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Contact metamorphic effects of the Stillwater Complex, Montana: the concordant iron-formation: a reply to the role of buffering in metamorphism of iron-formation

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We stated in Vaniman, *et al.* (1980), "We propose that the low fO_2 and the loss of internally buffered fO_2 variability in contact-metamorphosed iron-formations is due to redox interaction with the adjacent intrusions." Frost (1982) takes issue with this statement and considers that all iron-formations were metamorphosed under closed system conditions such that the mineral assemblages had buffered fluid compositions. Here, we shall clarify our statement and, in the spirit of that proposal, suggest a mechanism whereby contact-metamorphosed iron-formation could have interacted with the gabbroic heat-source.

First, let us consider whether the rock system of the Stillwater Iron-Formation was open or closed to oxygen. That is, did oxygen behave as a boundary-value or initial-value component (Zen, 1963)? If the iron-formation was open to oxygen, then the value of fO_2 during metamorphism should have been constant over a relatively large volume of rock. The values of fO_2 are recorded by the assemblages, and, for example, a rock volume metamorphosed under the externally imposed fO_2, T conditions of B in Figure 1 will contain one of the assemblages Opx + Q + M, Opx + Ol + M, Q + Opx, Ol + Opx, Opx + M, or Ol + M depending on the ratio of Fe:Mg:Si. All rocks containing the assemblage Opx + Q + M will have the same composition of orthopyroxene. Changes in T, at constant fO_2, result in changes in the compositions of olivine and orthopyroxene that coexist with quartz and magnetite because of the temperature dependence of the equilibrium constants for the various continuous reactions. If, on the other hand, small volumes of rock were closed to oxygen during metamorphism, assemblages in rocks metamorphosed at identical temperatures should have equilibrated at a variety of values of fO_2 reflecting the various ratios of Fe^{3+}/Fe^{2+} in the sedimentary protolith. The assemblages Opx + Q + M, Ol + Q + M, Ol + Opx + M, Opx + M, and Ol + Q, each representing a different value of fO_2, can be intimately associated, and the buffer assemblage Ol + Opx + Q + M can be common in a large region over which a temperature gradient existed. Furthermore, rocks that contain the same assemblage, for example Opx + Q + M, but that equilibrated within a range in fO_2 have pyroxene compositions that range in Mg/Fe, as described by Frost. The common occurrence of regional metamorphic terrains that exhibit closed-system behavior with respect to oxygen has long been recognized by Eugster (1959) and Thompson (1972).

Knowing the status of the boundary of the system with respect to oxygen, predicting the effects of metamorphism of a particular bulk composition is relatively easy, as we have done above. But con-

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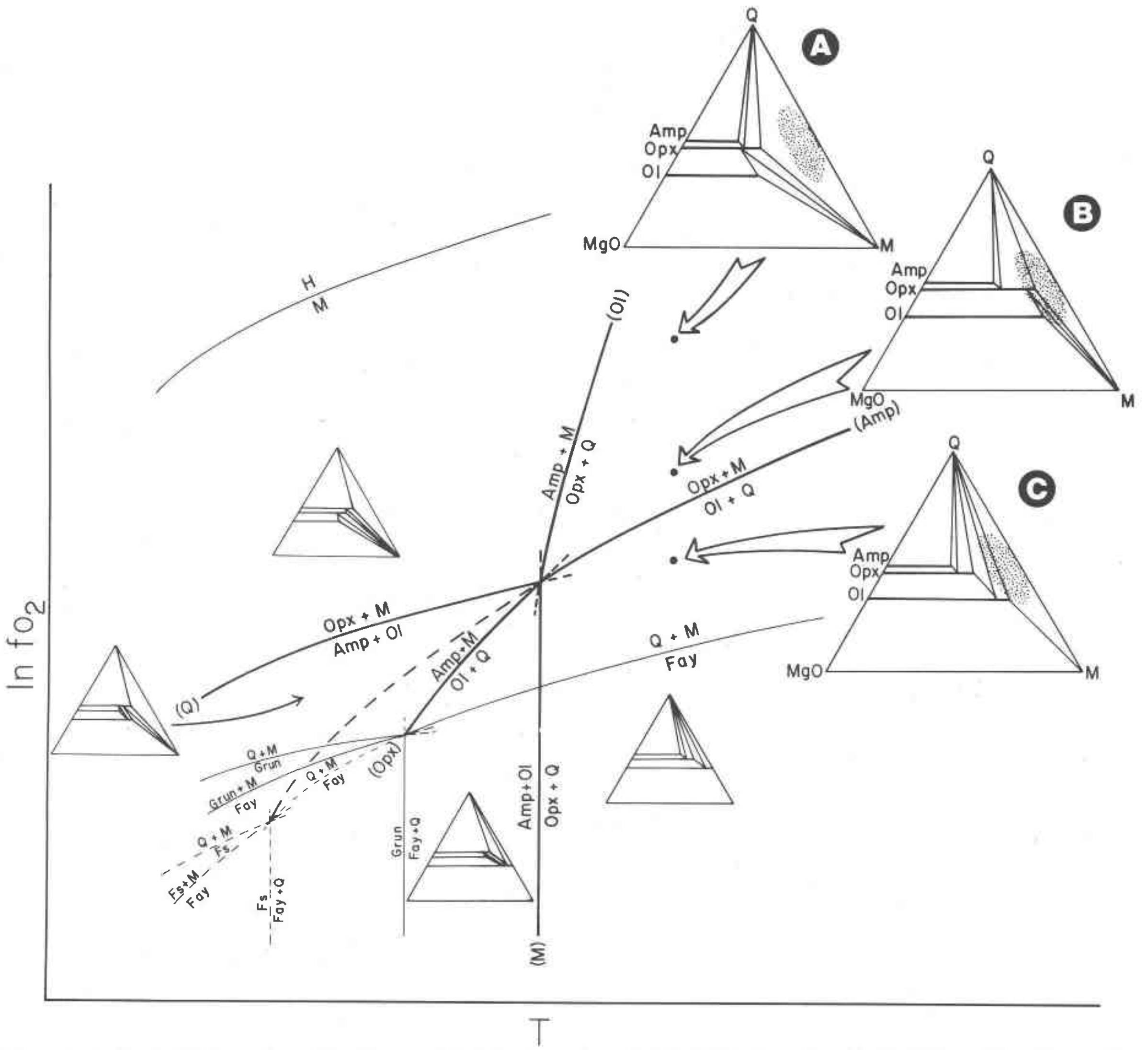

Figure 1. Isobaric T–fO_2 section of the phase equilibria in the system Fe–Mg–Si–O–H pertinent to the high-grade metamorphism of iron-formation at constant p_{H_2O}. Abbreviations are Q for quartz, M for magnetite, Amp for amphibole, Opx for orthopyroxene, Ol for olivine, Fay for Fe-end-member fayalite, Grun for Fe-end-member grunerite, Fs for end-member ferrosilite, and H for hematite. Facies-types for the T–fO_2 conditions indicated by A, B, and C are constructed from the observed distribution of Fe and Mg among coexisting phases: $(Fe/Mg)_{Amp} < (Fe/Mg)_{Opx} < (Fe/Mg)_{Ol}$ (Bonnichsen, 1969; Immega and Klein, 1976; and Labotka, unpublished data). Light lines are equilibria in the Fe-end-member system. Univariant equilibria in the Mg-bearing system are shown as heavy lines. Metastable equilibria are shown as dashed lines. Univariant equilibria are labeled by the absent phase, enclosed in parentheses. The equilibria (Opx) and (Amp) terminate at the invariant points for the corresponding pure-Fe systems (Thompson, 1975). The three facies-types for the conditions A, B, and C show compatibilities for an isothermal decrease in fO_2. Facies types A and B show the same mineral assemblages, but for the more oxidized conditions of A the compositions of Opx coexisting with Q and M is more Mg-rich than in the same assemblage under the more reducing conditions of B. Similarly, the assemblage Opx + Ol + M occurs in both A and B, but Ol and Opx in A are more Mg-rich than in B because an increase in fO_2 causes the continuous reaction $6\ Ol + O_2 = 6\ Opx + 2\ M$ to be driven to the right. The stippled region represents the range in Mg:Fe:Si for the compositions of unmetamorphosed iron-formations reported by James (1954). Under the oxidizing conditions of A, this range in bulk composition falls entirely within the field of Opx + O + M. Under the more reducing conditions of B and C, much of the range falls in fields represented by olivine-bearing assemblages.

fronted with the alternative of knowing the metamorphic effects, determining the boundary conditions can be a frustrating exercise, especially if the range in bulk composition and, thereby, the number of different assemblages are small. Most of the rocks in the Stillwater Iron-Formation equilibrated under the conditions illustrated by B in Figure 1 giving rise to the assemblages Opx + Q + M and Ol

+ Opx + M. Olivine and orthopyroxene in these assemblages have narrow ranges of composition indicating a relatively narrow range in fO_2 during metamorphism. This observation is consistent with, but does not prove, the boundary-value status of oxygen. In fact, one sample contains the assemblage Ol + Q + M which represents lower fO_2 values (C in Figure 1) than does the assemblage Opx + M from the same locality.

Similar relations occur in the contact metamorphosed Biwabik Formation, described by Bonnichsen (1969, 1975) and Morey *et al.* (1972), in which the assemblage Opx + Q + M is common, the pyroxene in this assemblage has a wide range in composition, the buffering assemblage Ol + Opx + Q + M commonly occurs, and the assemblage Ol + Q + M is present in drill core samples. Clearly, even within the range in temperatures that these samples experienced, the oxygen fugacity varied considerably. Indeed, Morey *et al.* (1972) indicated a bulk compositional control of fO_2 within the drill-core samples that they studied.

Contact metamorphism of iron-formation appears to have occurred under conditions such that oxygen behaved as an initial-value component. But, an intriguing aspect of metamorphosed iron-formations was mentioned by Eugster (1959) who noted the virtual absence of olivine from the regionally metamorphosed iron-formations of northern Michigan, which are the stratigraphic equivalents to the Biwabik and Gunflint Iron-Formations. Klein (1973) also remarked in a synthesis of metamorphism of iron-formations that olivine-bearing assemblages were described from only contact-metamorphic occurrences, although Gole and Klein (1981) described local occurrences of Ol + Q + M assemblages in regionally metamorphosed iron-formation from Western Australia. Contact-metamorphosed iron-formations clearly represent lower *average* fO_2 conditions (B and C in Figure 1) than regionally metamorphosed iron-formations (A in Figure 1) which also commonly contain hematite-bearing assemblages. Yet, both types of metamorphosed iron-formations exhibit effects of closed-system behavior with respect to oxygen fugacity. Did those iron-formations that were ultimately contact-metamorphosed fortuitously have lower average Fe^{3+}/Fe^{2+} initial ratios than those that were ultimately regionally metamorphosed? This is unlikely. In fact, the similarity of the fO_2 values represented by contact-metamorphosed iron-formations to the fO_2 values of the intrusive gabbros suggests interaction. We amplify our earlier proposal by suggesting a mechanism for this interaction.

In the Gunflint Iron-Formation, Floran and Papike (1978) provided textural evidence that indicated recrystallization of siderite and greenalite directly to orthopyroxene and olivine, overstepping intermediate minnesotaite- and grunerite-producing reactions. Adjacent to intrusions, the rate of change in temperature is greatest in a short interval following emplacement. For example, the temperature of the country rock adjacent to the contact of a 1200° C intrusion will rise to ~700° C almost instantly after intrusion. The actual values of temperature and rate of change in temperature depend on geometry, mechanism of heat transfer, and initial conditions (Carslaw and Jaeger, 1959); the occurrence of pigeonite in the iron-formation indicates that the temperature was locally in excess of 805° C (Vocke, 1981) in the Stillwater Iron-Formation. This rapid rise in temperature and concomitant overstepping of low-temperature reactions results in nearly complete devolatilization of the iron-formation. Floran (1975) showed that a large rate of devolatilization produces a solid volume decrease of 10 to 20%, and Vaniman *et al.* (1980) provided textural evidence for a similar volume change and for deformation during metamorphism of the Stillwater Iron-Formation. This large change in volume initially causes an increase in permeability by the creation of pores and cracks through which the escaping volatile phase can equilibrate with the reducing gabbroic intrusions. The tendency toward equilibration with the gabbro can result in reduction of Fe^{3+}/Fe^{2+} of the iron-formation. But, because the release of fluid by the iron-formation is dynamic and because the resulting porosity is inhomogeneous, which reflects both inhomogeneity in bulk composition and bedding-plane anisotropy of the iron-formation, complete and volumetrically uniform equilibration of the iron-formation with the gabbro is unlikely to occur. In addition, confining pressures in excess of 2 kbar prevents the iron-formation from retaining a 10 to 20% porosity (Hobbs *et al.*, 1976); the rock deforms in order to reduce the stress difference between grain–grain contacts and grain–pore contacts. The metamorphosed iron-formation now contains no visible microscopic porosity. Deformation isolates the iron-formation from the gabbro, and continued re-equilibration occurs under closed-system conditions. Inhomogeneities in equilibration with the fO_2 of the gabbro during the porous stage of the process are then preserved during the closed-system stage, resulting in contact-metamorphosed iron-formations with a range in fO_2 indicating "initial"-value status of oxygen, but with low average values of fO_2.

In summary, we agree with Frost that contact-metamorphosed iron-formations display a range in fO_2 conditions indicating metamorphism under closed-system conditions. We disagree, however, with the notion that the Stillwater and Duluth gabbros provided only the heat for metamorphism. The low average fO_2 of contact-metamorphosed iron-formations in comparison with regionally metamorphosed iron-formations is more than coincidental. We propose that contact metamorphism occurred in two stages. During the first stage when the rate of change in temperature was rapid and when a rapid decrease in solid volume resulting from nearly complete devolatilization occurred, the iron-formation underwent various degrees of reduction by interaction with the gabbroic intrusions. During the second stage, deformation effectively eliminated porosity, and closed-system metamorphism continued, preserving the inhomogeneities in Fe^{3+}/Fe^{2+} inherited from the first stage. Contrary to Frost's remarks about the intervening pelitic rocks that prevent exchange between gabbro and iron-formation, the Gunflint–Biwabik Iron-Formation is in direct contact with the Duluth Complex. The Stillwater Iron-Formation is separated from the Stillwater Complex by a few meters of pelitic rock, but the iron-formation and gabbro are intersected by high-angle, normal faults that were probably active during emplacement of the complex, although the most recent offset occurred during Laramide time (Page, 1977). Thus the pelitic rocks were not impediments to exchange between gabbro and iron-formation.

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