GENESIS OF LARGE PRECAMBRIAN IRON FORMATIONS

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Iron formations are a distinctive type of chemical sedimentary rock that are high in iron and silica and low in alumina. They are found in the oldest sedimentary successions on earth and persist as a widespread constituent of many volcanic and sedimentary successions until they disappear rather abruptly at or before 1.8 Ga. The only well-documented interval of global iron formation deposition postdating the Paleoproterozoic took place late in the Precambrian. Examples are found on all of the major cratons, but these late Precambrian iron formations "differ substantially in both character and bulk composition" from early Precambrian iron formations (James and Trendall, 1982, p. 207). Most notably, they tend to be richer in iron, closely associated with glacial deposits, and have a simple iron mineralogy dominated by oxides. Consequently, these late Precambrian iron formations may differ in origin from the early Precambrian ones and are not included in the summary that follows.

Few workers now dispute the marine origin of iron formations because of their uniform chemical and mineralogical composition (Gole & Klein 1981), their conformable contacts with definite marine units (Ojakangas 1983; Beukes 1983; Simonson 1985), and the sheer size and lack of variability of the largest ones. Moreover, paleocurrents in iron formations show complex polymodal patterns typical of shallow marine sands. Iron formations are the source of most large iron ore deposits, and the largest ore deposits are hosted by the uniquely large iron formations laid down during an interval of ≤ 800 million years in the Late Archean to Paleoproterozoic. The creation of these huge repositories of iron was probably the result of a unique conjunction of circumstances, as follows.

Undeformed iron formations can be divided into GRANULAR iron formations (**GIFs**) or BANDED iron formations (**BIFs**) on the basis of sedimentary textures. Most GIFs are well-sorted chemical sands in which the clasts (referred to as "granules") were derived intrabasinally via erosion and redeposition of pre-existing BIF. BIFs are by far the more abundant of the two and originally consisted of chemical muds instead of sands. Although they consist of a variety of different iron minerals, most BIFs exhibit a quasi- to highly rhythmic alternation of layers rich in iron and silica on a scale of millimeters to centimeters. The similarities of BIFs in all contexts suggest they originated via a common mechanism. Although the details of this mechanism have yet to be agreed upon, it is widely believed that the iron came from hydrothermal sources in the deep ocean. The link between hydrothermal activity and BIF deposition has been established with reasonably high confidence for some 'Algoma-type' BIFs via stratigraphic context and facies relationships. Various geochemical signatures also indicate a hydrothermal source for the larger BIFs (Klein & Beukes 1992). It is also widely believed that the iron minerals deposited in larger iron formations were originally precipitated along chemoclines in a stratified ocean (Beukes & Klein 1992; Simonson & Hassler 1996). This would help explain why iron formations were deposited in many different tectonic settings and are associated with a wide variety of rock types (Gross 1983; Blake & Barley 1993; Fralick & Barrett 1995). Isley (1995) has outlined a plausible scenario whereby iron dissolved in plumes released from deep-sea hydrothermal systems could travel long distances and reach distant depocenters in the early Precambrian. Recent isotopic work also indicates microbes played a role in the precipitation and/or diagenetic reorganization of the iron in iron formations (Beard *et al.* 1999).

Traditionally, the large iron formations of the Late Archean and early Paleoproterozoic have been referred to as 'Superior-type' and contrasted with so-called 'Algoma-type' iron formations, which are thought of as generally smaller and typical of older successions. In his original definition, Gross (1965) also stated 'Superior-type' iron formations are GIFs characterized by granules and oolites and associated with shallow-water strata such as quartzites and dolomites, whereas 'Algoma-type' iron formations are BIFs and associated primarily with volcanic rocks and deeper-water flysch-type graywackes and slates. Although this subdivison accurately reflects a secular change in the nature of iron formations through geologic time, not all large iron formations display granular textures or show a close association with shallow-water deposits. The large iron-formations of the Hamersley Group, for example, are true BIFs (Trendall & Blockley 1970; Trendall 1983) and are interbedded with turbidites and other deeper water strata (Simonson *et al.* 1993; Hassler 1993; Simonson & Hassler 1996). Gross (1983) subsequently acknowledged his two types are end members of a continuum and that some of the large iron formations were deposited in deeper water on continental shelves and slopes.

If all iron formations have the same source of solutes and precipitated via similar processes, an increase in the average size of iron formations through time could result from either 1) an increase in the average size of the depositional basins available, or 2) some change in the nature of ocean basins that permitted iron to be dispersed over greater distances. All of the large iron formations are underlain by continental basement, and many were deposited in passive margin settings, yet coincide with stratigraphic evidence of accelerated subsidence (Hoffman 1987; Simonson & Hassler 1996). Extensive shelf environments only became widely available after a surge in the growth of continental crust in the Late Archean (Goodwin 1991; Lowe 1992; Eriksson 1995). Therefore the first appearance of extensive areas of seafloor on continental shelves is probably one of the main reasons that the first large or 'Superior-type' iron formations appeared around 2.6 Ga. The difference in the ages of the largest iron formations on different continents could reflect the fact that the 'cratonization' of shields was a highly diachronous process (Eriksson & Donaldson 1986; Eriksson 1995). A connection between iron formation deposition and mantle superplumes could also help explain why 'Superior-type' iron formations do not appear to be evenly distributed in either time or space. Isley and Abbott (1999) recently documented a statistically significant correlation between the isotopic ages of iron formations and episodes of mantle plume activity detected via proxies such as komatiites and flood basalts.

It is also possible that the growth in the average size of iron formations during the early Precambrian was made possible by changes in the bathymetry and/or chemistry of the deep ocean which permitted the dispersal of iron over longer distances. Paradoxically, this would require a greater mobility of dissolved iron in Late Archean and early Paleoproterozoic oceans than it had earlier in the Archean. This runs counter to the prevailing view that the concentration of free oxygen in the atmosphere was increasing throughout this time period (Eriksson 1995). If this was a factor, either the concentration of free oxygen in the atmosphere was actually decreasing during this time interval, or any increase in atmospheric oxygen was more than offset by other factors such as changes in the circulation patterns and/or water chemistry of the deep ocean.

In addition to an increase in the average size of iron formations, there is also an increase in the abundance of GIF relative to BIF with geologic time. Only a handful of GIFs have been reported in 'Algoma-type' iron formations (Manikyamba 1999), and the iron formations deposited on the margins of the earliest continental nuclei are still pure BIFs, e.g., on the Kaapvaal and Zimbabwe Cratons (Eriksson 1979; Watchorn 1980; Fedo & Eriksson 1996). The majority of the older 'Superior-type' iron formations are pure BIFs deposited in deeper water settings such as distal shelf environments. The fine grain size and extreme lateral continuity of some of the layers in the Hamersley BIFs (Ewers & Morris 1981), for example, indicates they were deposited under "exceptionally still and quiet" conditions (Trendall 1983, p. 118-119). Nevertheless, the Hamersley Group contains a few thin beds of GIF-like chert (Simonson & Goode 1989), and the roughly coeval Griquatown Iron Formation of the Transvaal Supergroup (Beukes 1984; Beukes & Klein 1990) is the first true GIF. GIFs reached their acme ca. 2.0 Ga with the iron formations of North America (Morey 1983; Gross and Zajac 1983; Fralick & Barrett 1995) and the Nabberu Basin of Western Australia (Goode *et al.*, 1983; Bunting, 1986). These iron formations also contain oolitic and stromatolitic layers, but they are not volumetrically major components. In fact, GIF is subordinate to BIF even in the most GIF-rich 'Superior-type' iron formations. Nevertheless, the locus of iron formation deposition clearly shifted into progressively shallower environments through time until it ceased altogether at or before 1.8 Ga. Like the increase in size of 'Superior-type' iron formations, this increase in the relative abundance of GIF probably signals an expansion in the total area of continental shelves during the Late Archean and Paleoproterozoic. The migration of 'Superior-type' iron formation depocenters into shallow water could also signal a progressive shallowing of the global chemocline during this time interval.

Iron formations also differ from younger iron-rich sediments in having a high silica content. This is attributed to Precambrian seawater having higher silica concentrations than Phanerozoic seawater due to the absence of silica-secreting organisms (Maliva *et al.* 1989). However, much of this silica was added post-depositionally in the form of early cement. As well-sorted sands, GIFs' depositional porosities were ca. 40-50%, and much of this porosity was occluded by siliceous cement prior to compaction. The primary nature of this cement is indicated by competitive growth textures typical of void-fills, as well as its presence in intraclastic pebbles of GIF in some iron formations. Siever (1992) suggested the cements originated when silica diffused from silica-rich seawater into sediment porewaters at ambient seafloor temperatures, which would require concentration gradients that are the opposite of today's. Higher geothermal gradients could have also aided in the early diagenetic reorganization of silica in the shallow subsurface by accelerating dissolution and reprecipitation (Simonson 1987).

Early silica cements in GIFs are generally concentrated in oblate mottles or patches that are elongated parallel to stratification. Many GIFs also have relatively large cavities, cracks, and vugs filled largely with siliceous cements and lesser volumes of internal sediment. These features are part of a continuum that begins with small cracks confined to individual granules and believed to be true syneresis cracks, i.e., formed via shrinkage due to the dewatering of a gelatinous silica precursor (Gross 1972; Dimroth & Chauvel 1973; Beukes 1984). Some of the largest cracks in GIFs cut indiscriminately across both granules and cements, indicating the cements also shrank. Other cavities are stratiform, and the larger ones contain geopetally distributed internal sediment which is locally cross-laminated (Simonson 1987). Similar cracks and vugs developed in stromatolitic cherts in GIFs even contain evidence of cavity-dwelling microbes (Simonson & Lanier 1987). The fact that these cracks are in cemented sands (GIFs) rather than former muds (BIFs) indicates they are not mudcracks formed via subaerial desiccation. These features appear to be unique to iron formations and indicate that they are primary siliceous deposits. Many researchers have suggested iron formations formed via the silicification of sediments which were drastically different composition, e.g. carbonates or evaporites, "but the evidence against this concept is now so overwhelming ... that diagenetic replacement is no longer ... viable" (Kimberley 1989).

On average, GIFs appear to be more siliceous than BIFs. In part, this probably reflects primary depositional gradients. Oolitic and stromatolitic layers are frequently chert-rich and contain some of the lowest iron concentrations of any units in iron formations. Likewise, the younger iron formations with higher abundances of GIF tend to have lower overall concentrations of iron than the older, deeper-water iron formations. All of these observations are consistent with a deep-ocean source of iron, but it may also reflect a greater abundance of early silica cement in GIFs relative to BIFs. However, BIFs also show evidence of early cementation. Silica-rich bodies that are geometrically similar to the mottles in GIFs are widespread in BIFs, where they are generally referred to as chert pods. These pods have geometries identical to early-cemented concretions precipitated in fine-grained sediments of other compositions prior to compaction, e.g. carbonate concretions in shales. The ratio of the inside and outside thicknesses of individual layers that pass continuously through chert pods can be used to estimate the minimum amount of compaction. In BIFs, this ratio ranges as high as 14:1 (Simonson 1997), indicating the present thickness of such layers is < 10% of what it was prior to compaction. This in turn indicates the original precipitates that made up the proto-BIFs were highly porous, as is typical of muds in general. The textures and minerals shielded within these early-cemented volumes of BIF and GIF indicate proto-iron formations had a range of compositions not so different from what we find today. They did not, however, contain coarsely crystalline phases such as magnetite, which must have developed later in diagenesis.

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