Textures, Composition, Distribution and Diagenesis of Non-Clastic: Ferruginous

Almost all sedimentary rocks are iron-bearing in the sense that mudrocks, sandstones, and carbonates typically have an iron content of several percent. Nevertheless, sedimentary rocks in which the proportion of iron exceeds 15 percent are separately categorized as iron-rich.

Two major types of iron-rich sedimentary rocks are recognized, Iron formation, or Banded Iron Formation (BIF) these are Precambrian age and Ironstones, those of the Phanerozoic age.

There are important differences between those that formed in the early middle Precambrian and those of the Phanerozoic.

Iron formation, or banded iron formation (BIF)— are regionally extensive, locally thick sequences composed of alternating thin (millimetre to centimetre thick) layers of mainly crystalline-textured iron-rich minerals and chert.

Ironstone— are noncherty, essentially clastic-textured, iron-rich minerals of local extent. have a distinctly lower content of silica, and typically have oolitic textures.

Banded Iron Formation (BIF)

Banded-iron formation (BIF**)** are chemically precipitated sediment, typically thin bedded or laminated, consisting of 15 percent or more iron of sedimentary origin and layers of chert, chalcedony, jasper, or quartz. They are composed of alternating layers of iron-rich material and silica as you can observe in this photograph of BIF. Iron rich bands are normally composed of hematitie (Fe_2O_3), magnetite (Fe_3O_4), siderite ($FeCO_3$) or pyrite (FeS_2). The iron poor bands normally contain chert (fine-grained quartz) with lesser amounts of iron oxide.

Such formations occur on all the continents and most are between 1.8 to 2.2 billion years old; none are younger than Cambrian age. but range in age from early Precambrian to Devonian. Three periods of peak deposition are recognized: mid Archean (3400–2900 my), early Proterozoic (2000–2500 my), and late Proterozoic (750–500 my).

These deposits constitute the world’s major source of iron ore.

On the basis of relative abundance of major iron- bearing minerals, James (1966, 1992) defined four principal kinds of iron-rich sedimentary rocks, which he referred to as facies, the classis are:

Oxide facies: comprising of minerals dominantly Magnetite and Hematite

Silicate facies: comprising of minerals such as Cummingtonite, Grunerite, Greenalite, Minnesotaite, Stilpnomelane, Fe Chlorites, And Fe Amphiboles.
Carbonate facies: comprising of minerals such Ankerite and Siderite.
Sulphide facies: dominated by Pyrite.

Banded iron formation has also been classified based on their geologic setting the two main classes are a) Algoma-type and b) Superior-type
Algoma-type: these are dominantly Archaean though younger Algoma type iron ores are also known. These were formed by turbidity currents near island arc trenches and typically sub marine volcanogenic origin, Island arc depositional environment. They are thinly banded or laminated, lack oolitic or granular textures, and commonly extend laterally for only a few kilometres. Important examples of these types can be found in the Canadian Shield and Western Australia.

Superior-type: these are Early Proterozoic age formed between 2.7 and 1.8 billion years ago, they are Shelf sediments precipitated in stable shelf or marginal basin (foreland or back-arc basins), continental margin depositional environment. These not associated with volcanic rocks but commonly occur with quartzite, black carbonaceous shale, conglomerate, dolomite, massive chert, chert breccia, and argillite. They contain granular and oolitic textures, and commonly extend laterally for hundreds of kilometers. Important examples of these type can be found in the Lake superior deposits of Michigan and Minnesota, Hamersley Basin of Australia.
Raritan type deposits are a third kind of iron-formation. These deposits occur in several basins of late Precambrian age and are associated in part with glacial deposits.

Origin of Banded Iron Formation

The origin of banded iron formation is not clearly understood Because there is no modern analog for the iron-rich sediments, depositional models for these sediments have to be formulated mainly on the basis of the ancient record. Geologists with access to the same data may draw different inferences from these data; therefore, numerous models for the origin of iron-rich sedimentary rocks have been advanced – none of which has gained complete acceptance.
Points of discussion with the origin of precambrian iron formation centre on for example, Why, are cherty, banded iron-formations confined primarily to the Precambrian? Why are iron-formations and ironstones not forming today? What is the source of the vast amount of iron locked up in the known iron-rich deposits? Was volcanism involved in some way in the formation of some or all of these iron deposits? How was iron transported from its source to the depositional site, and what was the depositional environment in which iron-rich sediments accumulated?
Controversy also continues over the ultimate source of iron (weathering as opposed to magmatic iron escaping from Earth’s interior) and over the possible role of microorganisms such as bacteria and algae in the precipitation of the iron.

Many sedimentary petrologists consequently conclude that banded iron formation deposition is a uniquely Precambrian occurrence made possible by, and supporting the existence of, an earlier anaerobic Earth atmosphere (one lacking free oxygen) quite unlike that in existence today.
Iron has two main oxidation states
Fe^{2+} (ferrous)
Fe^{3+} (ferric)
Fe^{2+} is soluble in water, but is rapidly oxidized to Fe^{3+} in the presence of oxygen.
Fe^{3+} is insoluble. It flocculates and settles out. Thus: The deposition of iron-rich rocks is facilitated by the presence of O_2.
During the early Archean, O_2 was present only in trace quantities, so ferrous iron could exist in solution in the oceans. Indeed, because the contemporary atmosphere was rich in CO_2, fresh and ocean waters were probably more acidic, facilitating the release of iron through weathering. Beginning ~3.0 ga, photosynthesizing cyanobacteria began releasing O_2 into the oceans. At the same time, we begin to see BIFs - finely interbedded cherts and iron-rich mudrocks.
The exact mechanism of deposition is enigmatic, but seems related to scavenging of ferrous iron by free oxygen in the oceans.
The iron constituted an oxygen sink, binding oxygen into oceanic sediment until they were saturated at ~1.8 ga. After 1.8 ga, little dissolved iron was left in the oceans and the formation of BIF essentially stopped.

Ironstones

After oceanic oxygen sinks were saturated, O_2 could begin to accumulate in the atmosphere, allowing the rise of iron oxide minerals like hematite and goethite in terrestrial environments. Major iron mineral deposits are shallow marine, in which these minerals form ooids (sand-sized concretions) around a mineral nucleus.
Phanerozoic ironstones vary considerably in grade, lithology and iron minerals present. Ironstones form thin, massive, or poorly banded sequences, a few meters to a few tens of meters thick, in sharp contrast to the much thicker, well-banded iron-formations. E.g. of classical ironstone deposits can be found in Ordovician Wabana Formation of Newfoundland and the Silurian Clinton Group of the central and southern Appalachians.

The most important are the oolitic ironstones, which generally are hematite–chamosite in the Palaeozoic and goethite–berthierine in the Mesozoic. Of less significance are the siderite mudstones and sulphide ironstones. Major iron minerals are goethite, hematite, and chamosite. Nonferrous minerals in ironstones may include detrital quartz, calcite, dolomite, authigenic phosphorite, and authigenic chert.
They generally have an oolitic texture, and they may contain fossils that have been partly or completely replaced by iron minerals.
Ironstones are commonly interbedded with carbonates, mudrocks, and fine-grained sandstones of shelf to shallow-marine origin.