ~3.8 Ga until the GOE.

After the GOE, deposition of IFs continued for ~600 million years (Figure 3) but with marked differences in IF sedimentology. The thinly laminated, mud-sized chemical precipitates of pre-GOE IFs evolved to include thick units of cross-bedded, sand-sized granular clasts and ooids of iron-rich minerals (Clout & Simonson, 2005; Simonson, 2003). These sedimentological observations indicate that the depositional environment(s) for IFs moved from generally deeper sedimentation below wave base prior to the GOE to more shallow-water high-energy environments after the GOE, suggesting the locus of iron precipitation migrated to shallower water depths (Clout & Simonson, 2005; Simonson, 2003). The decline of sustained IF deposition after two billion years, beginning ~1.7 Ga, may relate to an increase in oceanic sulfate (Kump & Seyfried, 2005), sulfide (Anbar & Knoll, 2002; Canfield, 1998), or environmental oxygen (Holland, 2006; Huston & Logan, 2004; Sleep & Bird, 2008). Although few IFs are known between 1.7 and 1.0 Ga (Figure 3), IF deposition did not completely stop and strata from this period may yet yield more evidence of infrequent but significant iron precipitation, such as the ~1.4 Ga Xiamaling IF (Canfield et al., 2018).

4. Conclusions

The two-billion-year, preservation-corrected history of iron- and silica-rich precipitation and deposition (Figure 3) implies that IFs record a persistent and stable (bio)geochemical marine iron and silica cycle. The temporal pattern of IFs may not have been limited by depositional environments (Trendall, 2002), nor by influxes of ferrous iron (Isley & Abbott, 1999), and need not reflect an interplay of tectonic, chemical, and biological factors (Bekker et al., 2010; Clout & Simonson, 2005). Instead, IFs may chronicle a widespread signal that was subsequently obscured and dismembered to form an incomplete record on Earth's surface today. Rather than focus on the ~2.5 Ga peak in IF volumes, which may be an artifact of continental growth and preservational biases, future research might do well to concentrate on understanding the features of pre-GOE IFs and how they changed post-GOE, including the reason(s) for (1) the deposition of IFs for ~1.5 billion years prior to the GOE, (2) the continued deposition of IFs after the GOE for hundreds of millions of years, and (3) the decrease, and eventual demise, of IFs after ~1.7 Ga.

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